IMPROVEMENT IN THERMOSPHERIC NEUTRAL DENSITY ESTIMATIONS OF THE NUMERICAL TIE-GCM BY INCORPORATING HELIUM DATA FROM THE EMPIRICAL NRLMSISE-00 MODEL

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The total atmospheric neutral densities derived from the CHAMP (CHAllenging Minisatellite Payload) and the GRACE (Gravity Recovery And Climate Experiment) accelerometer data are used to investigate the accuracy of the empirical as well as numerical thermospheric neutral density models during the solar maximum (year 2002) through the solar minimum (year 2007). The models used in this thesis include the empirical models of the Jacchia-Bowman models (JB2006 and JB2008) and the MSIS (Mass Spectrometer Incoherent Scatter)-class model, and the numerical model of the NCAR’s TIE-GCM (National Center for Atmospheric Research’s Thermosphere Ionosphere Electrodynamics General Circulation Model). The thermospheric neutral density models show good agreement to the variations of neutral densities from the accelerometer data, but still had uncertainties which should be taken into account for better prediction of satellites’ position in orbit. The TIE-GCM shows larger uncertainties in the root mean square (RMS) in percent deviations at 400 km compared to the empirical models: 47.1% for the TIE-GCM compared to 15.7%, 10.3%, and 20.3% for the JB2006, JB2008, and NRLMSISE-00 models, respectively. The errors gradually increase with the decline in the solar activity. The partial pressures of helium obtained from the Naval Research Laboratory’s MSIS Extension 2000 (NRLMSISE-00) model are incorporated into the TIE-GCM to reflect the helium effect in calculating the molecular viscosity, the thermal conductivity, and the specific heat. As a result, the secular increases of the percent deviations are eliminated and the RMS of the TIE-GCM is improved to 21.4% and 22.8% for the densities from the CHAMP and the GRACE-A accelerometer data, respectively, with the incorporation of 71% partial pressures of helium from the NRLMSISE-00 model.
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NOMENCLATURE

- $a_p$ 3-hourly geomagnetic index
- $\vec{a}_d$ acceleration caused by drag (m/s$^2$)
- $\vec{a}_l$ acceleration caused by lift (m/s$^2$)
- $\vec{a}_{dl}$ acceleration caused by drag and lift (m/s$^2$)
- $c_{rd}$ coefficient of diffusive reflectivity
- $c_{rs}$ coefficient of specular reflectivity
- $g$ gravitational acceleration (m/s$^2$)
- $h$ height (km)
- $h_0$ arbitrary reference height
- $k$ Boltzmann constant ($= 1.38 \times 10^{-23}$ J/K)
- $m$ satellite mass for calculating neutral density (kg)
- $\overline{m}$ mean molecular weight (g/mol)
- $m_i$ molecular weight of individual species (g/mol)
- $n$ total number density (cm$^{-3}$)
- $n_i$ number density of individual species (cm$^{-3}$)
- $\hat{n}_i$ unit vector normal to $i^{th}$ satellite plate
- $u$ zonal neutral wind (eastward velocity, m/s)
- $v$ meridional neutral wind (northward velocity, m/s)
- $\vec{v}$ satellite velocity with respect to the surrounding atmosphere ($\vec{v} = \vec{v}_c - \vec{w}_c$)
- $\vec{v}_c$ satellite velocity relative to a co-rotating atmosphere
- $w$ vertical velocity ($w = dz/dt$)
- $\vec{w}_c$ wind velocity relative to a co-rotating atmosphere
- $\hat{\mathbf{y}}$ unit vector in the along-track direction of the CHAMP satellite
- $z$ vertical pressure level ($z = \ln (P_0/P)$)
- $\delta_\odot$ declination of the Sun
- $\lambda$ wavelength (nm)
- $\nu$ frequency of the photon
- $\theta_i$ angle between $\hat{n}_i$ and $\vec{v}_c$
- $\rho$ atmospheric neutral density (kg/m$^3$)
- $\psi$ mass mixing ratio
- $\mu$ molecular viscosity (gm cm$^{-1}$ sec$^{-1}$)
- $\mu_i$ molecular viscosity of individual species (gm cm$^{-1}$ sec$^{-1}$)
- $A_i$ area for the $i^{th}$ plate of satellite (m$^2$)
- $A_p$ daily averaged geomagnetic index
- $C_{Di}$ drag coefficient for the $i^{th}$ plate of satellite
- $C_{Li}$ lift coefficient for the $i^{th}$ plate of satellite
- $C_p$ specific heat per unit mass at constant pressure (J gm$^{-1}$ K$^{-1}$)
- $C_{p_i}$ specific heat of individual species at constant pressure
- $C_v$ specific heat per unit mass at constant volume
- $C_{v_i}$ specific heat of individual species at constant volume
- $F_{10.7}$ 10.7 cm solar flux (2800 MHz, $sfu = 10^{22}$ W m$^{-2}$ Hz$^{-1}$)
- $F_{10.7A}$ 81-day averaged 10.7 cm solar flux
- $H$ pressure scale height ($= kT/\overline{m}g$)
- $H_i$ pressure scale height of individual species ($= kT/m_ig$)
- $K_E$ eddy diffusion coefficient (s$^{-1}$)
- $K_T$ molecular thermal conductivity (erg cm$^{-1}$ sec$^{-1}$ K$^{-1}$)
$K_T$  molecular thermal conductivity of individual species
$L$  cooling rate
$M_{10.7}$  160 nm solar irradiance (sfu)
$M_{10.7A}$  81-day average value of $M_{10.7}$
$P$  pressure
$P_0$  reference pressure ($5 \times 10^{-4} \mu b$)
$Q$  heating rate
$R_E$  mean Earth’s radius (= 6356.766 km)
$S_{10.7}$  26-34 nm solar irradiance (sfu)
$S_{10.7A}$  81-day average value of $S_{10.7}$
$T$  neutral temperature of the atmosphere (K)
$T_c$  global minimum nighttime exospheric temperature
$T_l$  local exospheric temperature based on $a_p = 0$
$T_x$  exospheric temperature or thermopause temperature
$\vec{V}$  horizontal velocity vector
$Y_{10.7}$  0.1-0.8 nm X-ray and Lyman-α solar irradiance (sfu)
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Chapter 1

INTRODUCTION

Uncertainty in the atmospheric neutral density estimation is a crucial error source for prediction of the drag forces on a satellite below about 600 km, and results in uncertainty in satellites’ orbits in low earth orbit (LEO). Thermospheric neutral densities have been investigated for the past several decades, and empirical as well as theoretical thermospheric neutral density models have been developed and improved in order to predict or forecast variability in neutral densities with space weather events such as geomagnetic activity, solar flares, coronal mass ejections (CMEs), solar extreme ultraviolet (EUV) and ultraviolet (UV) on the overall thermal structure of the middle and upper atmosphere [Killeen, 1987]. Using satellite drag measurements and a numerical global mean upper atmosphere model, Qian et al. [2006] showed that average mass density at 400 km decreased secularly at a rate of 1.7% per decade from 1970 to 2000. Similarly, based on the satellite measurements and empirical Naval Research Laboratory’s Mass Spectrometer Incoherent Scatter Extension 2000 (NRLMSISE-00) model, Emmert et al. [2008] found that the secular decrease rate of average mass density at 400 km was about 3.55% per decade from 1967 to 2007. During the time of solar cycle 23/24 minimum (from 2007 to 2009), the unusually very low solar activity caused the thermosphere to be cooler and lower in mass density than expected, and it was proven that the primary cause of the extremely low mass density was due to the unusually very low solar EUV radiation [Solomon et al., 2010].

Both empirical and theoretical models tend to focus on longer time scales, though scientific interest in short time scale variability, even down to diurnal time scales, exists especially related to gravity wave and other wave or tidal influences. Even though theoretical models currently have larger uncertainties in thermospheric neutral density estimations than
empirical models, the improvement of theoretical models helps to investigate short term variability of thermospheric neutral density which is crucial for prediction of satellites’ orbital behavior in LEO. In this dissertation, the empirical Jacchia-Bowman JB2006/JB2008 models and the NRLMSISE-00 model were used with the theoretical National Center for Atmospheric Research (NCAR)’s Thermosphere-Ionosphere-Electrodynamics General Circulation Model (TIE-GCM) for comparison of thermospheric neutral density estimations. The TIE-GCM is suitable and efficient for thermosphere/ionosphere studies and thus is selected to be the primary model for this dissertation. However, the TIE-GCM does not include any physics due to helium in the atmosphere in it. Helium plays an important role in the upper thermosphere during the solar minimum time and the error in neutral density estimation of the TIE-GCM gradually increases with the decline in the solar activity. The goal of this dissertation is to eliminate the gradually increasing errors of the TIE-GCM with approaching the solar minimum time and suggest methods to improve the accuracy in neutral density estimation of the TIE-GCM by incorporating helium data calculated from the empirical NRLMSISE-00 model. Density data derived from the accelerometer data of the CHAMP (CHAllenging Mini-Satellite Payload) and GRACE (Gravity Recovery And Climate Experiment) satellites were used to compare the accuracy of these models.

Chapter 2 reviews the atmospheric structure of the Earth, primarily the thermosphere and the ionosphere, and discusses how they are characterized by thermal distributions, charged particles, chemical constituents, among other components. Energy sources from the space environment, such as solar irradiances and geomagnetic activity, which are closely related to the variation of thermospheric neutral density, are also discussed. Several empirical and theoretical thermospheric neutral density models are also introduced. Chapter 3 focuses on detailing the TIE-GCM, JB2006, JB2008, and NRLMSISE-00 models, including the physics details in the TIE-GCM, and data reduction for model formulation and methods for computing neutral
temperature and density in the other empirical models. Chapter 4 introduces thermospheric neutral density derived from the accelerometer data of the CHAMP and GRACE satellites, which also show the effect of helium in the upper thermosphere over the period from the solar maximum to the solar minimum time. It also presents the improvement of the accuracy in neutral density estimation of the TIE-GCM by incorporating helium data calculated from the empirical NRLMSISE-00 model into the TIE-GCM. Since the NRLMSISE-00 model also has its own uncertainty in neutral density estimation, several cases were considered in order to show the sensitivity of the TIE-GCM compared to neutral density derived from the CHAMP and GRACE accelerometer data. Even though helium is a minor constituent in the atmosphere, it should be considered as one of the major neutral constituents in the upper thermosphere, especially during the solar minimum period. This is because the number density of helium becomes greater than that of molecular nitrogen at around 450 km during the solar maximum and this threshold altitude lowers to around 350 km during the solar minimum. The estimation errors in neutral density of the TIE-GCM, specifically root mean square (RMS), can be significantly reduced by incorporating the effects of helium from the NRLMSISE-00 model. Chapter 5 presents conclusion and discusses future work.
Chapter 2

OVERVIEW

The atmospheric structures of the Earth can be divided into several distinct regions categorized by specific characteristics of vertical thermal distribution, neutral composition, and charged particles. These atmospheric structures are dramatically affected by external forces from the space environment, with the primary energy being from the Sun. The solar irradiances vary periodically as well as sporadically and, as a result, affect the understanding of these space weather events and their effects on the Earth’s atmosphere. This understanding is also vital in predicting or forecasting atmospheric weather in the middle and upper atmosphere. The primary uncertainty in predicting the orbital behavior in LEO results in the uncertainty of atmospheric drag estimates exerted on satellites from the thermospheric neutral density. The thermospheric neutral density variations are closely correlated with the variations of solar irradiances. During the past several decades, thermospheric neutral density models, empirical as well as theoretical models, have been developed and improved to reduce the uncertainty of neutral density estimation and to better understand the thermospheric physics controlling the space weather.

Earth’s Atmospheric Structure

The Earth’s atmosphere is an extremely complicated system. In order to understand and model the atmosphere of the Earth, it is useful to divide it into sub-regions, appropriate to the field of study, according to the characteristics of the air. First, from the point of view of the vertical temperature gradient of the neutral atmosphere, the atmosphere can be divided into four regions, ranging from the troposphere to the thermosphere. Second, while the neutral constituents
are well mixed in the lower atmosphere due to the turbulence, the composition of the air is almost uniform from the Earth’s surface up to a certain height. The effect of the turbulence is independent of the vertical distribution of gases and, as a result, the composition of the air is no longer uniform but varies with height in the upper atmosphere. Therefore, the neutral density can be derived by different methods between the two layers. Third, neutral constituents of the atmosphere are ionized by solar radiation, and, in the upper atmosphere, ions and electrons created by the ionization process can exist for short periods of time due to the thin air. Since the composition of ions varies with height, the atmosphere can be classified by the characteristics of these charged particles.

**Atmospheric structure with thermal characteristics**

The atmosphere of the Earth can be divided into several layers according to the vertical distribution of the neutral temperature, as shown in Figure 2-1, which is the most common classification. The height profiles of the temperature and mass density in Figure 2-1 are calculated from the NRLMSISE-00 model near the solar maximum on day 180 in 2002. The lowest layer is the troposphere containing about 99% of the water vapor in the atmosphere. The daily weather occurs in this layer. It extends from the Earth’s surface to the tropopause varying from about 8 km height at the geographical poles to about 18 km height at the equator. Starting from the mean surface temperature of about 288 K, the temperature in the troposphere decreases with the temperature lapse rate of -6 K/km, as most of the solar visible radiation is absorbed in the Earth’s surface. The negative vertical temperature gradient in this layer leads to the temperatures between about 190 K at the equator and about 220 K at high latitudes at the tropopause [Banks and Kockarts, 1973]. Continuing upward, 90% of the total amount of atmospheric ozone is present in the tropopause height to about 50 km, known as the stratosphere.
In this atmospheric region, the UV solar radiation above 242 nm wavelengths are absorbed by the trace ozone gas [Prölls, 2004] so that the temperature rises again up to about 270 K at the upper boundary (the stratopause). Above the stratopause, the mesosphere extends up to about 90 km above the Earth. In this layer, concentrations of ozone and water vapor are negligible so that very little of the solar UV radiation are absorbed in the atmosphere and the temperature is lower than that of the lower atmospheric regions. Instead of the absorption of solar radiation, the infrared (IR) cooling, particularly due to the trace gas carbon dioxide, plays an important role. As a result, the temperature in the mesosphere decreases again with altitude up to 180 K~190 K at its upper

Figure 2-1 Height profiles of total mass density and atmospheric structures of the Earth according to the characteristics of neutral temperature, composition, and thermal plasma (ionosphere). Neutral temperature and total mass density are calculated from the NRLMSISE-00 model near solar maximum (day 180 in 2002).
boundary of about 90 km (beginning of the mesopause). The temperature in the mesopause is the lowest in the Earth’s atmosphere, and it can be as low as 120 K under extreme conditions. The atmospheric region from the stratosphere to the mesopause is called the middle atmosphere. The dominant molecules from the Earth’s surface to the mesosphere are molecular nitrogen (N$_2$) and molecular oxygen (O$_2$). Number densities for these constituents can be seen in Figure 2-2.

Figure 2-2  Height profiles of number density of neutral constituents calculated from the NRLMSISE-00 model near solar maximum (day 180 in 2002). The exospheric temperature, T$_\infty$, is around 1,120 K.
The region above the mesopause with the minimum temperature in the Earth’s atmosphere is called the thermosphere and is regarded as the upper atmosphere. Here, chemical reactions occur much faster than on the Earth’s surface. In the lower thermosphere, up to about 250 km, the temperature dramatically increases with height due to the absorption of solar UV radiations below 242 nm wavelengths and the lack of effective heat loss processes. At heights above about 250 km, the increase in temperature eventually slows down due to the decrease of absorption of solar UV radiation. The temperature approaches the exospheric or thermopause temperature, typically 1,000 K, but occasionally varies from 600 K to 2,500 K under extreme solar conditions. As seen in Figure 2-2, the dominant molecule below around 200 km is N₂. Since the neutral constituents are diffusively separated with their own molecular weights above around 105 km, the light species, such as atomic oxygen (O), helium (He), and atomic hydrogen (H), prevail over the heavier neutral constituents like N₂ and O₂. Therefore, the number density of O becomes greater than that of N₂ above around 200 km. Furthermore, the He prevails over the N₂ above around 500 km, and is considered as a major constituent. It overtakes the O above around 700 km. However, these heights are very sensitive to the solar activity. The highly reactive gas, O, is not a natural species in the lower atmosphere. It is created by the photo-dissociation of O₂ (Eq. (2.1)) by the solar radiations below a wavelength of 242 nm, primarily by the Schumann-Runge bands and continuum. It is created as by-products of the several ion-neutral and neutral-neutral chemical reactions in the thermosphere [Fuller-Rowell and Rees, 1996]. The other highly reactive gas, H, is very similar to O. It is also created by the dissociation of water molecules (Eq. (2.2)) and has long life time in the middle and upper thermosphere since the time of recombination process as a loss process is extremely slow.

\[ O_2 + \text{photon} \ (\lambda \leq 242 \ \text{nm}) \rightarrow O + O \]  

(2.1)
\[
\text{H}_2\text{O} + \text{photon} (\lambda \leq 242 \text{ nm}) \rightarrow \text{H} + \text{OH}
\] (2.2)

**Atmospheric structure with characteristics of composition**

In addition to the classification of the Earth’s atmosphere according to the vertical distribution of the neutral temperature, it can be divided into two main regions by the characteristics of composition and is very useful for modeling the atmosphere: homosphere and heterosphere. The Earth’s atmosphere at the ground is mainly composed of N\textsubscript{2}, O\textsubscript{2}, Ar, and CO\textsubscript{2} contributing 78\%, 21\%, 0.9\%, and 0.03\%, respectively by volume. The remaining atmosphere is

![Figure 2-3](image)

**Figure 2-3** Height profiles of the molecular diffusion coefficient \((D)\), the time constant for molecular diffusion \((\tau_D)\), the eddy diffusion coefficient \((K_E)\), and the time constant for eddy diffusion \((\tau_E)\). The homopause height is defined as the height where \(K_E(h_{HP}) = D(h_{HP}) \) or \(\tau_D(h_{HP}) = \tau_E(h_{HP})\) [Prölls, 2004]
composed of other minor constituents. The mixture holds all the way up to around 105 km since the rate of mixing of atmospheric constituents by winds and dissipative turbulence is sufficiently rapid to mix the gases in order for the relative composition to be almost uniform in the region. The homopause height is determined as the height where the molecular diffusion coefficient equals the eddy diffusion coefficient, somewhere around 105 km above the Earth, depending on the solar activity; in other words, the height that the time constant for molecular diffusion becomes the same as that for eddy diffusion (Figure 2-3). The atmospheric region below the homopause, where the mixture of the air is uniform and, therefore, the atmosphere has constant mean molecular weight, is called the homosphere or turbosphere. Above the homopause, the atmosphere is no longer well mixed since the turbulence does not greatly affect the vertical distribution of gases. This region is called the heterosphere. In the heterosphere, the composition of the atmosphere varies with altitude since the molecules are more independent upon each other and are gravitationally separated in the way that heavy gases, such as O₂, N₂, and O, fall off rapidly and light gases, like He and H, fall off gradually with altitude (Figure 2-2).

The Earth’s atmosphere can be also divided by gravitational binding in regions, named the barosphere and exosphere. The boundary between the barosphere and exosphere is called the exobase, and ranges from around 500 km to 1,000 km above the Earth. Below the exobase, particles moving outward lose their energy by colliding with other particles frequently so as not to escape from the Earth’s gravitational influence. In the upper thermosphere, the mean free path, defined as the average distance between collisions for a gas molecule, becomes so large that particles rarely collide with other particles and, as a result, they can escape from the gravitational bounding of the Earth because they have sufficient energy. The exobase is defined as the height where the mean free path equals the pressure scale height [Prölls, 2004]. In the barosphere, the hydrostatic equation, Eq. (2.3), which explains the relation between the pressure gradient and the
gravitational forces, can be derived from the assumption of force equilibrium, i.e. the gravitational force is equal to and opposite from the pressure force.

\[
\frac{dp(z)}{dz} = -p(z) g(z)
\]  

(2.3)

Before solving the hydrostatic equation, it is necessary to introduce one of the most important parameters in the atmospheric characteristics; the pressure scale height, \(H\), defined in Eq. (2.4). The pressure scale height represents the height interval over which the atmospheric pressure decreases by a factor of \(1/e\). The scale height close to the Earth’s surface is around 8.43 km in contrast to around 50 km at 300 km above the Earth [Brekke, 1997], which means the atmospheric density decreases more rapidly in the lower atmosphere than in the upper atmosphere.

\[
H(h) = \frac{k T(h)}{m(h) g(h)}
\]  

(2.4)

The solution of the hydrostatic equation with the ideal gas law is called the well-known barometric law; Eq. (2.5) for atmospheric pressure and Eq. (2.6) for number density of neutral constituents. These equations can be directly applied to the homosphere for obtaining atmospheric pressure and number density, provided the temperature profile is determined. In the heterosphere, however, since the neutral constituents are gravitationally separated, the force equilibrium relation between the pressure gradient and gravitational force should be applied to individual species. As a result, the number density profile of individual neutral constituent in the heterosphere, i.e. the barometric law in the heterosphere, can be obtained by Eq. (2.7) with the assumption of thermodynamic equilibrium \((T = T_i)\) within a gas mixture.

\[
p(h) = p(h_0) \exp \left\{ - \int_{h_0}^{h} \frac{dz}{H(z)} \right\}
\]  

(2.5)
Using the barometric law, the number density at a certain height \( h \) can be calculated from the number density at a reference height \( h_0 \) using the temperature profile. Therefore, the temperature profile should be determined in order to calculate the neutral density profile in the barosphere. The Bates-Walker temperature profile, expressed by Eq. (2.8), can represent the neutral temperature in reasonably good agreement with the observations after many efforts to improve the approximation since it was first introduced [Bates, 1959; Jacchia, 1964; Walker, 1965]. Typically, \( h_0 = 120 \) km, \( T(h_0) = 350 \) K, and \( s = 0.021 \) km\(^{-1}\) show good agreement with the observations [Prölss, 2004].

\[
T(h) = T_\infty - \left(T_\infty - T(h_0)\right)e^{-s(h-h_0)}
\]

(2.8)

**Atmospheric structure with characteristics of charged particles**

The neutral atoms and molecules are subject to ionization caused by the solar radiation. Since the atmospheric constituents are dramatically rarefied in the upper atmosphere, free electrons produced from the ionization can exist for short period of time before they are captured by the positive ions (ionosphere) unlike in the lower atmosphere where free electrons cannot exist. Therefore, the atmosphere can be divided into several sub-regions by the characteristics of the behavior of charged particles. The ionosphere is usually divided into three sub-regions by the peaks of the height profile of the total electron density: D-, E-, and F-Regions.
The F-Region can be further subdivided into F1-Region and F2-Region, as shown in Figure 2-1. The ionospheric region above the F-Region peak is called the topside ionosphere. The classification of the ionospheric regions on daytime mid-latitude ionosphere is shown in Table 2-1, but the altitude range varies significantly with the solar activity. Since the primary cause in the low- and mid-latitude ionosphere is the photo-ionization of the neutral constituents, the ionic composition generally follows the composition of neutral atmosphere with some exceptions, as shown in Figure 2-4. And, since the main energy source of the photo-ionization of neutral constituents is the solar EUV radiation, the height profile of electron density in the ionosphere strongly varies with solar activity and the Earth’s rotation. Figure 2-5 shows the variations of the electron density profile during the solar maximum/minimum and day/night time. Typically, the maximum electron density is around 300 km, varying between around 200 km and 600 km with solar activity [Brekke, 1997].

<table>
<thead>
<tr>
<th>Regions</th>
<th>Height</th>
<th>Major Component</th>
<th>Production Cause</th>
</tr>
</thead>
<tbody>
<tr>
<td>D</td>
<td>60-90 km</td>
<td>NO⁺, O₂⁺</td>
<td>Lyman-α, hard X-rays</td>
</tr>
<tr>
<td>E</td>
<td>90-140 km</td>
<td>O₃⁺, NO⁺</td>
<td>Lyman- β, soft X-rays, UV Continuum</td>
</tr>
<tr>
<td>F1</td>
<td>140-200 km</td>
<td>O⁺, NO⁺</td>
<td>He II, UV Continuum (10-80 nm)</td>
</tr>
<tr>
<td>F2</td>
<td>200-400 km</td>
<td>O⁺, N⁺</td>
<td>He II, UV Continuum (10-80 nm)</td>
</tr>
<tr>
<td>Topside Ionosphere</td>
<td>&gt; 400 km</td>
<td>O⁺</td>
<td>Transport from below</td>
</tr>
<tr>
<td>Plasmasphere</td>
<td>&gt; 1200 km</td>
<td>H⁺</td>
<td>Transport from below</td>
</tr>
</tbody>
</table>
The D-Region is the lowermost layer of the ionosphere between about 60 km and 90 km, where NO$^+$ and O$_2^+$ ions are the major components. NO$^+$ ions are directly created by the photo-ionization process by the Lyman-$\alpha$ line and are created by the charge exchange from N$_2^+$ and O$_2^+$ ions mainly ionized by the hard X-rays below 1 nm wavelengths. As a result, the minor constituent, NO, in the neutral atmosphere surprisingly becomes the major component, NO$^+$, in the D-Region ionosphere. On the contrary, the major neutral constituent, N$_2$, becomes the minor component, N$_2^+$, in the lower ionosphere. Above the D-Region, O$_2^+$ and NO$^+$ ions are still major ion concentrations up to about 140 km, called E-Region. N$_2^+$ and O$_2^+$ ions are created by soft X-rays between 1 nm and 10 nm, and O$_2^+$ ions are also created by the Lyman-$\beta$ line. O$^+$ ions becomes the dominant ionic constituent above the E-Region up to the boundary of the

Figure 2-4  The composition of the ionosphere above 90 km observed during the solar minimum day at mid-latitudes [Johnson, 1966]
ionosphere, called F-Region and Topside Ionosphere. Above the topside ionosphere, the $\text{H}^+$ ions are the abundant species in the region called the plasmasphere.

The production and loss of the ions and electrons, which determines the height profile of the ionosphere, are caused by several physical processes. The primary ionization process is the photo-ionization of thermospheric neutral constituents by solar EUV and X-ray radiations. The photo-ionization processes of the predominant neutral constituents O, $\text{N}_2$, and $\text{O}_2$ in the upper atmosphere are [Prölss, 2004]

\begin{align*}
\text{O} + \text{photon } (\lambda \leq 91 \text{ nm}) & \rightarrow \text{O}^+ + e \quad (2.9) \\
\text{N}_2 + \text{photon } (\lambda \leq 80 \text{ nm}) & \rightarrow \text{N}_2^+ + e \quad (2.10) \\
\text{O}_2 + \text{photon } (\lambda \leq 103 \text{ nm}) & \rightarrow \text{O}_2^+ + e \quad (2.11)
\end{align*}

where, $e$ is the photoelectron. The excess energy is created after the photo-ionization processes, appears in the form of kinetic energy of the photoelectron released in the photo-ionization processes, and is one of the important heat sources in the thermosphere and ionosphere. In addition to the primary ionization process of the photo-ionization of the neutral constituents, there are secondary but important ionization processes to determine the structure of the ionosphere: the charge exchange and particle precipitation. The photoelectron released from the photo-ionization process can attain sufficiently high energy to ionize neutral atmosphere with the charge exchange reactions. Equations (2.12) through (2.18) are the important charge exchange reactions in the ionosphere [Schunk, 1983]. In the lower ionosphere, the important sources of $\text{O}_2^+$ ions are reactions (2.13) and (2.15). The charge exchange reaction plays more important role in the upper ionosphere. Above the Topside Ionosphere, i.e. the plasmasphere, the reaction (2.14) is the main source for the $\text{H}^+$ ions. Unlike to the simple charge exchange reactions, $\text{NO}^+$ ions are created by somewhat more complicated charge exchange reactions with dissociation. The reactions (2.12),
(2.17), and (2.18) are the main sources of NO$^+$ ions, and explain the way that NO$^+$ ion is the dominant species in the lower ionosphere even though it is a minor neutral constituent in the lower thermosphere. In addition, N$_2^+$ ions created from the photoionization process (2.10) are rapidly converted into NO$^+$ ions with the reaction (2.17). The other secondary ionization process in the upper atmosphere is the particle precipitation at particularly higher latitudes. Energetic particles, mostly electrons but also protons and other ions, flow into the upper atmosphere along the magnetic field lines from the magnetosphere. These energetic particles can reach down to the mostly lower thermosphere around 100 km, but some can reach the troposphere or even down to the Earth’s surface depending on the particles’ energy. The high energetic charged particles precipitating into the upper atmosphere collide with other particles. Through the collisions, the charged particles ionize and dissociate the atoms and molecules in the upper atmosphere.

While the ions are produced by the ionization processes, i.e. photo-ionization, charge exchange, and particle precipitation, they are also neutralized by the ionization loss processes: the

\[
\begin{align*}
O^+ + N_2 &\rightarrow NO^+ + N & (2.12) \\
O^+ + O_2 &\rightarrow O_2^+ + O & (2.13) \\
O^+ + H &\leftrightarrow H^+ + O & (2.14) \\
N_2^+ + O_2 &\rightarrow O_2^+ + N_2 & (2.15) \\
N_2^+ + O &\rightarrow O^+ + N_2 & (2.16) \\
N_2^+ + O &\rightarrow NO^+ + N & (2.17) \\
N^+ + O_2 &\rightarrow NO^+ + O & (2.18)
\end{align*}
\]
dissociative recombination, radiative recombination, and charge exchange. Ions are produced as well as neutralized by the charge exchange process such as reactions (2.12) through (2.18), while the ionization density as a whole does not change by this process. The dissociative and radiative recombinations are the most important loss processes for molecular and atomic ions, respectively. The dissociative recombination of molecular ion is the ionization loss process which a positive ion recombines with a negative electron creating new neutral constituents from the recombination. The important examples of the dissociative recombination processes in the ionosphere are the reactions (2.19) and (2.20). O₂ ions are dissociated into atomic oxygen, and NO ions are separated into atomic nitrogen and atomic oxygen by the dissociative recombination. The dissociative recombination processes of NO⁺ and O₂⁺ play an important role in the lower ionosphere. The radiative recombination of atomic ion, relatively slow process in the ionosphere, is the ionization loss process which a positive ion recombines with an electron creating a neutral constituent and entirely or partly radiating the excess energy. The reaction (2.21) is an example of the radiative recombination.

\[ \text{O}_2^+ + e \rightarrow O + O \]  
(2.19)

\[ \text{NO}^+ + e \rightarrow N + O \]  
(2.20)

\[ \text{O}^+ + e \rightarrow O + \text{photon} \]  
(2.21)
Energy sources from the space environment

Abundant energy from the space environment affects the Earth’s atmospheric structure such as thermal distribution, neutral and charged-particle composition, magnetic field, among others. Photons, neutral and charged particles, and interplanetary magnetic field (IMF) coming from the Sun, which is the primary energy source from the space environment to the Earth, interact complexly with the Earth’s atmosphere and, as a result, the variations of the Earth’s atmospheric structure primarily depend on the solar activity such as the 11-year sunspot cycle, 27-day solar rotation, solar flares, coronal mass ejections (CMEs), among other phenomena.

Figure 2-5  Height profile of total electron density on day and night time with ionospheric regions during the solar maximum and the solar minimum time [Jursa, 1985]
Solar variations and irradiance

The integrated emission from the whole solar disc can be divided into three components with the criterion of the timescales of the intensity variations: (1) R-component (rapid variations such as flares with the durations less than an hour), (2) S-component (slow variations over hours to years), and (3) a quiet component extrapolated to zero plage area, termed as “Quiet Sun Level” [Tapping and DeTracey, 1989]. Since the first discovery of S-component by Covington [1947], it has been turned out that the S-component at 10.7 cm wavelength is strongly correlated with the sunspots and the 10.7 cm (2,800 MHz) solar flux, $F_{10.7}$, can be used as the proxy of the integrated fluxes in UV and EUV [Chapman and Neupert, 1974; Bossy, 1983; Donnelly et al., 1983; Oster, 1983; Hedin, 1984; Nicolet and Bossy, 1985; Lean, 1987]. As a result, $F_{10.7}$ index has been widely used in the empirical as well as the theoretical atmospheric models as an input for the level of solar activity. It is produced daily by the ground-based Dominion Radio Astrophysical Observatory (DRAO) located in Penticton, British Columbia, Canada. Observations of the $F_{10.7}$ values are made at 17, 20, and 23 UT each day and made available through the Space Weather Canada website*. The 20 UT (local noon) observed values are archived at the National Geophysical Data Center (NGDC)/World Data Center†.

One of the most significant manifestations of the solar magnetic activity is the 11-year solar cycle which can be characterized by waxes and wanes of sunspots with approximately 11-year periodicity. It was first introduced by Wolf [1852] after examining sunspots from the Zürich Observatory sunspot records. Sunspot, dark area in photosphere of the Sun lasting from several hours up to several months, is a vortex of gas associated with stormy localized magnetic field and is a relatively cooler area compared to the surroundings resulted from less emission of electromagnetic radiations from below. Since intense magnetic activity exists in sunspot areas,

* http://www.spaceweather.ca/sx-eng.php [retrieved 22 August 2010]
the variation of sunspots is strongly related to the amount of solar radiation emitting to the space from the Sun and, as a result, significantly affects the variation of the Earth’s atmosphere. International sunspot number [Clette et al., 2007] represents the variation of sunspots and has been archived in Solar Influences Data Analysis Center (SIDC) in the Royal Observatory of Belgium. Figure 2-6 shows the yearly averaged sunspot numbers from the year 1610 to 2000. It evidently shows the approximately 11-year periodicity of the recurrences of sunspots and two remarkable events of low solar activity; named as Maunder Minimum in about 1645~1715 and Dalton Minimum in about 1790~1830 [Eddy, 1976]. During especially the Maunder Minimum, the Sun was extremely inactive with few sunspots observed and this period closely coincides with one of the coldest periods, so-called “Little Ice Age”. Figure 2-7 (a) and (b) illustrate the

![Yearly averaged sunspot numbers 1610-2000](http://solarscience.msfc.nasa.gov/images/ssn_yearly.jpg) [retrieved 22 August 2010]

Figure 2-6 Yearly averaged sunspot numbers from the year 1610 to 2000 (courtesy of NASA/Marshall Space Flight Center)[§]

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Figure 2-7  (a) Monthly averaged sunspot number, (b) daily and 81-day averaged 10.7 cm solar flux**, and (c) daily averaged geomagnetic storm index ($A_p$)††

monthly averaged sunspot numbers and $F_{10.7}/F_{10.7A}$ indices during the past several solar cycles. The figures show the strong correlations between the sunspot numbers and $F_{10.7}$ indices, and the increases and decreases of $F_{10.7}$ indices show the approximately 11-year periodicity like sunspot numbers do.

Other than the 11-year periodicity of the recurrences of sunspots, the most prominent variation of the Sun is the approximately 27-day solar rotation attributing to the rotation of sunspot groups. Since the Sun is a gaseous planet, it rotates differentially with the location: about 26 days at the equator and about 35 days at the poles. Furthermore, it has been shown that sunspots and magnetic features rotate faster than the local plasma on the solar surface [Howard and Harvey, 1970; Howard, 1984; D’Silva and Howard, 1994]. As a result, the rotating system is totally complicated and it is very difficult to predict the variations of the solar irradiances. Figure 2-8 shows the approximately 27-day solar rotational effects by the daily $F_{10.7}$ values from the year of 1992 through 2000, which covers the period from the declining phase of solar cycle 22 to the ascending phase of solar cycle 23 [Kane, 2003]. The deviations in percentage from the 27-day running means are plotted to show the variations of the daily $F_{10.7}$ values. Dots and crosses represent the maxima and minima of deviations, respectively. The numbers in the figure show the intervals between two consecutive maxima, the parentheses are used for the intervals exceeding 36 days, and the boxes are used for the intervals under 14 days. Intervals exceeding 35 days are broken up into subintervals so as to be about equal parts of rotational periodicity. Even though the overall average interval of the maxima is about 27 days, the intervals vary in wide range about 13~40 days. One of the possible reasons to the wide range of fluctuations in periodicity is that the active regions like sunspots move rapidly forward or backward compared to local plasma.
Figure 2-8 Deviations in percent of daily $F_{10.7}$ values from the 27-day running means from the year of 1992 to 2000. Dots and crosses represent the maxima and minima, respectively, and the numbers indicate the intervals of consecutive maxima. Parentheses indicate over 36-day intervals and boxes represent under 14-day intervals [Kane, 2003].
Sometimes there are powerful explosions in the atmosphere above the sunspots, which are called solar flares. A sudden release of magnetic energy built up in the solar atmosphere is the cause of solar flares. A solar flare, a bright flash of light, lasts only a few minutes to a few hours, but the explosion radiates across virtually the entire electromagnetic spectrum from radio waves to X-/γ-rays and sends bursts of energetic particles including electrons and protons into space. Since a solar flare is the source of sporadic particle and electromagnetic emissions that affect the Earth’s upper atmosphere, it is important in solar-terrestrial relations. Solar flares are categorized into class B, C, M, and X according to their intensity (magnitude of peak burst) as listed in Table 2-2 [Bray and Loughhead, 1965].

<table>
<thead>
<tr>
<th>Class</th>
<th>Magnitude of peak burst (W/m²)</th>
</tr>
</thead>
<tbody>
<tr>
<td>B</td>
<td>~ 10^6</td>
</tr>
<tr>
<td>C</td>
<td>10^6 ~ 10^5</td>
</tr>
<tr>
<td>M</td>
<td>10^4 ~ 10^4</td>
</tr>
<tr>
<td>X</td>
<td>10^4 ~</td>
</tr>
</tbody>
</table>

Another form of mass release is a coronal mass ejection (CME). CMEs are due to the break-outs of magnetically closed regions in the corona on the Sun, which suddenly release as much as 10^{13} kg of plasma into the space [Hundhausen, 1988; Kahler, 1988]. They expand as they move away from the Sun, called the solar wind, and the speed of solar material ranges from less than 50 km/s to greater than 1,200 km/s at the height of several solar radii from the surface [Howard at el., 1985]. CMEs are often classified into two categories according to the initial speed: slow CMEs for less than 400 km/s and fast CMEs for higher than 400 km/s [Borgazzi at el., 2009]. Several scientists suggested that CMEs, particularly fast CMEs, were the causes of
non-recurrent geomagnetic storms [Tsurutani et al., 1988; Gosling et al., 1990 and 1991] and it was confirmed by Landi et al. [1998]. According to the study by Landi et al., the majority of the non-recurrent geomagnetic storms (about 70%) take place after the enhancements of solar soft X-ray emissions associated with CMEs within the interval of 5 days. Therefore, CMEs can be considered as the leading causes of non-recurrent geomagnetic storm events.

Since the late 1970s, satellites have been recording the total solar irradiances (TSI) with a series of different radiometers and satellites: HF on NIMBUS 7, ACRIM (Active Cavity Radiometer Irradiance Monitor) I on the SMM (Solar Maximum Mission), ACRIM II on the UARS (Upper Atmosphere Research Satellite), and VIRGO (Variability of solar Irradiance and Gravity Oscillations) on the SOHO (Solar and Heliospheric Observatory). However, there are differences in TSI values from the different satellites. These differences represent that the TSI estimates from different satellites are not exactly the same and that these are in the range of approximately 1,365–1,373 W-m$^{-2}$. Fröhlich [2006] reconstructed these time series by shifting each of them to a certain level and merged them together. Figure 2-9 represents those corrected TSI values during the three solar cycles: solar cycles 21, 22, and 23 [Fröhlich, 2006]. Different colors represent the TSI values from the different radiometers in different satellites, and the thick black line represents the 81-day average values. The average TSI value is about 1,366 W-m$^{-2}$ and the average minimum TSI value during solar minima is about 1,365.5 W-m$^{-2}$. Furthermore, the variation of the TSI values from solar maxima to solar minima is within about 0.2%.
The huge amounts of energy approximately 1,366 W-m\(^{-2}\) from the Sun ranging almost the entire electromagnetic spectrum are crucial to the atmosphere of the Earth. While some solar indices are almost absorbed in the thermosphere, others can reach to the Earth’s surface. While some solar indices largely affect temperature and density of the Earth’s atmosphere, others can be negligible. The International Standard Organization (ISO) has developed the standards for solar irradiances to categorize solar irradiances according to the characteristics. Table 2-3 shows the spectral subcategories of solar irradiance products from the hard X-rays to the radio frequencies, the source regions of the Sun, and the thermal absorption regions by the Earth’s atmosphere [ISO, 2007; Tobiska and Nusinove, 2000].

Figure 2-9 Total solar irradiances from solar cycles 21, 22, and 23. Colors represent datasets from different radiometers in space; HF radiometer, ACRIM and VIRGO. Thick black line shows the 81-day average values [Fröhlich, 2006].
Table 2-3  Classification of solar irradiance spectrum and atmospheric heating

<table>
<thead>
<tr>
<th>ISO 21348 Spectral Subcategory</th>
<th>Wavelength Range</th>
<th>Solar source Region</th>
<th>Atmospheric thermal absorption region</th>
</tr>
</thead>
<tbody>
<tr>
<td>hard X-rays</td>
<td>~ 1 nm</td>
<td>Corona</td>
<td>Thermosphere</td>
</tr>
<tr>
<td>soft X-rays ultraviolet (XUV)</td>
<td>1 nm ~ 30 nm</td>
<td>Corona</td>
<td>Thermosphere</td>
</tr>
<tr>
<td>extreme ultraviolet (EUV)</td>
<td>30 nm ~ 120 nm</td>
<td>Chromosphere</td>
<td>Thermosphere</td>
</tr>
<tr>
<td>far ultraviolet (FUV)</td>
<td>120 nm ~ 200 nm</td>
<td>Chromosphere</td>
<td>Thermosphere</td>
</tr>
<tr>
<td>middle ultraviolet (MUV)</td>
<td>200 nm ~ 300 nm</td>
<td>Chromosphere</td>
<td>Mesosphere</td>
</tr>
<tr>
<td>ultraviolet (UV)</td>
<td>300 nm ~ 400 nm</td>
<td>Chromosphere</td>
<td>Stratosphere</td>
</tr>
<tr>
<td>visible (VIS)</td>
<td>400 nm ~ 760 nm</td>
<td>Photosphere</td>
<td>Troposphere</td>
</tr>
<tr>
<td>infrared (IR)</td>
<td>0.76 µm ~ 10 µm</td>
<td>Photosphere</td>
<td>Troposphere</td>
</tr>
<tr>
<td>far infrared</td>
<td>10 µm ~ 1 mm</td>
<td>Photosphere</td>
<td>Troposphere</td>
</tr>
<tr>
<td>radio</td>
<td>1 mm ~</td>
<td>Cool corona</td>
<td>Thermosphere</td>
</tr>
</tbody>
</table>

The $F_{10.7}$ index has been used as a proxy for EUV emissions for thermospheric heating in various thermospheric density models because thermospheric heating is dominated by the solar EUV emissions. However, the thermosphere is also heated by the solar energy other than the EUV emissions, such as the coronal soft X-rays, Lyman-$\alpha$, FUV wavelengths, among others [Banks and Kockarts, 1973]. Tobiska et al. developed several solar indices for representing these solar emissions to the JB2006 model ($S_{10.7}$ and $M_{10.7}$ were used in the model). These new solar indices were published in 2008 [Tobiska et al., 2008]. Another solar index, $Y_{10.7}$, was developed and incorporated into the latest Jacchia-class model (JB2008 model) [Bowman et al., 2008b].
• **$S_{10.7}$ index**

Since the launch of the NASA/ESA SOHO research satellite in December 1995, the integrated 26~34 nm solar irradiance, $S_{10.7}$, has been measured by one of the instruments on the satellite; Solar EUV Monitor (SEM). The orbit for the SOHO research satellite was chosen to a halo orbit at the Lagrange Point 1 ($L_1$) on the Earth-Sun line, approximately 1.5 million km from the Earth, in order not to be interrupted by the Earth or the Moon for observing the Sun. The photons from 26~34 nm wavelengths are mostly absorbed by atomic oxygen at the thermosphere above around 200 km.

• **$M_{10.7}$ index**

The $M_{10.7}$ index represents photospheric and lower chromospheric solar FUV Schumann-Runge Continuum emission near 160 nm. The photons in this wavelength are mostly deposited into the lower thermosphere by photo-dissociation process of molecular oxygen. The $M_{10.7}$ index is not directly produced by measuring the 160 nm solar FUV emissions but indirectly calculated from the chromospheric Mg II core-to-wing ratio (cwr) of 279.56 and 280.27 nm solar MUV emissions since the Mg II cwr is a good proxy for the 160 nm solar FUV emission. Mg II cwr is produced by the Solar Backscatter UV (SBUV) spectrometer equipped in the NOAA series operational satellites, and provided daily by NOAA’s Space Environment Center (SEC).

• **$Y_{10.7}$ index**

The $Y_{10.7}$ index is incorporated into the JB2008 model in order to represent the heating in the mesosphere and the lower thermosphere, approximately 85~100 km, by the solar X-rays (0.1~0.8 nm) and Lyman-$\alpha$ (121 nm). The photons of solar X-rays arriving the Earth are deposited in this atmospheric layer by ionizing $O_2$ and $N_2$ and, as a result, they create the ionospheric D-region. Solar X-ray emissions are measured by the X-ray
Spectrometer (XRS) on the GOES (Geostationary Operational Environmental Satellite) series operational satellites. On the other hand, the photons of Lyman-α emission are absorbed in the same atmospheric layer by dissociating NO and participating in H₂O chemistry. The solar X-ray and Lyman-α emissions are competing drivers in the mesosphere and lower thermosphere. During high solar activity, the solar X-ray emissions are major energy source in this atmospheric region. However, the Lyman-α emission takes over the dominance during moderate and low solar activity. With this competing relation between the solar X-ray and Lyman-α emission, the composite solar index, Y₁₀.7, was developed to be weighted for representing mostly solar X-ray emissions during high solar activity and representing mostly Lyman-α emissions during moderate and low solar activity.

**Geomagnetic activity**

The geomagnetic activity results from the interaction between the solar wind and the Earth’s magnetic field. The steadily streaming outflow of fully ionized plasma (solar wind) carries with it the solar magnetic field, which is known as the interplanetary magnetic field (IMF). The solar wind can vary markedly on an hourly basis and is highly structured throughout the solar system because of time variations, CMEs, and flares. The solar wind generally takes two to three days to reach the Earth. When these high speed solar winds hit the Earth’s magnetic field, a free-standing shock wave, called a bow shock, is formed and the Earth’s magnetic field is compressed to a distance about 10 Earth radii on the sun-facing side because the Earth’s magnetic field acts as a hard obstacle to the solar wind [Briggs et al., 1996]. Although the bulk of the solar wind is deflected around the Earth outside the magnetopause, some of it can cross the magnetopause and enter the magnetosphere through the magnetic reconnection, which takes place when the Sun’s
IMF joins with the Earth’s magnetic field and, subsequently, the magnetic field of the Sun and Earth becomes coupled together. As a result, direct entry of solar wind plasma occurs on the day side in the vicinity of the polar cusp (Figure 2-10). Those solar wind particles travel along geomagnetic field lines and deposit their energy into the upper atmosphere. The solar wind particles also get into the tail of the magnetosphere and, along with plasma escaped from the Earth’s upper atmosphere, populate a region known as the plasma sheet. These particles directly access to the Earth’s upper atmosphere on the night side along specific magnetic field lines and, at low altitudes, join the day side polar cusp forming the auroral oval.

Geomagnetic activity significantly affects the upper atmosphere. During a geomagnetic storm time, the increased solar UV emissions heat the Earth’s upper atmosphere, causing the upper atmosphere to expand, and, as a result, the density in the upper atmosphere increases significantly. In order to understand and forecast the heat generated and the density increase due to the geomagnetic activity, it is useful to classify the disturbance levels to numerical values. Among other geomagnetic indices, the $K_p$ index (from the German Kennzahl, planetar), introduced by Chapman and Bartles [1940], is the most widely used geomagnetic planetary index for indicating the severity of the magnetic fluctuations. The $K$ variations can be defined as follows [Siebert and Meyer, 1996]:

\[
K \text{ variations are all irregular disturbances of the geomagnetic field caused by solar particle radiation within the 3-hour interval concerned. All other regular and irregular disturbances are non } K \text{ variations. Geomagnetic activity is the occurrence of } K \text{ variations.}
\]
Thirteen selected subauroral observatories at mid-latitude between 48°N and 63°S latitude measure K values with a time resolution of three hours and those thirteen K values are then averaged after applying latitude corrections in order to form the $K_p$ index. The $K_p$ index is a quasi-logarithmic, worldwide average of geomagnetic disturbances ranging from 0.0 for very quiet to 9.0 for extremely severe geomagnetic activity. It represents a sufficiently good measure for the general level of magnetic activity and thus indirectly for the intensity of the solar wind energy dissipation rate. The $a_p$ index is also widely used to represent the level of magnetic activity. The linear $a_p$ index is directly converted from the quasi-logarithmic $K_p$ index with the conversion between two indices in Table 2-4, which can be used for calculating sums and averages because the logarithmic scaled $K_p$ index is not suitable for simple averaging to obtain a

Figure 2-10 Magnetic reconnection and current systems of the Earth’s magnetosphere (courtesy of T.A. Potemra, Johns Hopkins Applied Physics Laboratory)
daily index. The mean value of eight $a_p$ values for one day in universal time (UT), $A_p$, is also widely used in the empirical models. Figure 2-7 (c) shows $A_p$ values during several solar cycles.

Table 2-4 Conversion of the $K_p$ and $a_p$ indices [Vallado and McClain, 2007]

<table>
<thead>
<tr>
<th>$K_p$</th>
<th>0.0</th>
<th>0.33</th>
<th>0.67</th>
<th>1.0</th>
<th>1.33</th>
<th>1.67</th>
<th>2.0</th>
<th>2.33</th>
<th>2.67</th>
<th>3.0</th>
<th>3.33</th>
<th>3.67</th>
<th>4.0</th>
<th>4.33</th>
</tr>
</thead>
<tbody>
<tr>
<td>$a_p$</td>
<td>0</td>
<td>2</td>
<td>3</td>
<td>4</td>
<td>5</td>
<td>6</td>
<td>7</td>
<td>9</td>
<td>12</td>
<td>15</td>
<td>18</td>
<td>22</td>
<td>27</td>
<td>32</td>
</tr>
<tr>
<td>$K_p$</td>
<td>4.67</td>
<td>5.0</td>
<td>5.33</td>
<td>5.67</td>
<td>6.0</td>
<td>6.33</td>
<td>6.67</td>
<td>7.0</td>
<td>7.33</td>
<td>7.67</td>
<td>8.0</td>
<td>8.33</td>
<td>8.67</td>
<td>9.0</td>
</tr>
<tr>
<td>$a_p$</td>
<td>39</td>
<td>48</td>
<td>56</td>
<td>67</td>
<td>80</td>
<td>94</td>
<td>111</td>
<td>132</td>
<td>154</td>
<td>179</td>
<td>207</td>
<td>236</td>
<td>300</td>
<td>400</td>
</tr>
</tbody>
</table>

While the $K_p$ index is observed at mid-latitude observatories, the $D_s$ (Disturbance Storm Time) index is based on the measurements from low-latitude observatories. The $D_s$ index (nT, nano Tesla) is the magnitude of the normalized storm time ring current in the magnetosphere (Figure 2-10). Sugiura [1964] introduced the derivation of $D_s$ values using the hourly values of the horizontal force at eight selected observatories in low latitudes in order to minimize the influence of the equatorial electrojet. Today, four observatories have been used for calculating $D_s$ values based on the quality of observation, sufficient distance from the auroral/equatorial electrojets, and distribution in longitude as evenly as possible: Hermanus (South Africa), Kakioka (Japan), Honolulu (Hawaii), and San Juan (Puerto Rico). The $D_s$ indices are archived in the World Data Center (WDC) in Kyoto, Japan. Figure 2-11 is an example of geomagnetic storm indices of $a_p$ and $D_s$ on November 2003‡‡. Typical geomagnetic storms have three phases in the view of the $D_s$ index: initial, main, and recovery phase. Initial phase is characterized by the sharp increase in $D_s$, called the storm sudden commencement, lasting a few hours only, which is caused by the compression of the magnetosphere on the arrival of a burst of solar plasma. After the sharp increase in $D_s$ for a short period, a large decrease in $D_s$ follows due to the increase of ring

‡‡ http://wdc.kugi.kyoto-u.ac.jp/dstdir/ [retrieved 4 October 2010]
current energy during the storm’s main phase building up to a maximum in about a day. In contrast to the initial and main phase of geomagnetic storm, the recovery of the storm to a normal state takes a few days with significant changes in slope; the early recovery phase has a steep slope and the late recovery phase has a gentle slope [Bowman et al., 2008b].

Figure 2-11 Variations of (a) $a_p$ and (b) $D_{st}$ geomagnetic indices during the geomagnetic storm on November 2003. The variation in $D_{st}$ values during the most geomagnetic storm time typically shows three phases; initial, main, and recovery phase.
**Thermospheric neutral density variations**

The primary energy source for heating the Earth’s thermosphere is the solar EUV emission. The photons of EUV emission from the Sun are mostly deposited in the thermosphere, causing the variations in neutral temperature and, as a result, causing the variations in neutral densities with the neutral temperature-density relation expressed in Eqs. (2.5) through (2.7). Because the solar EUV emission is totally related to the solar activity, thermospheric neutral density also varies directly with the solar activity, such as the 11-year solar cycle variation, 27-day solar rotational variation, solar flares, and CMEs, among others. The relation between solar EUV emission and thermospheric neutral densities can be identified with the relation of the neutral density and the proxy of solar EUV emission, the $F_{10.7}$ index. Figure 2-12 (a) shows the neutral density profiles simulated from the JB2006 model with two different $F_{10.7}$ values: 200 and 80 to represent solar maximum and solar minimum time, respectively (Figure 2-7 (b)). The thermospheric neutral density significantly increases with the increase of $F_{10.7}$ value, i.e. during the high solar activity. Figure 2-12 (b) represents the daily averaged neutral densities normalized to 400 km calculated from the CHAMP accelerometer data\(^8\) [Sutton et al., 2007] and $F_{10.7}$ values during the first half of the year in 2003. The variations of neutral densities show good agreements to the 27-day solar rotational variations in $F_{10.7}$ values, the proxy of solar EUV emissions. In order to study the neutral density response to the solar EUV flares and CMEs, Sutton et al. [2006] investigated the neutral density variations near 400 km and 490 km from the CHAMP and GRACE accelerometer data, respectively, for the two X-class solar flares on 28 October and 4 November, 2003. Thermospheric neutral density increases about 50–60% and 35–45% at low to mid-latitude during the first and second flare, respectively. Therefore,

---

thermospheric neutral density variations are strongly correlated with the solar activity, i.e. the variations of the mostly solar EUV emission.

After an analysis of the atmospheric neutral densities derived from satellite drag data, Paetzold and Zschomer [1961] first discovered an additional global density variation, the semiannual density variation, varying with a six-month periodicity, which maxima occurred in April and October shortly after the equinoxes and minima occurred in around January and July near the solstices. The semiannual variation is a major change in thermospheric neutral density aside from the 11-year and 27-day solar cycle variation, however, the main driving mechanism for the semiannual density variation still remains a mystery. Fuller-Rowell [1998] suggested a possible mechanism for the semiannual density variation that an asymmetric solar heating near solstices results in the strong large scale interhemisphere thermospheric circulation, the less diffusive separation due to the well mixing of lighter/heavier species, and, therefore, the decrease of the scale height because of the increase of mean molecular weight and the decrease of thermospheric neutral density at a certain pressure level. On the other hand, the interhemisphere thermospheric circulation is weak at the equinoxes, since the solar heating is more uniform than near the solstices. Rowell used the term “thermospheric spoon” for this effect in the upper atmosphere. In order to investigate the thermospheric semiannual density variation, Bowman et al. [2008c] analyzed observational data from 28 satellites with perigee heights ranging from 200 km to 1100 km during the period 1969~2006. They found that the dates of maximum and minimum neutral densities are the same within each year regardless of latitude, local solar time, or altitude. The magnitude of the variation can change from year to year by 60% during the solar minimum to over 250% during the solar maximum, and the yearly variation is strongly correlated with solar activity. Qian et al. [2009] suggested the modified eddy diffusion coefficients at the lower boundary with Fourier series of four harmonics per year in order to decrease the semiannual density errors in the numerical TIE-GCM. Figure 2-12 (c) shows the semiannual
density variation incorporated into the JB2006 model at 400 km with constant solar EUV and geomagnetic indices.

The diurnal variation or day-and-night variation in thermospheric neutral density is directly related to the local solar time since there are the differences in solar irradiances between the dark side of the Earth and the sunlit hemisphere. In general, the maximum and minimum densities occur at around 2 p.m. and 4 a.m. local solar time, respectively. In thermospheric neutral density models, local solar time or the difference in right ascension between the Sun and the local position is used as the parameter of diurnal density variation. Figure 2-12 (d) represents the diurnal density variation calculated from the JB2006 model in the location of 45° latitude, 0° longitude, and 400 km height on day 60 in 2003. During the day, the amplitude of the maximum density near 2 p.m. compared to the minimum density near 4 a.m. local solar time is the order of about 100%.
Figure 2-12 Thermospheric neutral density variations. (a) represents density variations in solar maximum and minimum time with the densities from the JB2006 model. (b) shows the 27-day solar rotational density variations with the daily averaged densities from the CHAMP accelerometer data and $F_{10.7}$ values. (c) and (d) show semiannual and diurnal density variations, respectively, with the densities from the JB2006 model.
Thermospheric neutral density models

Uncertainty in the atmospheric neutral density is a crucial error source for the prediction of the drag forces on a satellite and results in uncertainty in satellites positions in LEO below about 600 km. As a result, establishing accurate thermospheric neutral density models are important to predict the trajectory of these satellites. Thermospheric neutral density models, empirical as well as numerical models, have been developed to increase the accuracy of thermospheric neutral density estimation for several decades.

Empirical thermospheric neutral density models

Empirical thermospheric neutral density models have been developed using simplified physical concepts from in situ measurements and satellite-tracking data. Recently, thermospheric neutral density data derived from the very accurate accelerometer data aboard in the CHAMP and GRACE satellites were incorporated into some empirical models (Jacchia-class models) and, as a result, helped the models to have more accurate density estimation.

Drag Temperature Model (DTM)

Barlier et al. [1978] first developed the Drag Temperature Model (DTM-78) using the measurements of exospheric temperature and atmospheric densities derived from satellite drag data. However, it is not representative of low solar activity condition due to the data set used, which causes uncertain predictions during low solar activity. With extended data and an improved algorithm, it was refined to respond well especially under extreme solar and geomagnetic conditions (DTM-94) [Berger et al., 1998]. The lower boundary condition both the DTM-78 and DTM-94 models were set to constant temperature and its vertical gradient at 120
km. Since significant variations of temperature at 120 km have been found from incoherent scatter radar observations, these constant boundary conditions cause the model to have large uncertainties in the lower thermosphere. Bruinsma et al. [2003] improved the DTM-94 model using the temperature with incoherent scatter radar observations, and temperature gradient with temperature/wind data from the upper atmosphere research satellite. As a result, the temperature gradient was modeled as a function of local time, latitude, and season at 120 km (DTM-2000). Atmospheric Explorer satellite (AE-C and -E) data were also incorporated into the model in order to improve the prediction of temperature under low solar activity conditions.

**NASA Marshall Engineering Thermosphere (MET) model**

In order to calculate atmospheric properties for altitudes between 90 km and 2,500 km, the NASA Marshall Space Flight Center (MSFC) used the MSFC/J70 [Johnson and Smith, 1985], which was similar to the J70 model. The model has an operational limitation, as it could only be run on a specific computer since the model was coded in Fortran IV. Hickey [1988] first developed the formal version of MET model (MET-1988) which is an essentially a modified J70 model with the incorporation of spatial and temporal variations from the J71 model. It improved the calculation of atmospheric thermodynamic properties and updated the entire program from Fortran IV to Fortran 77 so that it could be executed on other computers. The model was improved to MET-1999 so as to correctly adapt to the change of the new millennium by including some improvements in the calculation of the solar coordinates among others [Owens et al., 2000]. The latest version of the model is the MET-V2.0 model and was improved in 2002 [Owens, 2002]. The MET-V2.0 model provides total mass density, temperature, and individual species number densities in the altitude range between 90 km and 2,500 km. Compared to the other models, it
uses the $F_{10.7}$ index with the 162-day averaged 10.7 cm solar flux. The MET-V2.0 is used by a number of aerospace organizations.

**Jacchia-class model**

The empirical Jacchia-class models, first developed in 1965 (J65) with the lower altitude boundary at 120 km [Jacchia, 1965], are more operational than the MSIS-class models, and are used in tracking and predicting satellite orbital behavior. The Jacchia-class models are mostly based on drag data obtained by observing the orbital motion of numerous satellites. The lower altitude boundary was lowered to 90 km in 1970 (J70) [Jacchia, 1970] and the J70 model was reformulated by using newer and more complete data in 1971 (J71) [Jacchia, 1971]. In 1977, Jacchia developed the J77 model by including satellite mass spectrometer data and revising some equations in the J71 model [Jacchia, 1977]. In 2006, Bowman et al. [2008a] improved the Jacchia-class model to the JB2006 model by incorporating new solar indices $S_{10.7}$ and $M_{10.7}$ in order to calculate global minimum nighttime exospheric temperature. New exospheric temperature and semiannual density equations for representing thermospheric density variations, temperature correction equations for diurnal and latitudinal effects, and density correction factors for model corrections required at high altitude were included in the JB2006 model. The latest version of the Jacchia-class models is the JB2008 model which includes a new solar index, $Y_{10.7}$, and geomagnetic index, $D_{st}$ [Bowman et al., 2008b].

**MSIS-class model**

Satellite mass spectrometer and ground-based incoherent scatter radar data are the major data sources of the empirical MSIS-class models, which is widely used in the research
community. Hedin et al. [1977] first developed the MSIS-1 model with the lower altitude boundary at 120 km. Since the data sets are mostly from low to medium solar activity conditions, it has a weakness under high solar activity conditions. Hedin [1983] incorporated data from several rocket flights, satellites, and incoherent scatter radars during high solar activity conditions, lowered the altitude boundary to 90 km, and named as the MSIS-83 model. In addition to the daily averaged $A_p$ index, the 3-hourly $a_p$ indices were adopted to the model. Hedin developed the MSIS-86 model by adding and changing terms to better represent seasonal variations in the polar regions [Hedin, 1987], and the lower altitude boundary was further extended to the Earth’s surface in the MSISE-90 model (E stands for extension) [Hedin, 1991]. The U.S. Naval Research Laboratory (NRL) improved the MSISE-90 model to the NRLMSISE-00 model by including total mass density data from satellite accelerometers and orbit determination, molecular oxygen number density data from solar UV occultation aboard the SMM, and temperature data from incoherent scatter radar (ISR) [Picone et al., 2002].

**Numerical thermospheric neutral density models**

In contrast to the empirical models, physics-based numerical models try to account for the complex evolution of the thermosphere closely coupled with the ionosphere, the magnetosphere, and the mesosphere energetically, dynamically, and chemically. Conservation laws (mass, momentum, and energy) and atmospheric constituent models were incorporated into physical density models.
**Coupled Thermosphere Ionosphere Model (CTIM)**

The CTIM [Fuller-Rowell et al., 1987, 1996] evolved from a three-dimensional, time-dependent, global model of the thermosphere developed at University College London [Fuller-Rowell and Rees, 1980, 1983] and an ionospheric model developed at Sheffield University [Quegan et al., 1982] coupling the thermosphere and ionosphere through energy and momentum interchange. The model calculates thermospheric parameters with a 1-min time step on a grid of 20 longitude bins (every 18°), 91 latitude bins (every 2°), and 15 constant pressure level surfaces with a vertical resolution of one scale height from 80 km to around 300–700 km depending on solar activity. The CTIM assumes hydrostatic equilibrium for pressure levels and solves the energy, momentum, and continuity equations for the three major neutral species O, N₂, and O₂ as well as other minor neutral species. The CTIM calculates ions O⁺, H⁺, O₂⁺, N₂⁺, and NO⁺ with the assumption of chemical equilibrium. Millward et al. [1996] developed the Coupled Thermosphere-Ionosphere-Plasmasphere (CTIP) model by adding a low-latitude, self-consistent plasmasphere into the CTIM extended to 10,000 km. The CTIP model calculates H⁺, O⁺ ion densities and temperatures ranging from 100 km to 10,000 km and N₂⁺, O₂⁺, NO⁺, N⁺ ion densities below 400 km. The self-consistent electrodynamics solver for the mid- and low-latitude electropotential from Richmond and Roble [1987] was incorporated into the CTIP model and named CTIPe (Coupled Thermosphere-Ionosphere-Plasmasphere electrodynamics) model [Millward et al., 2001]. The Coupled Middle Atmosphere and Thermosphere (CMAT) model evolved from the CTIP model covering a vertical range from around 30 km to around 300–600 km depending on solar activity [Harries et al., 2002]. The CMAT model uses outputs from the MSISE-90 model for lower boundary seasonal forcing [Hedin, 1991] and outputs from the Global Scale Wave Model (GSWM) [Hagan et al., 1999] for lower boundary tidal forcing.
Global Ionosphere-Thermosphere Model (GITM)

Ridely et al. [2006] developed the Global Ionosphere-Thermosphere Model (GITM) using a three-dimensional spherical grid system. The GITM has a fixed resolution in longitude (5°), but has adjustable, non-uniform resolutions in both latitude and altitude coordinates. The most noticeable feature of the GITM is that it does not use the assumption of hydrostatic equilibrium for solving dynamics equations, which allows the model to more realistically capture physics in the high-latitude region, where auroral heating is prevalent. Unlikely to the other models using pressure-based vertical coordinate system, the GITM uses an altitude-based coordinate system. The code of the GITM can be run in either 1-D or 3-D mode.

Thermosphere-Ionosphere-Electrodynamic General Circulation Model (TIE-GCM)

The NCAR’s TIE-GCM is one of the foremost models of three-dimensional, time-dependent numerical models, which solves continuity, momentum, and energy equation of the upper atmosphere self-consistently with the assumption of the hydrostatic equilibrium. It extends from around 97 km (assumed level of peak concentrations of atomic oxygen) to around 600 km depending on solar activity with 0.5 scale vertical height resolution. The first version of the NCAR’s general circulation model, TGCM (Thermospheric General Circulation Model), was developed in 1981 and focused on only the thermosphere above about 100 km [Dickinson et al., 1981]. The TGCM used a global empirical model (MSIS-1) to specify ion drag and the neutral gas background properties such as gas mixture molecular conductivity, viscosity, specific heat, and mean scale height, among others. The coupling of dynamics with neutral composition described by Dickinson and Ridley [1972, 1975] was included by extending the model equations of the TGCM [Dickinson et al., 1984]. Roble et al. [1988] included a self-consistent aeronomic
scheme of the thermosphere and ionosphere developed by Roble et al. [1987] and Roble and Ridley [1987] and named the TIGCM (Thermosphere-Ionosphere General Circulation Model). The TIGCM calculates electron/ion density and ion temperature with winds, neutral temperature and composition. The scheme for electrodynamics coupling between the thermosphere and the ionosphere was developed and implemented into the TIGCM [TIE-GCM, Richmond et al., 1992]. The TIE-GCM includes vertical propagation of semi-diurnal and diurnal tidal perturbations from the mesosphere to the thermosphere caused by solar radiation specified by the climatological tidal data or the GSWM for the lower boundary. The ionospheric convection pattern is specified using the empirical model developed by Heelis et al. [1982]. The TIE-GCM calculates the global distribution of neutral temperature, zonal/meridional/vertical winds, mass mixing ratio of major (N\textsubscript{2}, O\textsubscript{2}, and O) and minor (N(\textsuperscript{2}D), N(\textsuperscript{4}S), and NO) neutral constituents for the neutral atmosphere, and ion/electron temperature, electron number densities, and ionic constituents (O\textsuperscript{+}, O\textsubscript{2}\textsuperscript{+}, N\textsuperscript{+}, N\textsubscript{2}\textsuperscript{+}, and NO\textsuperscript{+}). However, helium and hydrogen and their ions are not presently included in the TIE-GCM. Roble and Ridley [1994] lowered the lower altitude boundary of the TIE-GCM from around 97 km to 30 km including the physical and chemical processes for the mesosphere and the upper stratosphere, and named the TIME-GCM (Thermosphere-Ionosphere-Mesosphere-Electrodynamics General Circulation Model). The TIME-GCM calculates the long-lived minor neutral constituents H\textsubscript{2}O, H\textsubscript{2}, CH\textsubscript{4}, CO, and CO\textsubscript{2}, but excludes Cl\textsubscript{x}, N\textsubscript{2}O, NO\textsubscript{3}, N\textsubscript{2}O\textsubscript{5}, and so forth, since the stratospheric chemistry is limited.

**Thermosphere Ionosphere Nested Grid (TING) model**

The three-dimensional, time-dependent, high resolution, nested grid model (TING model) which is an extension of the NCAR’s TIGCM, was developed to study meso-scale and small-scale processes in the coupled thermosphere-ionosphere system [Wang et al., 1999]. The TING
model simultaneously calculates global (coarse grid, TIGCM) and local (nested grid) distributions of neutral temperatures, winds, neutral constituents, and ionospheric parameters with the same solvers as the TIGCM but different resolutions between the two grid system. The vertical resolution of the nested grid of the TING model cannot be increased and is the same as the resolution of the coarse grid, which are 25 constant pressure levels between around 97 km and around 500 km. The horizontal resolution for the global coarse grid is 5° latitude by 5° longitude, and that for the nested grid is three times higher than the coarse grid’s (1.67° for the first level resolution). The TING model can adjust the size and location of the nested grid suitable to the specific application. The TING model is typically used in the northern hemisphere auroral region to study geomagnetic activity at high latitude.

In this dissertation, the Jacchia-class models (JB2006 and JB2008) and the NRLMSISE-00 model for the empirical model were chosen to compare the models’ accuracies since those models are widely used. In addition, it was shown that the JB2008 model is the most accurate empirical model in thermospheric neutral density estimation. On the other hand, the NRLMSISE-00 model calculates major neutral constituents’ number densities as well as total mass density and neutral temperature. The primary model of this dissertation is the TIE-GCM due to the reason that it is the most common theoretical model for the Earth’s thermosphere and ionosphere and it is suitable to study the physics in the thermosphere. This dissertation uses version 1.92 of the TIE-GCM. Both empirical and theoretical models are further described in Chapter 3.
Chapter 3

MODEL DESCRIPTIONS

In this chapter, the JB2006, JB2008, NRLMSISE-00, and TIE-GCM models are described in detail. The JB2006 and JB2008 models are more operational than the NRLMSISE-00 model, and are used for tracking and predicting satellites’ orbital behavior. They are mostly based on drag data from the orbital motion of numerous satellites. They calculate total mass density, neutral temperature at the height of interest, and exospheric temperature. On the other hand, the major data sources of the NRLMSISE-00 model are the satellite mass spectrometer and ground-based incoherent scatter radar data. It is widely used in the research community, and calculates major species’ number densities from the surface of the Earth in addition to the total mass density and neutral temperature unlike to the JB2006 and JB2008 model which the lower altitude boundary is 90 km. The TIE-GCM solves the Eulerian continuity, momentum, and energy equations for the coupled thermosphere/ ionosphere system. It calculates concentrations of major as well as minor neutral constituents, concentrations of ions, electron density, and wind components in the altitude region of approximately 97 km ~ 600 km depending on solar activity.

Jacchia-class models

The main drawback of the first Jacchia-class model (J65) is the constant boundary conditions at 120 km altitude. The lower altitude boundary conditions were lowered to 90 km (J70), newer and more complete data were incorporated into the model (J71), and some equations were revised with satellite mass spectrometer data (J77). In 2006, new solar indices S_{10.7} and
M_{10.7} were included for global minimum nighttime temperature calculation (JB2006). New exospheric temperature and semiannual density equations, temperature corrections for diurnal and latitudinal effects, and density corrections factors at high altitude were included in the JB2006 model. The JB2006 model was improved to the JB2008 model with a new solar index Y_{10.7} and geomagnetic index D_{st}. The JB2008 model has been shown that the model provides the highest accurate thermospheric neutral densities among other thermospheric density models [Bowman et al., 2008a, 2008b; Tobiska et al., 2008]. This is possibly due to the contribution of the D_{st} index representing the energy deposit into the high-latitude thermosphere during geomagnetic storms and sub-storms. Because of the highest accuracy in thermospheric neutral density estimation of the JB2008 model, the U.S. Air Force (USAF) plans to operationally implement this model [Tobiska, 2010] and the Committee on Space Research (COSPAR)’s International Reference Atmosphere (CIRA-08) model adopted total mass density profiles from the JB2008 model above 120 km altitude [ANS, 2008]. Space Environment Technologies (SET) provides total mass densities in LEO satellite operations in their LEO Alert and Prediction System (LAPS)^* which uses thermospheric densities from the JB2008 model [Tobiska and Bouwer, 2010].

**JB2006 model**

The JB2006 model was developed from the previous Jacchia’s models. New density data from the analysis of satellite drag data were used to develop new formula for the neutral temperature and the neutral density. New solar indices were included to better represent thermospheric responses to the solar activity and new equations for the semiannual density variations were developed.

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^* http://spacewx.com/Products.html [retrieved 10 January 2011]
Data reduction

The primary data set of the JB2006 model is from the satellite drag data. Very accurate daily densities were computed by the method developed by Bowman et al. [2004] from the drag data of numerous satellites at the perigee altitudes from 175 km to 1,100 km. Daily temperature values were also computed from these satellite drag data using a special energy dissipation rate method [Bowman et al., 2004] in the period from 1978 through 2004. Density values calculated by this method show less than 4% uncertainties.

Model formulation

The input arguments of the JB2006 model are the modified Julian day, right ascension and declination of the Sun, latitude/longitude, \( F_{10.7}/S_{10.7} \) for the previous day, \( M_{10.7} \) for 5 days earlier, and \( F_{10.7A}/S_{10.7A}/M_{10.7A} \), and \( a_p \) index for 6.7 hours earlier. It assumed that the homosphere extends to 105 km and becomes the heterosphere above that height. As a result, the barometric equation is used to calculate neutral density in the mixing region of 90–105 km and the diffusion equation is used in the diffusion region above 105 km. The number density ratio of \( n(O)/n(O_2) \) is assumed to be 1.5 at 120 km for the best fit to satellite drag data.

The atmosphere is assumed to be composed of only nitrogen, oxygen, argon, helium, and hydrogen, although hydrogen is included in the region only above 500 km. In the homosphere, the mean molecular weight is calculated by Eq. (3.1), based on the sea level molecular weight [Jacchia, 1970],

\[
\bar{m}(h) = \sum_{n=0}^{6} c_n (h - 100)^n; \quad (90 \text{ km} < h < 105 \text{ km})
\] (3.1)
where the coefficients are:

\[
\begin{align*}
  c_0 &= 28.15204 \\
  c_1 &= -0.085586 \\
  c_2 &= 1.2840 \times 10^{-4} \\
  c_3 &= -1.0056 \times 10^{-5} \\
  c_4 &= -1.0210 \times 10^{-5} \\
  c_5 &= 1.5044 \times 10^{-6} \\
  c_6 &= 9.9826 \times 10^{-8}
\end{align*}
\]

The sea level composition of the Jacchia-class model is from the U.S. Standard Atmosphere 1962 [COESA, 1962] and the mean molecular weight at sea level is assumed to be 28.960 g/mol.

**Computation of neutral temperature**

The process for computing neutral temperature in the JB2006 model is to compute (a) the global minimum nighttime exospheric temperature, (b) the local exospheric temperature with no geomagnetic conditions, (c) the geomagnetic activity correction, (d) the daily temperature correction, (e) the inflection point temperature, and (f) the local temperature at a point of interest.

- **Global minimum nighttime exospheric temperature \( (T_c) \)**

  The \( T_c \) was computed with only \( F_{10.7} \) and \( F_{10.7A} \) used in the previous Jacchia-class model [Jacchia, 1965, 1970, 1971, 1977]. New solar indices, \( S_{10.7} \) and \( M_{10.7} \), were adopted in the \( T_c \) equation for better density variation correlations with UV radiation. Equation (3.2) is the \( T_c \) equation in the JB2006 model, where the \( \Delta \) prefix represents the difference between the daily and 81-day average value of each solar index.

\[
T_c = 379.0 + 3.353 \ F_{10.7A} + 0.358 \ \Delta F_{10.7} + 2.094 \ \Delta S_{10.7} + 0.343 \ \Delta M_{10.7} \quad (3.2)
\]

- **Local exospheric temperature with no geomagnetic conditions \( (T_l) \)**

  The \( T_c \) is the exospheric temperature without any corrections such as local solar time, geomagnetic activity, among others. After applying the diurnal variation by the
correction of local solar time, $T_i$ is derived from $T_c$ based on no geomagnetic activity ($a_p = 0$). The assumption for the calculation of $T_i$ is that the maximum daytime and the minimum nighttime exospheric temperatures occur at a latitude, $\phi$, equal to the declination of the Sun, $\delta_\odot$, on the day side and at $\phi$ equal to $-\delta_\odot$ on the night side of the Earth, respectively. The ratio of the maximum to the minimum temperature $(1+R)$ varies with solar cycle, but it was shown that a good average for the ratio is $R = 0.31$. Consequently, $T_i$ is calculated by Eq. (3.3) with the input angles explaining the position of interest determined by Eqs. (3.4) through (3.6),

$$T_i = T_c (1 + R \sin^m \theta) \left( 1 + R \frac{\cos^m \eta - \sin^m \theta}{1 + R \sin^m \theta} \cos^n \frac{\tau}{2} \right)$$ \hspace{1cm} (3.3)

$$\eta = \frac{1}{2} |\phi - \delta_\odot|$$ \hspace{1cm} (3.4)

$$\theta = \frac{1}{2} |\phi + \delta_\odot|$$ \hspace{1cm} (3.5)

$$\tau = H + \beta + p \sin (H + \gamma)$$ \hspace{1cm} (3.6)

where the parameters are:

$$H = \text{hour angle of the Sun} \hspace{1cm} \beta = -37^\circ$$

$$m = 2.5 \hspace{1cm} p = 6^\circ$$

$$n = 3.0 \hspace{1cm} \gamma = 43^\circ$$

- **Geomagnetic activity correction ($\Delta T_g$)**

The JB2006 model has adopted the approximated equation for representing geomagnetic activity effects on the exospheric temperature changes developed by Jacchia et al. [1967]. After analyzing temperature variations according to geomagnetic activity conditions from the several satellites with perigee heights of 250 km ~ 615 km and apogee heights 800 km ~ 2,500 km, they developed Eq. (3.7) for representing geomagnetic perturbations on
the exospheric temperature. Equation (3.7) shows good approximation for the perturbations, although the uncertainty at higher latitudes above 55° is larger than those at latitudes below 55°. They also found a time delay between the geomagnetic activity and the temperature perturbations. The average time lag at all latitudes is 0.28±0.012 day, i.e. 6.7±0.3 hours and a smaller mean time lag at higher latitudes (0.24±0.02 day for \(|\phi| \geq 55^\circ\)) than that at lower latitudes (0.30±0.01 day for \(|\phi| < 55^\circ\)). The JB2006 model adopted the average time lag 6.7 hours prior to the time of interest.

\[
\Delta T_g = \begin{cases} 
28^\circ K_p + 0.03^\circ e^{K_p} & \text{for } K_p \text{ index} \\
1.0^\circ a_p + 100^\circ (1 - e^{-0.98 a_p}) & \text{for } a_p \text{ index}
\end{cases}
\] (3.7)

- **Daily temperature correction (\(\Delta T_c\) or \(\Delta T_x\))**

After the correction of the geomagnetic activity, the temperature profile still showed significant error varying with height. Therefore, it was necessary to correct that error by adding temperature variations to the exospheric temperature or inflection point temperature. The Jacchia-models prior the JB2006 model considered that as a function of local solar time and latitude only. However, Bowman et al. [2008a] developed the polynomial fits with least squared method as a function of local solar time, latitude, height, and F_{10.7}. They divided three height regions for this correction: Eq. (3.8) for 250 km ~ 700 km, Eq. (3.9) for 200 km ~ 250 km, and Eq. (3.10) for 140 km ~ 200 km. The temperature corrections above 200 km (\(\Delta T_c\)) are added to the exospheric temperature. On the other hand, those below 200 km (\(\Delta T_x\)) are added to the inflection point temperature at 125 km, since it is better than the exospheric temperature correction for these very low heights.
- For $h = 250$ km ~ $700$ km, $H=\frac{h}{100}$
\[
\Delta T_c = B_1 + F \left( B_2 + B_3 \theta + B_4 \theta^2 + B_5 \theta^3 + B_6 \theta^4 + B_7 \theta^5 \right) \\
+ \varphi \left( B_9 \theta + B_{10} \theta^2 + B_{11} \theta^3 + B_{12} \theta^4 \right) \\
+ \varphi H \left( B_{13} + B_{14} \theta + B_{15} \theta^2 + B_{16} \theta^3 + B_{17} \theta^4 + B_{18} \theta^5 \right) + B_{19} \varphi
\]

- For $h = 200$ km ~ $250$ km, $H=\frac{(h-200)}{50}$
\[
\Delta T_c = H C_1 + HF \left( C_2 + C_3 \theta + C_4 \theta^2 + C_5 \theta^3 + C_6 \theta^4 + C_7 \theta^5 \right) \\
+ H \varphi \left( C_9 \theta + C_{10} \theta^2 + C_{11} \theta^3 + C_{12} \theta^4 + C_{13} + C_{14} F + C_{15} F \theta + C_{16} F \theta^2 \right) \\
+ C_{17} + \varphi \left( C_{18} \theta + C_{19} \theta^2 + C_{20} \theta^3 + C_{21} F + C_{22} F \theta + C_{23} F \theta^2 \right)
\]

- For $h = 140$ km ~ $200$ km, $H=\frac{h}{100}$
\[
\Delta T_x = D_1 + \left( D_2 \theta + D_3 \theta^2 + D_4 \theta^3 + D_5 \theta^4 \right) \\
+ H \left( D_6 + D_7 \theta + D_8 \theta^2 + D_9 \theta^3 \right) + F \left( D_{10} + D_{11} \theta + D_{12} \theta^2 \right)
\]

where the values of coefficients $B$, $C$, and $D$ are listed in Table 3-1, and
\[
F = \frac{(F_{10.7} - 100)}{100} \quad \theta = \frac{\text{(local solar time in hour)}}{24} \quad \varphi = \cos(\phi)
\]

Finally, with these temperature and temperature corrections, the local exospheric temperature, $T_\infty$, is obtained with the summation of them ($T_\infty = T_l + \Delta T_g + \Delta T_c$).

- **Inflection point temperature ($T_x$)**

In order to represent the vertical temperature profile in the thermosphere, two local parameters were defined: a low altitude parameter (inflection point temperature, $T_x$) for the temperature at 125 km and a high altitude parameter (exospheric temperature, $T_\infty$) for asymptotic temperature in the exosphere. The $T_\infty$ is determined with $T_c$, $T_h$, $\Delta T_g$, and $\Delta T_c$, and, subsequently, the $T_x$ is calculated by Eq. (3.11) using the exospheric temperature.
$$T_x = 444.3807 + 0.02385 \ T_\infty - 392.8292 \ e^{(-0.0021357 \ T_\infty)}$$

(3.11)

Table 3-1 Coefficient values of the equations for the daily temperature correction [Bowman et al., 2008a]

<table>
<thead>
<tr>
<th>Coefficient</th>
<th>$B$</th>
<th>$C$</th>
<th>$D$</th>
</tr>
</thead>
<tbody>
<tr>
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<td>-0.457512297E+01</td>
<td>-0.155986211E+02</td>
<td>0.828727388E+00</td>
</tr>
<tr>
<td>2</td>
<td>-0.512114909E+01</td>
<td>-0.512114909E+01</td>
<td>0.124730376E+02</td>
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<tr>
<td>3</td>
<td>-0.693003609E+02</td>
<td>-0.693003609E+02</td>
<td>-0.318077422E+03</td>
</tr>
<tr>
<td>4</td>
<td>0.203716701E+03</td>
<td>0.203716701E+03</td>
<td>0.645905237E+03</td>
</tr>
<tr>
<td>5</td>
<td>0.703316291E+03</td>
<td>0.703316291E+03</td>
<td>-0.322577619E+03</td>
</tr>
<tr>
<td>6</td>
<td>-0.194349234E+04</td>
<td>-0.194349234E+04</td>
<td>0.709584573E+01</td>
</tr>
<tr>
<td>7</td>
<td>0.110651308E+04</td>
<td>0.110651308E+04</td>
<td>0.126347673E+02</td>
</tr>
<tr>
<td>8</td>
<td>-0.174378996E+03</td>
<td>-0.220835117E+03</td>
<td>-0.744574342E+02</td>
</tr>
<tr>
<td>9</td>
<td>0.188594601E+04</td>
<td>0.143256989E+04</td>
<td>0.511241177E+02</td>
</tr>
<tr>
<td>10</td>
<td>-0.709371517E+04</td>
<td>-0.318481844E+04</td>
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<td>-0.223771015E+02</td>
</tr>
<tr>
<td>13</td>
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<td>0.199956489E+02</td>
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<tr>
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<td>-0.127093998E+02</td>
<td></td>
</tr>
<tr>
<td>15</td>
<td>-0.482006560E+03</td>
<td>0.212825156E+02</td>
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<td>16</td>
<td>0.181870931E+04</td>
<td>-0.275555432E+01</td>
<td></td>
</tr>
<tr>
<td>17</td>
<td>-0.237389204E+04</td>
<td>0.110234982E+02</td>
<td></td>
</tr>
<tr>
<td>18</td>
<td>0.996703815E+03</td>
<td>0.148881951E+03</td>
<td></td>
</tr>
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<td>19</td>
<td>0.361416936E+02</td>
<td>-0.751640284E+03</td>
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<td>20</td>
<td>0.637876542E+03</td>
<td></td>
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</tr>
<tr>
<td>21</td>
<td>0.127093998E+02</td>
<td></td>
<td></td>
</tr>
<tr>
<td>22</td>
<td>-0.212825156E+02</td>
<td></td>
<td></td>
</tr>
<tr>
<td>23</td>
<td>0.275555432E+01</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

- **Local temperature at a specific height ($T_h$)**

The lower boundary condition of the Jacchia-class model is the constant temperature ($T_0 = 183$ K) at the height $h_0 = 90$ km. The temperature gradient ($G = dT/dh$) at the lower boundary is set to $G_0 = 0$. The temperature and temperature gradient increase from the boundary condition with height and the temperature gradient at the inflection point is
defined as Eq. (3.12) with the factor 1.90 for the best fits. Lastly, the local temperature at
$h$ can be calculated by Eq. (3.13) below $h_x$ with $T_x$ and $G_x$, and by Eq. (3.14) above $h_x$
with $T_x$, $G_x$, and $T_\infty$, where $h_x$ is 125 km.

$$G_x = \left( \frac{dT}{dh} \right)_{h=h_x} = 1.90 \frac{T_x - T_0}{h_x - h_0} \quad (3.12)$$

$$T_h = T_x + G_x (h - h_x) \left[ \{ -9.8204695 \times 10^{-6} (h - h_x) - 7.3039742 \\ \times 10^{-4} \} (h - h_x)^2 + 1 \right] \quad (3.13)$$

$$T_h = T_x + \frac{2}{\pi} (T_\infty - T_x) \tan^{-1} \left[ \frac{\pi G_x}{2 (T_\infty - T_x)} (h - h_x) \{ 1 + 4.5 \times 10^{-6} (h - h_x)^2 \}^2 \right] \quad (3.14)$$

**Computation of neutral density**

Using the temperature profiles derived from Eqs. (3.13) and (3.14), the JB2006 model
calculates the thermospheric neutral densities based on the assumption of two distinct
atmospheric regions: the mixing region up to 105 km and the diffusion region above 105 km.
Hydrogen density is further considered above 500 km in the diffusion region. Then, seasonal-
latitudinal and semiannual density variations are applied for the neutral densities.

- **Mixing region density calculation**

In the mixing region between 90 km and 105 km, the barometric equation expressed by
Eq. (3.15) is integrated to obtain the thermospheric neutral density, where $R$ is the
universal gas constant ($R$=8.31432 joules K$^{-1}$ mol$^{-1}$), $\bar{m}$ is calculated by Eq. (3.1), and the
gravitational acceleration as a function of height is obtained by Eq. (3.16).

$$d\ln(\rho) = d\ln \left( \frac{\bar{m}}{T} \right) - \frac{\bar{mg}}{RT} dh \quad (3.15)$$
Using the total thermospheric neutral density calculated from Eq. (3.15), the total number of particles per unit volume \( (N) \) is obtained by Eq. (3.17), where \( A \) is Avogadro’s number \( (A=6.02257\times10^{23} \text{ mol}^{-1}) \). Number densities of individual species, \( n(i) \), can be computed with the total number of particles. \( N_2, \) Ar, and He number densities are obtained by Eq. (3.18), \( O \) number density by Eq. (3.19), and \( O_2 \) number density by Eq. (3.20). \( q_0(i) \) is the volume fraction of individual species at sea level and \( \bar{m}_0 \) is the mean molecular weight of air at sea level \( (\bar{m}_0 = 28.960 \text{ g/mol}) \).

\[
N = \frac{A \rho}{\bar{m}}
\]

(3.17)

\[
n(i) = q_0(i) \frac{\bar{m}}{\bar{m}_0} N
\]

(3.18)

\[
n(O) = 2N \left(1 - \frac{\bar{m}}{\bar{m}_0}\right)
\]

(3.19)

\[
n(O_2) = N \left(\frac{\bar{m}}{\bar{m}_0} [1 + q_0(O_2)] - 1\right)
\]

(3.20)

- **Diffusion region density calculation**

In contrast to the mixing region below 105 km, the number densities of individual species are first calculated, and then total thermospheric neutral density is obtained by the summation of the number densities of individual species in the diffusion region above 105 km. The number density of individual species can be computed by integrating the diffusion equation expressed by Eq. (3.21), where \( m_i \) is the molecular weight of the individual species and \( \alpha_i \) is the thermal diffusion coefficient of the individual species. The JB2006 model adopted \( \alpha = -0.38 \) for helium and \( \alpha = 0 \) for everything else.
\[
\frac{dn(i)}{n(i)} = - \frac{m_i g}{RT} dh - \frac{dT}{T} (1 + \alpha_i)
\]  
(3.21)

The JB2006 model includes hydrogen number density above 500 km based on the assumption of diffusive equilibrium at the altitude region. The concentration of hydrogen at 500 km is calculated by Eq. (3.22) and no hydrogen number densities are included below 500 km.

\[
\log_{10} n(H)_{500} = 73.13 - 39.40 \log_{10} T_\infty + 5.5 (\log_{10} T_\infty)^2
\]  
(3.22)

- **Correction of seasonal-latitudinal density variation**

The seasonal-latitudinal density variations occur in the lower thermosphere below around 200 km. The maximum amplitude of the seasonal-latitudinal density variations appear in the altitude between 105 km and 120 km [Jacchia, 1970]. Equation (3.23) represents these density variations in the JB2006 model, where \( Y \) is days in a year (365 or 366) and \( d \) is the day of year since January 1,

\[
\Delta \log_{10} \rho = 0.02 (h - 90) \frac{\phi}{|\phi|} e^{-0.045 (h - 90)} \sin^2 \phi \sin \left[ \frac{360^\circ}{Y} (d + 100) \right]
\]  
(3.23)

- **Correction of semiannual density variation**

In order to represent the semiannual density variations, the JB2006 model adopted the form expressed by Eq. (3.24) from the J71 [Jacchia, 1971] model, modifying the components of the equation, where \( z = \text{height (km)} / 1,000 \) and \( t = \text{day of year} \).

\[
\Delta \log_{10} \rho = F(z) G(t)
\]  
(3.24)
The amplitude of the semiannual density variation, $F(z)$, is the difference between the yearly maximum and the yearly minimum log$_{10}$ neutral density values as a function of height. Bowman [2004] modified and simplified Jacchia’s $F(z)$ equation to a quadratic polynomial equation without losing any fidelity. Bowman et al. [2008a] further modified the equation expressed by Eq. (3.25) so as to cover all years and all heights. The coefficients of Eq. (3.25) are listed in Table 3-2.

$$F(z) = B_1 + B_2 F_{10.7A} + B_3 F_{10.7A}^2 + B_4 F_{10.7A} z + B_5 F_{10.7A}^2 z^2 + B_6 F_{10.7A}^2 z^2 \quad (3.25)$$

Table 3-2 $B_i$ coefficients of Eq. (3.25) [Bowman et al., 2008a]

<table>
<thead>
<tr>
<th>$i$</th>
<th>$B_i$</th>
<th>$i$</th>
<th>$B_i$</th>
</tr>
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<tr>
<td>1</td>
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<td>-1.00064E-02</td>
</tr>
<tr>
<td>2</td>
<td>-1.59000E-03</td>
<td>5</td>
<td>-2.37509E-05</td>
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<tr>
<td>3</td>
<td>1.26190E-02</td>
<td>6</td>
<td>2.60759E-05</td>
</tr>
</tbody>
</table>

The average density variation as a function of time, $G(t)$, was normalized to the amplitude 1 with the difference between the yearly maximum and the year minimum values for the year. Bowman [2004] modified Jacchia’s $G(t)$ equation to the nine-term Fourier series. Later, it was expanded for global coverage with including the $F_{10.7A}$ values expressed by Eq. (3.26) since the $G(t)$ showed the dependency on solar activity [Bowman et al., 2008a]. Here, $\omega = 2\pi\theta$ and $\theta = (t - 1.0)/365$. The least-squared fitted values for the coefficients are listed in Table 3-3. Equations (3.24), (3.25), and (3.26) represent the semiannual density variations in the JB2006 model.
Figure 3-1 shows the computation flow of thermospheric neutral mass density and neutral temperature on the JB2006 model [Kim, 2008].

\[
G(t) = C_1 + C_2 \sin(\omega) + C_3 \cos(\omega) + C_4 \sin(2\omega) + C_5 \cos(2\omega) \\
+ C_6 \sin(3\omega) + C_7 \cos(3\omega) + C_8 \sin(4\omega) + C_9 \cos(4\omega) \\
+ F_{10.7A} \left( C_{10} + C_{11} \sin(\omega) + C_{12} \cos(\omega) + C_{13} \sin(2\omega) + C_{14} \cos(2\omega) \right) \\
+ F_{10.7A}^2 \left( C_{19} + C_{20} \sin(\omega) + C_{21} \cos(\omega) + C_{22} \sin(2\omega) + C_{23} \cos(2\omega) \right) 
\]

(3.26)

Table 3-3 \(C_i\) coefficients of Eq. (3.26) [Bowman et al., 2008a]

<table>
<thead>
<tr>
<th>(i)</th>
<th>(C_i)</th>
<th>(i)</th>
<th>(C_i)</th>
<th>(i)</th>
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<td>0.206763E-02</td>
<td>21</td>
<td>0.134078E-04</td>
</tr>
<tr>
<td>6</td>
<td>-0.165537E-01</td>
<td>14</td>
<td>-0.142888E-02</td>
<td>22</td>
<td>-0.614176E-05</td>
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<tr>
<td>7</td>
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<td>15</td>
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<td>16</td>
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<td></td>
</tr>
</tbody>
</table>
Figure 3-1 Computation flow of total mass density and neutral temperature on the JB2006 model
**JB2008 model**

The JB2008 model was developed from the JB2006 model by including new solar and geomagnetic indices. The new formula for the neutral temperature calculation and the semianual density variations were developed based on the analysis of recent drag data from numerous satellites. The accuracy of the neutral density estimation during the geomagnetic storm time was improved by using these new indices.

**Data reduction**

The JB2008 model was developed from the JB2006 model by using four different thermospheric neutral density data sources [Bowman et al., 2008b]. One of the density data sources is the same as that used in developing the JB2006 model, but the data are more recent ones from the period 1997 through 2007 compared to the period 1978 through 2004 in the JB2006 model. The second data source is the density values obtained from the Air Force Space Command’s High Accuracy Satellite Drag Model (HASDM) processes [Storz et al., 2002; Bowman and Storz, 2003]. Drag information from the trajectories of 75 ~ 80 inactive payloads and debris (calibration satellites) were analyzed for the global correction to the thermosphere using the HASDM processes covering the altitude range of 200 km ~ 800 km. Another neutral density data source is the CHAMP accelerometer data every 10 seconds for the 2001 through 2005 time period. The last data source is the GRACE accelerometer data every 5 seconds from 2002 through 2005.
New solar indices for temperature calculation

The $F_{10.7}$ index has been used as the proxy of thermospheric heating in various models. However, it has been shown that using only $F_{10.7A}$ index cannot fully capture the 11-year solar cycle variations, especially during the solar minimum period [Marcos et al., 2005 and 2006]. In order to capture the 11-year solar cycle variations in neutral density model, Bowman et al. [2008b] developed a new 11-year solar index $F_S$, weighting the $F_{10.7A}$ index of great majority being supplemented by the $S_{10.7A}$ index during the solar minimum period. Equation (3.27) shows the weighting scheme of the 11-year solar index. With the $F_S$ index, the new $T_c$ equation was developed for the JB2008 model from the analysis using numerous satellites from 1997 through 2007 expressed by Eq. (3.28), where the $\Delta$ prefix represents the difference between the daily and the 81-day average value of each solar index. The lag time of $F_{10.7}$ and $S_{10.7}$ indices was determined as 1 day, which is the same as that of the JB2006 model, and the lag time of the newly adopted $Y_{10.7}$ in the JB2008 model was chosen as 5 days for the least squares minimum. For the $M_{10.7}$ index, the different lag time of 2 days compared to 5 days in the JB2006 model was applied for the JB2008 model since the new $Y_{10.7}$ index absorbing X-rays and Lyman-\(\alpha\) emissions replaces $M_{10.7}$ index with capturing the low altitude longer absorption at around 80 km ~ 90 km [Bowman et al., 2008b],

$$F_S = F_{10.7A} \times W_T + S_{10.7A} \times (1 - W_T) \quad \text{where,} \quad W_T = \frac{\sqrt{(F_{10.7A}/240)}}{4} \quad (3.27)$$

$$T_c = 392.4 + 3.227F_S + 0.298\Delta F_{10.7} + 2.259\Delta S_{10.7} + 0.312\Delta M_{10.7} + 0.178\Delta Y_{10.7} \quad (3.28)$$
Modeling semiannual density variations

For representing the semiannual density variations, the JB2008 model basically used the same formulation of Eq. (3.24) as the JB2006 model but different expressions for \( F(z) \) and \( G(t) \). From the work by Bowman et al. [2008c], additional solar EUV indices were included in the JB2008 model so as to capture the semiannual density variations not modeled in the JB2006 model. Equation (3.29) is the new solar index for \( F(z) \) computed using \( S_{10.7A} \) and \( M_{10.7A} \) indices in addition to the previously used \( F_{10.7A} \) index, where the \( \bar{F}_j \), \( \bar{S}_j \), and \( \bar{M}_j \) indices are the July averages of \( F_{10.7A} \), \( S_{10.7A} \), and \( M_{10.7A} \) indices, respectively. Using the new solar indices for the semiannual density variations, the new \( F(z) \) equation for the JB2008 model was developed in the form of Eq. (3.30) and the coefficients of the equation are listed in Table 3-4,

\[
\bar{F}_{SMJ} = 1.00 \bar{F}_j - 0.70 \bar{S}_j - 0.04 \bar{M}_j
\]  

\[
F(z) = B_1 + B_2 \bar{F}_{SMJ} + B_3 z \bar{F}_{SMJ} + B_4 z^2 \bar{F}_{SMJ} + B_5 z \bar{F}_{SMJ}^2
\]

Table 3-4 \( B_i \) coefficients of Eq. (3.30) [Bowman et al., 2008b]

<table>
<thead>
<tr>
<th>( i )</th>
<th>( B_i )</th>
<th>( i )</th>
<th>( B_i )</th>
</tr>
</thead>
<tbody>
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<td>-2.78E-02</td>
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</tr>
<tr>
<td>3</td>
<td>2.78E-02</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Similarly to the \( F(z) \) analysis, the semiannual yearly periodic function, \( G(t) \), was determined to include long term EUV and MUV heating in the equation. A new solar index, \( \bar{F}_{SM} \), for the \( G(t) \) equation was developed in the similar way to the \( F(z) \) equation using \( S_{10.7A} \) and \( M_{10.7A} \) indices in addition to \( F_{10.7A} \) index. Equation (3.31) shows the new solar index for \( G(t) \) equation as
a linear function of the EUV and MUV solar indices. Using the new $F_{SM}$ solar index, the new $G(t)$ function for the JB2008 model was developed as Eq. (3.32), where $\omega$ is the same values as defined in the JB2006 model and the coefficient values are listed in Table 3-5, 

$$F_{SM} = 1.00 F_{10.7A} - 0.75 S_{10.7A} - 0.37 M_{10.7A}$$  \hspace{1cm} (3.31)$$

$$G(t) = C_1 + C_2 \sin(\omega) + C_3 \cos(\omega) + C_4 \sin(2\omega) + C_5 \cos(2\omega)$$  \hspace{1cm} (3.32)$$

$$+F_{SM}\{C_6 + C_7 \sin(\omega) + C_8 \cos(\omega) + C_9 \sin(2\omega) + C_{10} \cos(2\omega)\}$$

**Table 3-5** $C_i$ coefficients of Eq. (3.32) [Bowman et al., 2008b]

<table>
<thead>
<tr>
<th>$i$</th>
<th>$C_i$</th>
<th>$i$</th>
<th>$C_i$</th>
<th>$i$</th>
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<td>-1.79E-02</td>
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<td>-1.25E-03</td>
</tr>
</tbody>
</table>

**Modeling geomagnetic storm variations**

One of the significant improvements of the JB2008 model compared to the other model is that it adopted $D_s$ values for representing geomagnetic storm effects since the $D_s$ index is the indication of the strength of the storm time ring current in the magnetosphere and the time resolution of the $D_s$ index is 1 hour while that of $a_p/K_p$ index is 3 hour which are used in the other models. Bowman et al. [2008c] has adopted the equation determining the exospheric temperature responses as a function of the $D_s$ index developed by Wilson et al. [2006] and Burke et al. [2009]. Equation (3.33) shows the relation between the exospheric temperature responses and the $D_s$ index by Burke et al., where the subscript 1 is the one time step beyond the subscript 0 time.
After analyzing the storms from the GRACE accelerometer data, it was found that $\tau_1 = 6.5$ hours, $\tau_2 = 7.7$ hours, and the slope $S \approx 1.58$ were best fits for the equation. The exospheric temperature responses can be obtained by integrating the equation from storm commencement time until the end of storm time. In the JB2008 model, the magnetic storm was defined only when the minimum $D_{st}$ is less than -75 nT for practical reasons. Above -75 nT, the exospheric temperature responses by the geomagnetic effects is calculated using the 3-hourly $a_p$ index like the JB2006 model.

$$dT_{c,1} = \left(1 - \frac{1}{\tau_1}\right) dT_{c,0} + S \left\{D_{st,1} - \left(1 - \frac{1}{\tau_2}\right) D_{st,0}\right\} \tag{3.33}$$

The JB2008 model calculates $T_e$ corrections divided into several sub-stages of geomagnetic storm: the main phase, sub-storms during the main phase, recovery phase, and recovery slope change to the end of the storm. The sample progress of geomagnetic storm is in Figure 2-11 (b). During the main phase and sub-storm, it was decided to correct the slope $S$ from Eq. (3.33) for the optimization to the JB2008 model while $\tau_1$ and $\tau_2$ remained unchanged. The new quadratic function for $S$ instead of the constant value 1.58 was developed as Eq. (3.34). It is a function of minimum $D_{st}$ values only ($D_{st, \text{min}}$). As a result, the exospheric temperature corrections for the main phase of geomagnetic storms are calculated by Eq. (3.33) with the slope $S$ from Eq. (3.34),

$$S = -1.5050 \times 10^{-5} \left(D_{st, \text{min}}\right)^2 - 1.0604 \times 10^{-2} D_{st, \text{min}} - 3.20 \tag{3.34}$$

During the main phase of geomagnetic storms, there are usually sub-storms which the $D_{st}$ variations become positive for short time period. After analyzing the neutral densities from the accelerometer data and the HASDM processes, it was found that the neutral densities did not drop as expected even though $D_{st}$ values increased during the sub-storm periods in the main phase. It
was, therefore, determined to develop an additional equation for these time periods. Equation (3.35) shows the relation during the sub-storms, where \( S \) is calculated by Eq. (3.34) and the best factor \( S_{FAC} = 0.3 \). For the best fits to the thermospheric neutral density variations, it was decided to apply a small lag time for the exospheric temperature responses: no lag time for large storms \( (D_{st} < -350 \text{ nT}) \), 1 hour lag time for moderate storms \( (-350 \text{ nT} < D_{st} < -250 \text{ nT}) \), and 2 hours lag time for minor storms \( (-250 < D_{st}) \).

\[
dT_{c,1} = dT_{c,0} - S_{FAC} S \left( D_{st,1} - D_{st,0} \right)
\]

(3.35)

After analyzing recovery phases during geomagnetic storms and many trials for optimizing \( \tau_1, \tau_2, \) and \( S \), it was determined that \( \tau_1 = \infty, \tau_2 = 1, \) and \( S = 0.13 \) were best fits during the recovery phases. The resulting Eq. (3.36) is used for calculating the exospheric temperature responses for recovery phases until the recovery slope changes.

\[
dT_{c,1} = dT_{c,0} + 0.13 D_{st,1}
\]

(3.36)

The last phase of geomagnetic storms is from the recovery slope change to the end of the storm. The simple equation for this time period in calculating the exospheric temperature responses was developed as Eq. (3.37) with the constant slope \( S = -2.5 \),

\[
dT_{c,1} = dT_{c,0} + S \left( D_{st,1} - D_{st,0} \right)
\]

(3.37)

With the newly developed solar index \( (Y_{10.7}) \) for the heating in the lower thermosphere, the new semiannual density corrections, and the newly adopted geomagnetic storm disturbances \( (D_n) \) index, the accuracy of the JB2008 model in thermospheric neutral density estimation was significantly improved especially during the solar minimum periods and during the major geomagnetic storm periods.
NRLMSISE-00 model

Since the development of the first version of the MSIS-class models in 1977 (MSIS-1) [Hedin et al., 1977] based primarily on the ISR data under low to moderate solar activity conditions with the lower altitude boundary of 120 km, the model has been improved to the MSIS-83 [Hedin, 1983], MSIS-86 [Hedin, 1987], MSISE-90 [Hedin, 1991], and NRLMSISE-00 [Picone et al., 2002] by lowering the lower altitude boundary from 120 km to the surface of the Earth and by including drag measurements and satellite-borne accelerometer data among others in addition to the ISR data under low to high solar activity conditions. Therefore, the accuracy of the thermospheric neutral density estimation of the NRLMSISE-00 model improved during even high solar activity conditions. One of the remarkable changes in the NRLMSISE-00 model is that the model adopted a new component, named anomalous oxygen. It allows for O+ and hot atomic oxygen contributions to the total mass density near the exobase above 500 km. As a result, the NRLMSISE-00 model provide the users with two different mass densities: the total neutral mass density just like the previous models and the effective mass density which is the sum of the total neutral mass density and the anomalous oxygen contribution above 500 km.

Data reduction

The MSISE-90 model is primarily based on the ISR and mass spectrometer data with the additional data from solar UV occultation, pressure gauge, and falling sphere, among others, covering the period from 1965 through 1983 [Hedin, 1991]. The main drawback of the MSIS-class models (prior to the MSISE-90 model) was that they had not included any drag measurements and satellite-borne accelerometer data like the Jacchia-class models. That caused the MSIS-class models to have constraints for tracking and predicting satellites’ orbital behavior.
However, Picone et al. [2002] increased the accuracy of the MSIS-class models by expanding the data sets to include drag measurements and satellite-borne accelerometer data, among others in addition to the data sets of the MSISE-90 model.

The core of the NRLMSISE-00 model is still the ISR data directly influencing the temperature profile. More recent ISR data (~1998) were included into the model having the increased quality of the values with the help of the significantly improved processing methods for the data. In order to remove the constraints for tracking and predicting satellites’ orbital behavior, the drag measurements and accelerometer data used in the Jacchia models were included. One of the important data sets newly included in the NRLMSISE-00 model for the lower thermosphere was from the Millstone Hill data [Goncharenko and Salah, 1998]. The high quality lower thermospheric neutral temperature data from the Millstone Hill covering the altitude range of 100 km through 130 km were very important to model the temperature profile, not only in the lower thermosphere, but also near the mesopause. The last data sets included in the model were the O$_2$ number density from the SMM mission covering the altitude range of 140 km through 220 km [Aikin et al., 1993]. The SMM mission enabled direct measurement of O$_2$ above 150 km during high solar activity conditions and these data were used in the NRLMSISE-00 model for determining the dependence on the solar EUV radiation and magnetic activity conditions. With these newly adopted data sets from the several sources, the accuracy of the NRLMSISE-00 model had been improved for predicting orbital behavior and for estimating neutral density in the lower thermosphere.

**Model formulation**

For the temperature calculation, the NRLMSISE-00 model uses the Bates-Walker equations [Walker, 1965] for the upper thermosphere which is developed from Jacchia’s
temperature profile and an inverse polynomial [Hedin, 1991] for the lower thermosphere with several nodes. These temperature profiles allow to exactly integrating the hydrostatic equation so as the diffusive equilibrium density profiles to be determined for each important species. In the MSIS-class model, different turbopause heights were used according to species: 105 km for O₂, Ar, and O, 100 km for He, and 95 km for H. The turbopause height for N₂ used to be also set to 105 km in the previous MSIS-class models, but the height for N₂ in the NRLMSISE-00 model is not fixed to a specific value but varies with respect to latitude and day of year near 105 km.

The NRLMSISE-00 model uses the F₁₀.₇ value for the previous day and F₁₀.₇A indices only as the input for the solar irradiances. To represent the geomagnetic activity, it uses the daily averaged geomagnetic index (Aₚ) and/or 3-hourly geomagnetic index (aₚ). Since the effects of the solar irradiances and geomagnetic indices are small and not well established below around 80 km, the NRLMSISE-00 model sets these variables to 150, 150, and 4, respectively. The user can choose whether the model uses both Aₚ and aₚ indices or Aₚ index only. The Aₚ index only can be used during geomagnetic quiet time period. It is simple and fast to operate the model but inaccurate during geomagnetic storm time. Using both the Aₚ and aₚ indices is the most common and accurate method for representing the geomagnetic storm conditions. The aₚ indices up to 57 hours prior to the current time including the Aₚ index are used in this case. The exospheric temperature for the altitudes below 200 km is set to the global mean value of $T_\infty = 1027.32$ K.

**Computation of neutral temperature**

In order to calculate temperature and temperature gradient, the NRLMSISE-00 model uses the basic expansion formula, $G(L)$. The basic expansion formula is expressed by the sixth degree Legendre polynomials and is used to correct global averages of temperature and temperature gradient applying several variations: solar activity, annual/semiannual variations,
diurnal/semidiurnal/terdiurnal variations, magnetic activity, longitudinal variation, universal time, among others. The NRLMSISE-00 model incorporated the anomalous oxygen model profile into the basic expansion formula for above 500 km. Basically, the NRLMSISE-00 model uses the Bates-Walker temperature profile expressed by Eq. (3.38) for the upper thermosphere above 123.4 km ($h_\alpha$). The $\xi$, $\sigma$, $T_\infty$, and $T_{ib}$ are calculated by Eqs. (3.39) through (3.43), where the $T_{ib}$ is the temperature at 120 km ($h_{ib}$), $R_E$ is the Earth’s radius, $T'_{ib}$ is the temperature gradient at $h_{ib}$, and the $\bar{}$ (over bar) means the global average values,

$$T(h) = T_\infty - (T_\infty - T_{ib}) \exp\{-\sigma \xi(h,h_{ib})\} \quad (h \geq h_\alpha) \quad (3.38)$$

$$\xi(h,h_{ib}) = \frac{(h - h_{ib})(R_E + h_{ib})}{(R_E + h)} \quad (3.39)$$

$$\sigma = \frac{T'_{ib}}{(T_\infty - T_{ib})} \quad (3.40)$$

$$T'_{ib} = \bar{T}'_{ib} \{1 + G(L)\} \quad (3.41)$$

$$T_\infty = \bar{T}_\infty \{1 + G(L)\} \quad (3.42)$$

$$T_{ib} = \bar{T}_{ib} \{1 + G(L)\} \quad (3.43)$$

The altitude $h_\alpha$ is used as a fitting parameter and it was first set to 116.5 km in the MSIS-83 model. It has been changed to 117.2 km in the MSIS-86 model, 122.8 km in the MSISE-90 model, and 123.4 km in the NRLMSISE-00 model to provide a better overall fit to the data. Prior to the MSISE-90 model, the temperature profile below the altitude $h_\alpha$ was determined by a sixth-order polynomial. However, the temperature profile below the altitude $h_\alpha$ could not be expressed by polynomial in the MSISE-90 model, while extending the lower altitude boundary to the
Earth’s surface. Instead, a cubic spline as a function of inverse temperature with several chosen nodes for calculating the temperature below the altitude $h_a$ was used. The nodes used for calculating the inverse temperature are 123.4 km, 110 km, 100 km, 90 km, 72.5 km, 55 km, 45 km, 32.5 km, 20 km, 15 km, 10 km, and 0 km, which provides a reasonable representation of the temperature profile. For the temperature gradient profile, the altitudes of 72.5 km, 32.5 km, and 0 km were chosen as the nodes, which divides the lower atmosphere into three parts: the lower thermosphere/mesosphere (above 72.5 km), the lower mesosphere/upper stratosphere (32.5 km ~ 72.5 km), and the lower stratosphere/ troposphere (below 32.5 km).

The temperature and temperature gradient at the altitude $h_a$ are calculated by Eqs. (3.44) and (3.45). The neutral temperature is calculated with these values and the values for the chosen nodes for the lower thermosphere using cubic spline and the fixed global average exospheric temperature of $T_{\infty} = 1027.32$ K,

$$T_a = T_{\infty} - (T_{\infty} - T_{ib}) \exp\{-\sigma \xi (h_a, h_{ib})\}$$  \hspace{1cm} (3.44)

$$T'_a = (T_{\infty} - T_a) \sigma \left\{ \frac{(R_E + h_{ib})^2}{(R_E + h_a)} \right\}$$  \hspace{1cm} (3.45)

**Computation of neutral density**

With the temperature profiles, neutral density can be calculated by the hydrostatic equilibrium condition below the turbopause (mixing region) and by the diffusive equilibrium above the turbopause (diffusion region). The density profile of the NRLMSISE-00 model is a blend of the mixing and diffusive profiles multiplied by one or more factors for chemistry or dynamics flow effects in order to transit smoothly from the mixing to the diffusive equilibrium.
near the turbopause. As a result, the net density is expressed as a root of the sum of the diffusive
and the mixing densities by Eqs. (3.46) and (3.47),

\[ n(h, M) = \left\{ n_d(h, M)^A + n_m(h, M)^A \right\}^{1/A} \prod_{i=1}^{n} c_i(h) \]  \hspace{1cm} (3.46)

\[ A = \frac{M_t}{(M_o - M)} \]  \hspace{1cm} (3.47)

where,

\[ M_t = 28 \quad \bar{M}_o = 28.95 \]
\[ M = \text{molecular weight of gas species} \]
\[ n = \text{net number density} \]
\[ n_d = \text{number density of the diffusive profile} \]
\[ n_m = \text{number density of the mixing profile} \]
\[ c_i(h) = \text{lower thermospheric density multiplier (chemistry and dynamic flow effects)} \]

Neutral density in the mixing region is defined by the hydrostatic equilibrium and the
ideal gas law. In the diffusion region, the neutral density is calculated by Eqs. (3.48) through
(3.52). The number density at \( h_{lb} \) is calculated by applying the basic expansion formula for
correcting the global mean values like temperature and temperature gradient calculation.

\[ n_d(h, M) = n_{lb} \ D_B(h, M) \ \left\{ \frac{T(h_{lb})}{T(h)} \right\}^{(1+\alpha)} \]  \hspace{1cm} (3.48)

\[ D_B(h, M) = \ \left\{ \frac{T(h_{lb})}{T(h)} \right\}^{\gamma} \exp\{-\alpha \ \gamma \ \xi(h, h_{lb})\} \]  \hspace{1cm} (3.49)

\[ \gamma = \frac{M \ g_{lb}}{(\sigma \ R \ T_{\infty})} \]  \hspace{1cm} (3.50)
\[ g_{lb} = \frac{g_s}{\left(1 + \frac{h_{lb}}{R_E}\right)^2} \]  

(3.51)

\[ n_{lb} = \bar{n}_{lb} \exp\{G(L)\} \]  

(3.52)

where,

- \( g_s \) = gravitational acceleration at the Earth’s surface (9.80665 m/s\(^2\))
- \( g_{lb} \) = gravitational acceleration at \( h_{lb} \)
- \( n_{lb} \) = number density at \( h_{lb} \)
- \( R \) = universal gas constant
- \( \alpha \) = thermal diffusion coefficient (He and H: -0.38, Ar: 0.17, and others: 0.0)

One of the characteristics of the NRLMSISE-00 model is that it has adopted anomalous oxygen for \( O^+ \) and hot atomic oxygen contributions above 500 km. The number density of the anomalous oxygen, \( n_{O_a}(h) \), is calculated by Eq. (3.53) [Picone et al., 2002], where \( H(h, T) \) is the scale height with the mass of atomic oxygen, \( C = 76 \) km, \( h_a = 550 \) km, \( T_a = 4,000 \) K, \( h_{lb} = 120 \) km, and \( n_{O_a}(h_{lb}) = 6.0 \times 10^4 \) cm\(^{-3}\),

\[ n_{O_a}(h) = n_{O_a}(h_{lb}) \exp\left\{-\frac{\xi(h, h_{lb})}{H(h_{lb}, T_a)} + \frac{C}{H(h_a, T_a)}\left[1 - \exp\left(-\frac{h - h_a}{C}\right)\right]\right\} \]  

(3.53)

Figure 3-2 shows the computation flow of the neutral temperature and neutral density from the NRLMSISE-00 model [Kim, 2008]. Consequently, the accuracy of the NRLMSISE-00 model has been improved compared to the MSISE-90 model by incorporating new data from satellite-borne and ground-based measurements and by including the new component of the anomalous oxygen.
Figure 3-2  Computation flow of total mass density and neutral temperature on the NRLMSISE-00 model
TIE-GCM

The TIE-GCM solves continuity, momentum, and energy equations of the upper atmosphere ranging from around 97 km to around 600 km depending on solar activity with 0.5 scale vertical height resolution [Dickinson et al., 1981; Dickinson et al., 1984; Richmond et al., 1992]. It covers the latitude coordinates between 87.5° S and 87.5° N since the primitive equations of the dynamic meteorology have singularities at the poles. As a result, it was decided that the 87.5° latitude is the closest to the poles with numerical stability [personal communication with Ray Roble (NCAR), 2009]. It uses the pressure levels \( z = \ln \left( \frac{P_0}{P} \right) \) for representing vertical coordinates ranging from -7 to 7, where the reference pressure \( P_0 = 5 \times 10^{-4} \, \mu b \). While the empirical models use the \( F_{10.7} \) index of one day earlier, the TIE-GCM uses the interpolated \( F_{10.7} \) indices using four \( F_{10.7} \) values from two days earlier and two days later. It uses the \( K_p \) index instead of the \( a_p \) index for reflecting geomagnetic activity, unlike the empirical models. The TIE-GCM includes vertical propagation of semi-diurnal and diurnal tidal perturbations from the mesosphere to the thermosphere caused by solar radiation specified by the GSWM (Global Scale Wave Model) for the lower boundary [Hagan et al., 1999]. The ionospheric convection pattern is specified using the empirical model developed by Heelis et al. [1982]. The method for electrodynamic coupling between the thermosphere and the ionosphere was developed and implemented into the TIE-GCM [Richmond et al., 1992]. The TIE-GCM calculates the neutral temperature and winds, the height of the constant pressure level, the number densities of major neutral constituents (O, O\(_2\), and N\(_2\)), and the number densities of minor neutral constituents (N(\(^2\)D), N(\(^4\)S), and NO). It also calculates the electron density, ion and electron temperatures, and the concentrations of ions (O\(^+\), O\(_2^+\), NO\(^+\), N\(_2^+\), and N\(^+\)). Helium and hydrogen and their ions are not included in the model.
Boundary conditions

The lower altitude boundary of TIE-GGCM is set to the pressure level of -7, which is equivalent to approximately 97 km depending on the solar activity. It was assumed that the atomic oxygen number density is peak in the lower altitude boundary and uses a constant mass mixing ratio of 0.22 and 0.78 for \( \text{O}_2 \) and \( \text{N}_2 \), respectively. As background conditions, the neutral temperature and the neutral zonal/meridional winds were set to 181.0 K and 0 m/s, respectively, at an altitude of 96.372 km. The tidal perturbations can be added to these background conditions using the GSWM, which solves for migrating or nonmigrating waves with the steady-state assumptions from the Earth’s surface to 125 km [Hagan et al., 1999; Hagan and Forbes, 2002, 2003]. The forcing on the tidal perturbations is due to the thermospheric absorption of solar EUV radiation, the absorption of solar radiation in Schumann-Runge (S-R) bands and continuum in the mesopause region, the strato-mesospheric absorption of solar UV radiation, the tropospheric absorption of solar infrared (IR) radiation, and the tropospheric latent heating associated with deep convective activity (DCA), among others.

The upper altitude boundary of the TIE-GCM is the pressure level of +7. The diffusive equilibrium was assumed at the upper boundary for the thermosphere and the ionosphere. It was assumed that there were no sources of heat and momentum at sufficiently high levels of the atmosphere. As a result, the TIE-GCM uses the upper altitude boundary conditions that the vertical gradients of neutral temperature and winds equal 0 (\( \frac{\partial T}{\partial z} = \frac{\partial u}{\partial z} = \frac{\partial v}{\partial z} = \frac{\partial w}{\partial z} = 0 \)) [Dickenson et al., 1981].
Modeling in the thermosphere

- **Neutral temperature calculation**

With the lower and upper altitude boundary conditions, TIE-GCM calculates the neutral temperature, the zonal and meridional neutral winds, and the vertical neutral winds using the thermodynamic equation, the momentum equation, the continuity equation, and the hydrostatic equation. The neutral temperature is calculated by integrating neutral temperature gradient from the thermodynamic equation in the form of Eq. (3.54), where $\vec{V}$ is the horizontal velocity vector and $t$ is the time [Roble et al., 1988].

$$
\frac{\partial T}{\partial t} = \frac{g}{P_0 \overline{C_p}} \frac{\partial}{\partial z} \left( \frac{K_T}{H} \frac{\partial T}{\partial z} + K_E H^2 C_p \rho \left( \frac{g}{\overline{C_p}} + \frac{1}{H} \frac{\partial T}{\partial z} \right) \right) - \vec{V} \cdot \nabla T
$$

(3.54)

In the thermodynamic equation, the local time variation of the neutral temperature is calculated by the heating and cooling terms on the right hand side of the equation. The first term of the thermodynamic equation is the downward heat transfer caused by the vertical molecular thermal conduction. The second term is the adiabatic heating or cooling due to the downward eddy heat conduction. The third term is the horizontal heat advection due to the horizontal winds, and the fourth term is the vertical advection and adiabatic heating or cooling caused by the vertical winds. All the other heating and cooling sources such as the infrared cooling are included into the last term. The TIE-GCM adopted the results by Banks and Kockarts [1973] for calculating the specific heat coefficient ($C_p$) and the molecular thermal conductivity coefficient ($K_T$) in a gas mixture expressed by Eqs. (4.9) and (4.14), respectively, using only major neutral constituents $O_2$, $O$, and $N_2$. The TIE-GCM uses a constant eddy diffusion coefficient ($K_E = 5 \times$...
$10^{-6} \text{ s}^{-1}$ at the lower altitude boundary. From the specific value at the lower altitude boundary, the eddy diffusion coefficient is determined as the form of the global average $K_R = 5 \times 10^{-6} \exp (-7 - z)$ derived from a reasonable agreement to the observed global mean structure. For the upper boundary condition, thermal diffusive equilibrium condition is assumed.

- **Neutral winds calculation ($u$, $v$, and $w$)**

The horizontal neutral winds are calculated from the momentum equations. The zonal wind is obtained by integrating the eastward momentum equation and the meridional wind is by the northward momentum equation expressed by Eqs. (3.55) and (3.56), respectively [Dickinson et al., 1981]. The geopotential ($\Phi$) in the momentum equations is obtained by integrating the hydrostatic equation expressed by Eq. (3.57),

$$\frac{\partial u}{\partial t} = \frac{g e^z}{P_0} \frac{\partial}{\partial z} \left( \frac{\mu}{H} \frac{\partial u}{\partial z} \right) + \left( f + \frac{u}{r} \tan \phi \right) v + \lambda_{x\chi}(u_i - u) + \lambda_{x\gamma}(v_i - v) - \vec{V} \cdot \nabla u - w \frac{\partial u}{\partial z} - \frac{1}{r \cos \phi} \frac{\partial \Phi}{\partial \lambda}$$

$$\frac{\partial v}{\partial t} = \frac{g e^z}{P_0} \frac{\partial}{\partial z} \left( \frac{\mu}{H} \frac{\partial v}{\partial z} \right) + \left( f + \frac{u}{r} \tan \phi \right) u + \lambda_{y\gamma}(v_i - v) + \lambda_{y\chi}(u_i - u) - \vec{V} \cdot \nabla v - w \frac{\partial v}{\partial z} - \frac{1}{r} \frac{\partial \Phi}{\partial \phi}$$

$$\frac{\partial \Phi}{\partial z} = \frac{R(T_0 + T))}{m}$$

where,

$u = \text{eastward neutral velocity}$ \hspace{1cm} $u_i = \text{zonal ion drift velocity}$

$v = \text{northward neutral velocity}$ \hspace{1cm} $v_i = \text{meridional ion drift velocity}$

$f = \text{Coriolis parameter}$ \hspace{1cm} $\Phi = \text{geopotential}$

$r = \text{radial distance from the center of Earth}$ \hspace{1cm} $w = \text{vertical neutral wind}$
\( \lambda_{xx}, \lambda_{xy}, \lambda_{yy}, \lambda_{yx} = \text{ion drag tensor} \)

The time rate of change of horizontal neutral wind velocity is caused by the force due to the molecular and eddy viscosity, Coriolis force, ion drag force, the nonlinear horizontal momentum and advection force, vertical advection, and pressure gradient force.

The continuity equation of the thermospheric neutral gas follows the form of Eq. (3.58) and is solved for by obtaining the vertical velocity with the upper boundary condition of \( \partial w/\partial z = 0 \) and with the lower boundary conditions prescribed by the GSWM model [Dickinson et al., 1981].

\[
\frac{1}{r \cos \phi} \frac{\partial}{\partial \phi} (v \cos \phi) + \frac{1}{r \cos \phi} \frac{\partial u}{\partial \lambda} + e^z \frac{\partial}{\partial z} (e^{-z} w) = 0
\]  

(3.58)

- **Major species calculation (O\(_2\), O, and N\(_2\))**

The mass mixing ratio of the major neutral constituents of O\(_2\) and O are calculated using the continuity equation expressed by Eq. (3.59) [Dickinson et al., 1984]. The mass mixing ratio of N\(_2\) is obtained by \( \psi_{N2} = 1 - \psi_{O2} - \psi_O \),

\[
\frac{\partial}{\partial t} \psi = - \frac{e^z}{\tau} \frac{\partial}{\partial z} \left[ \frac{\bar{m}}{m_{N2}} \left( \frac{T_0}{T} \right)^{0.25} \alpha^{-1} L \psi \right] + S - R
\]

\[
+ e^z \frac{\partial}{\partial z} \left( K_E(z) e^{-z} \frac{\partial}{\partial z} \psi \right) - \nabla \cdot \nabla \psi - w \frac{\partial}{\partial z} \psi
\]

(3.59)

where,

\[ L_{ij} = \delta_{ij} \left\{ \frac{\partial}{\partial z} - \left( 1 - \frac{m_i}{\bar{m}} - \frac{1}{\bar{m}} \frac{\partial \bar{m}}{\partial z} \right) \right\} \]

\[ \psi = (\psi_{O2}, \psi_O) \text{, matrix for mass mixing ratio of O}_2 \text{ and O} \]

\[ \tau = \text{diffusion time scale (1.86 \times 10^3 \text{ s})} \quad T_0 = 273 \text{ K} \]

\[ \alpha = \text{matrix for diffusion coefficients} \quad \delta_{ij} = \text{Dirac delta function} \]
$S = \text{chemical production} \quad R = \text{chemical loss}$

Equation (3.59) shows that the time rate of change of the mass mixing ratio of major neutral species is caused by the molecular diffusion, the chemical production/loss, the eddy diffusion, and the horizontal/vertical advection. The term $S$ is for the production of atomic oxygen by the molecular oxygen photo-dissociation expressed by Eq. (2.1) which is the main source of atomic oxygen in the thermosphere in the Schumann-Runge bands and continuum. The term $R$ is for the loss of atomic oxygen and the loss rate is determined by the chemical recombination of $O$ into $O_2$. The upper boundary condition is the diffusive equilibrium condition which is defined as $\frac{\partial \psi}{\partial z} = 0$ in Eq. (3.59). The lower boundary conditions are the constant mass mixing ratio of 0.22 and 0.78 for $O_2$ and $N_2$, respectively, the maximum number density of atomic oxygen near 97 km, and the assumption of $\frac{\partial n(O)}{\partial z} = 0$.

- **Minor species calculation ($N(^2D)$, $N(^4S)$, and NO)**

The TIE-GCM included minor species transport of $N(^2D)$, $N(^4S)$, and NO. The transport equation expressed by Eq. (3.60) is used to solve for long-lived neutral constituents of $N(^4S)$ and NO,

$$
\frac{\partial}{\partial t} \psi = -e^{-z} \frac{\partial}{\partial z} A \{ \frac{\partial}{\partial z} - E \} \psi - \{ \nabla \cdot \nabla \psi + w \frac{\partial \psi}{\partial z} \} \\
+ e^z \frac{\partial}{\partial z} e^{-z} K_E(z) \left\{ \frac{\partial}{\partial z} + \frac{1}{m} \frac{\partial m}{\partial z} \right\} \psi + S - R
$$

(3.60)

where,

$$
E = \left\{ 1 - \frac{m_i}{m} - \frac{1}{m} \frac{\partial m}{\partial z} \right\} - \frac{1}{T} \frac{\partial T}{\partial z} + F \psi
$$

$A = \text{molecular diffusion coefficient} \quad \alpha = \text{thermal diffusion coefficient}$
$F_{\text{}} = \text{matrix operator for the frictional interaction between minor and major species}$

The time rate of change of the mass mixing ratio of minor neutral constituents is caused by the molecular diffusion, the horizontal/vertical advection, the eddy diffusion, and the chemical production/loss. The terms on the right hand side of $E$ are the gravitational force, the thermal diffusion, and the frictional interaction with major species. The diffusive equilibrium condition was assumed for both $N(S)$ and NO at the upper boundary. At the lower boundary, the photochemical equilibrium was assumed for $N(S)$ and a specified constant number density of $4 \times 10^6 \text{ cm}^{-3}$ was assumed for NO.

The number density of the short-lived minor neutral constituent of $N(D)$ is obtained from the assumption that $N(D)$ is in photochemical equilibrium throughout the thermosphere such that the production and loss rate of it is balanced. The production of $N(D)$ is mainly due to the solar EUV photo-dissociation of $N_2$ and ion/neutral chemistry.

**Modeling in the ionosphere**

- **Ion/electron number density calculation**

  The ion-neutral chemical reactions and reaction rates required for modeling the ionosphere are listed in Table 3-6 and Table 3-7, respectively [Roble et al., 1987]. The $O^+$ number density is calculated by the $O^+$ continuity equation considering both the electric field and the magnetic field expressed by Eq. (3.61) [Roble et al., 1988] and the number densities of the other ions ($O^+_2$, $N^+_2$, $NO^+$, and $N^+$) are calculated with the assumption of photochemical equilibrium with respect to the number density of atomic oxygen ions.
\[
\frac{\partial n(O^+)}{\partial t} = \eta(O^+) - L\ n(O^+) - \nabla \cdot [n(O^+) \ \vec{V}_i]
\]  
(3.61)

where,
\[
\vec{V}_i = \vec{V}_{i\parallel} + \vec{V}_{i\perp} \quad \vec{V}_{i\perp} = \frac{\vec{E} \times \vec{B}}{|B|}
\]
\[
\vec{V}_{i\parallel} = \left[ \vec{b} \cdot \frac{1}{v_{in}} \left\{ \vec{g} - \frac{1}{\rho_i} \nabla (P_i + P_e) \right\} + \vec{b} \cdot \vec{V} \right] \vec{b}
\]

\[n(O^+) = O^+ \text{ number density} \quad \eta(O^+) = O^+ \text{ production rate}\]

\[L = O^+ \text{ loss rate } \{L = k_1 n(O_2) + k_2 n(N_2)\}\]

\[\vec{V}_{i\perp} = \text{ ion velocity perpendicular to the magnetic field line by } \vec{E} \times \vec{B} \text{ drift}\]

\[\vec{V}_{i\parallel} = \text{ ion velocity parallel to the magnetic field line by ambipolar diffusion}\]

\[\vec{b} = \text{ unit vector along the magnetic field} \quad \vec{g} = \text{ gravitational acceleration}\]

\[v_{in} = O^+ \text{ ion-neutral collision frequency} \quad \rho_i = \text{ ion mass density}\]

\[P_i = \text{ ion pressure} \quad P_e = \text{ electron pressure}\]

\[\vec{V} = \text{ neutral wind vector} \quad \vec{E} = \text{ electric field vector}\]

\[\vec{B} = \text{ Earth’s magnetic field vector}\]

Equation (3.61) explains that the time rate of change of \(O^+\) number density is caused by the production/loss of \(O^+\) and by the transport of \(O^+\) induced from the \(\vec{E} \times \vec{B}\) drift and ambipolar diffusion. The production of atomic oxygen ions is mainly due to the ionization of oxygen atoms and due to the dissociative ionization of \(O_2\). The loss rate of \(O^+\) is determined by the charge exchange with \(O_2\) and by the chemical reactions with \(N_2\). The photochemical equilibrium of \(O^+\) is assumed at the lower altitude boundary and the transport of \(O^+\) from and to the plasmasphere is specified at the upper altitude boundary.

Once the \(O^+\) number density is determined by the continuity equation, the distribution of the other ions is solved with the assumption of photochemical equilibrium. A fourth-order polynomial was derived for solving the number density of electrons after equating production and loss terms in Table 3-6. The parameters for obtaining the electron number density are listed in
Table 3-6 and the chemical reaction rates are listed in Table 3-7. Using the electron number density obtained from the root of the fourth order polynomial, the number densities of $N_2^+$, $O_2^+$, NO$^+$, and $N^+$ can be calculated by Eqs. (3.62) through (3.65).

\[
n(N_2^+) = D / (E + \alpha_3 n_e) \tag{3.62}
\]

\[
n(O_2^+) = B / (C + \alpha_2 n_e) \tag{3.63}
\]

\[
n(NO^+) = \left\{ A + \frac{ED}{(E + \alpha_3 n_e)} + \frac{CB}{(C + \alpha_2 n_e)} \right\} / (\alpha_1 n_e) \tag{3.64}
\]

\[
n(N^+) = \eta(N^+) / \{(k_6 + k_7) n(O_2) + k_8 n(O)\} \tag{3.65}
\]

Table 3-6 Ion-neutral chemical reactions in the TIE-GCM [Roble et al., 1987]

<table>
<thead>
<tr>
<th>Reaction</th>
<th>Rate Constant</th>
<th>Product</th>
<th>Energy</th>
</tr>
</thead>
<tbody>
<tr>
<td>$O^+ (4S) + O_2 \rightarrow O_2^+ + O + 1.555 \text{ eV}$</td>
<td>$k_1$</td>
<td>$N^+ + O \rightarrow O^+ + N + 0.93 \text{ eV}$</td>
<td></td>
</tr>
<tr>
<td>$O^+ (4S) + N_2 \rightarrow NO^+ + N(4S)$</td>
<td>$k_2$</td>
<td>NO$^+ + e \rightarrow (20%) N(4S) + O + 2.75 \text{ eV}$</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>$N_2^+ + O \rightarrow NO^+ + N(2D) + 0.70 \text{ eV}$</td>
<td>$k_3$</td>
<td>$O_2^+ + e \rightarrow (15%) O(3P) + O(3P) + 6.95 \text{ eV}$</td>
<td></td>
</tr>
<tr>
<td>$O_2^+ + N(4S) \rightarrow NO^+ + O + 4.21 \text{ eV}$</td>
<td>$k_4$</td>
<td>$(85%) O(3P) + O(1D) + 4.98 \text{ eV}$</td>
<td></td>
</tr>
<tr>
<td>$O_2^+ + NO \rightarrow NO^+ + O_2 + 2.813 \text{ eV}$</td>
<td>$k_5$</td>
<td>N$^+_2 + e \rightarrow (10%) N(4S) + N(4S) + 5.82 \text{ eV}$</td>
<td></td>
</tr>
<tr>
<td>$N^+ + O_2 \rightarrow O_2^+ + N(4S) + 2.486 \text{ eV}$</td>
<td>$k_6$</td>
<td>$(90%) N(4S) + N(2D) + 3.44 \text{ eV}$</td>
<td></td>
</tr>
<tr>
<td>$N^+ + O_2 \rightarrow NO^+ + O + 6.699 \text{ eV}$</td>
<td>$k_7$</td>
<td>NO$^+ + h\nu_{Ly-\alpha} \rightarrow NO^+ + e$</td>
<td></td>
</tr>
</tbody>
</table>
Table 3-7 Ion-neutral chemical reaction rates in the TIE-GCM [Roble et al., 1987]

\[
k_1 = 2.82 \times 10^{-11} - 7.74 \times 10^{-12} \left( \frac{T_1}{300} \right) + 1.073 \times 10^{-12} \left( \frac{T_1}{300} \right)^2 -5.17 \times 10^{-14} \left( \frac{T_1}{300} \right)^3 + 9.65 \times 10^{-16} \left( \frac{T_1}{300} \right)^4 \quad 300 \leq T_2 \leq 6000 \text{ K}
\]

\[
k_2 = \begin{cases} 
1.533 \times 10^{-12} - 5.92 \times 10^{-13} \left( \frac{T_2}{300} \right) + 8.6 \times 10^{-14} \left( \frac{T_2}{300} \right)^2 & 300 < T_2 \leq 1700 \text{ K} \\
2.73 \times 10^{-12} - 1.155 \times 10^{-12} \left( \frac{T_2}{300} \right) + 1.483 \times 10^{-13} \left( \frac{T_2}{300} \right)^2 & 1700 < T_2 \leq 6000 \text{ K}
\end{cases}
\]

\[
k_3 = \begin{cases} 
1.4 \times 10^{-10} \left( \frac{300}{T_R} \right)^{0.44} & T_R < 1500 \text{ K} \\
5.2 \times 10^{-11} \left( \frac{T_R}{300} \right)^{0.2} & T_R \geq 1500 \text{ K}
\end{cases}
\]

\[
k_4 = 2.0 \times 10^{-10} \\
k_5 = 4.4 \times 10^{-10} \\
k_6 = 4.0 \times 10^{-10}
\]

\[
k_7 = 2.0 \times 10^{-10} \\
k_8 = 1.0 \times 10^{-12}
\]

\[
\alpha_1 = 4.2 \times 10^{-7} \left( \frac{300}{T_e} \right)^{0.85}
\]

\[
\alpha_2 = \begin{cases} 
1.6 \times 10^{-7} \left( \frac{300}{T_e} \right)^{0.55} & T_e \geq 1200 \text{ K} \\
2.7 \times 10^{-7} \left( \frac{300}{T_e} \right)^{0.7} & T_e < 1200 \text{ K}
\end{cases}
\]

\[
\alpha_3 = 1.8 \times 10^{-7} \left( \frac{300}{T_e} \right)^{0.39}
\]

where \( T_1 = 0.667 \ T_i + 0.333 \ T_n, \ T_2 = 0.6363 \ T_i + 0.3637 \ T_n, \ T_R = (T_i + T_n)/2 \)

\( T_i = \text{ion temperature, } T_e = \text{electron temperature, and } T_n = \text{neutral temperature} \)
Table 3-8 Parameters for calculating \( n_e, O_2^+, N_2^+, NO^+, \) and \( N^+ \) [Roble and Ridley, 1987]

\[
a_4 n_e^4 + a_3 n_e^3 + a_2 n_e^2 + a_1 n_e + a_0 = 0
\]

where,

\[
a_4 = \alpha_1 \alpha_2 \alpha_3
\]

\[
a_3 = \alpha_1 (\alpha_2 E + \alpha_3 C) - \alpha_1 \alpha_2 \alpha_3 (F + G)
\]

\[
a_2 = \alpha_1 EC - \alpha_1 (\alpha_2 E + \alpha_3 C)(F + G) - \alpha_1 \alpha_2 D - \alpha_1 \alpha_3 B - \alpha_2 \alpha_3 A
\]

\[
a_1 = -\alpha_1 \{EC(F + G) + DC + BE\} - \alpha_2 E(A + D) - \alpha_3 C(A + B)
\]

\[
a_0 = -EC(A + B + D)
\]

\[
A = k_2 n(O^+) n(N_2) + k_7 n(N^+) n(O_2) + \beta_9 n(NO)
\]

\[
B = \eta(O_2^+) + k_{11} n(O^+) n(O_2) + k_{16} n(N^+) n(O_2)
\]

\[
C = k_4 n\{N(4S)\} + k_9 n(NO)
\]

\[
D = \eta(N_2^+) \quad E = k_3 n(O) \quad F = n(O^+) \quad G = n(N^+)
\]

\[
n_e = \text{number density of electrons} \{n_e = n(O^+) + n(O_2^+) + n(N_2^+) + n(N^+) + n(NO^+)\}
\]

\[
\eta = \text{chemical production of ion}
\]
• **Ion/electron temperature calculation**

In order to calculate the electron temperature, the TIE-GCM adopted the electron energy conservation equation developed by Schunk and Nagy [1978]. The electron energy conservation equation can be written as Eq. (3.66) neglecting the chemical and viscous heating of the electrons, where \( \bar{u}_e \) is the electron bulk velocity, \( \bar{q}_e \) is the electron heat flow vector, and \( Q_e/L_e \) are the local electron heating/cooling rates. The assumption for the lower boundary is that the electron temperature is the same as the neutral temperature.

\[
\frac{3}{2} n_e k \frac{\partial T_e}{\partial t} = -n_e k T_e \nabla \cdot \bar{u}_e - \frac{3}{2} n_e k \bar{u}_e \cdot \nabla T_e - \nabla \cdot \bar{q}_e + \sum Q_e - \sum L_e 
\]  

(3.66)

The ion temperature in the TIE-GCM is calculated by Eq. (3.67), where \( L_{e,i} \) is the cooling rate due to the energy transfer between the electrons and ions, \( L_{i,n} \) is the cooling rate between the ions and the major neutral constituents, and \( Q_J \) is the Joule heating. The assumption for the ion temperature calculation is that the quasi-steady state of energy exchange among ions, electrons, and major neutral constituents.

\[
L_{e,i}(T_e - T_i) + \rho Q_J = L_{i,n}(T_i - T_n) 
\]

(3.67)

The TIE-GCM uses the same spherical coordinates and solves on the same geographic grid for the thermosphere and the ionosphere, allowing to couple between the neutral and the ionic atmospheres. As a result, the TIE-GCM is useful to produce the basic structure/behavior of the thermosphere/ionosphere, and useful to examine the interactions among thermodynamics, plasma-dynamics, and electrodynamics. Chapter 4 compares the results from the empirical/numerical thermospheric neutral density models to the accelerometer data from the CHAMP and GRACE satellites.
Chapter 4

DATA AND ANALYSIS OF MODEL SIMULATIONS

This chapter describes the CHAMP and GRACE satellites, and discusses the derivation method of thermospheric neutral density from the CHAMP and GRACE accelerometer data. With those thermospheric neutral density data, the accuracies of both empirical and numerical density models are compared at 400 km from near solar maximum period (2002) to near solar minimum period (2007). Specific heat, molecular viscosity, and molecular thermal conductivity are simulated with and without helium using the equations incorporated into the TIE-GCM, which are used in the thermodynamic equations affecting thermospheric neutral temperature and, as a result, affecting thermospheric neutral density. Finally, methods for incorporating helium data from the NRLMSISE-00 model into the TIE-GCM are suggested in order to improve the estimation of the neutral density of the TIE-GCM. The helium data were used to calculate the specific heat, molecular viscosity, and molecular thermal conductivity in the TIE-GCM, which are used in the thermodynamic equation for neutral temperature calculation in the thermosphere. Several increments of helium data from the NRLMSISE-00 model were simulated for finding optimal values and for comparing the sensitivity.

**Thermospheric neutral density data and model comparison**

Since the CHAMP and GRACE satellites were equipped with GPS receivers and very accurate accelerometers, nearly continuous thermospheric neutral density data have been available from nearly pole to pole using the sub-decimeter accuracy of the satellite tracking data. Those neutral density data help to better understand the responses of the upper atmosphere to the
variations of solar-terrestrial activity. Also, the accuracy of thermospheric neutral density models can be compared with the neutral density data derived from the accelerometer data.

**CHAMP and GRACE satellites**

Recently, with the help of the GPS satellite-to-satellite tracking measurements, the CHAMP satellite carrying the BlackJack GPS receiver provided almost continuous satellite tracking data with the sub-decimeter accuracy for radial, cross-track, and along-track components extending from pole to pole [Kuang et al., 2001]. The CHAMP satellite was launched on July 15, 2000 into an almost circular (~ 0.003 eccentricity), near polar (87.3° inclination) orbit with an initial altitude of 454 km. The mission was managed by the GeoForschungsZentrum (GFZ) in Potsdam, Germany. The nominal mission duration of the CHAMP satellite was five years, which covered the solar activity conditions from maximum (2001) to minimum (2005) of solar cycle 23. Thanks to the extraordinarily prolonged low solar activity condition during the time of solar cycle 23/24 minimum and four altitude maneuvers (twice in 2002, third time in 2006, and last time in 2009), the lifetime of the CHAMP satellite was substantially extended more than the five year beyond the nominal mission duration. It reentered the Earth’s atmosphere on September 20, 2010.

The primary mission objectives of the CHAMP satellite were mapping the magnetic and gravity fields of the Earth, and monitoring the ionosphere and troposphere. The monitoring of thermospheric neutral density was one of the secondary mission objectives. Thermospheric neutral density can be derived from the very accurate STAR (Spatial Tri-axial Accelerometer for Research) accelerometer data along with knowledge of the characteristics (area and mass) of the CHAMP satellite. Figure 4-1 shows the layout of the CHAMP satellite and the reference frame of the STAR accelerometer, which define the x-axis for radial, y-axis for along-track, and z-axis
as normal to the $x$-$y$ plane. The derivation method of thermospheric neutral density from the accelerometer data is described in next section in detail.

The GRACE satellites were launched on March 17, 2002 into an almost circular (~ 0.001 eccentricity), near polar orbit (89° inclination) with an initial altitude of 485 km. The GRACE mission succeeds the CHAMP mission and provides an even higher accuracy. In order to attain
the higher accuracy, the GRACE mission consisted of two identical satellites (GRACE-A and -B) separated by around 220 km along-track. They equipped the much more accurate SuperSTAR accelerometer rather than the STAR accelerometer. The SuperSTAR accelerometer provides the accuracy of approximately $10^{-10}$ m/s$^2$, while the accuracy of the STAR accelerometer is approximately $10^{-4}$ m/s$^2$ [Bruinsma et al., 2004; Kang et al., 2006]. The derivation method of thermospheric neutral density from the GRACE accelerometer observations is very similar to that from the CHAMP accelerometer data.

**Thermospheric neutral density derivation from the accelerometer data**

Currently, two sources of calculating density data from CHAMP accelerometer data are available: one from Sutton et al. of the University of Colorado and the other from Bruinsma et al. of the French Space Agency, CNES (Centre National d'Etudes Spatiales), who is studying DTM modeling. Bruinsma et al. [2004] retrieved total atmospheric neutral densities from the STAR accelerometer data with an absolute uncertainty of 10-15% for up to moderate geomagnetic activity conditions ($a_p = 15$). However, they used the DTM-2000 atmospheric density model for calculating observed total neutral density by scaling the density predicted with the model. On the other hand, Sutton et al. [2007] developed a method for deducing neutral densities and winds in the thermosphere using the STAR accelerometer data from the CHAMP satellite without using a scaling process from any density models. In this dissertation, the neutral density data by Sutton et al. are used.

The acceleration caused by the drag and lift accelerations can be expressed by Eqs. (4.1) and (4.2), respectively, where $A$ is the total plate area,

---

*http://sisko.colorado.edu/sutton/data.html, [retrieved 15 October 2010]*
The CHAMP satellite consisted of 13 flat plates. Table 4-1 shows areas, coefficients of diffusive and specular reflectivity in visible/IR emissions, and normalized vectors normal to the area for these 13 flat surfaces [Lühr et al., 2002]. Sutton et al. developed the relation between the total acceleration caused by drag and lift forces and the sum of those from each plate. Therefore, Eq. (4.3) for calculating neutral density from the relation was obtained. When estimating neutral density, the satellite velocity, $\vec{v} = \vec{v}_c - \vec{w}_e$, was approximated to $\vec{v}_c$ with the assumption of negligible $\vec{w}_e$ compared to the velocity of the satellite,

$$\rho = \frac{-2 m (\vec{a}_{dt} \cdot \hat{y})}{\sum_{i=1}^{13} \left( A_i C_{Di} (\vec{v}_c \cdot \hat{n}_i) \vec{v}_c + A_i C_{Ui} (\cos \theta_i / \sin \theta_i) (\vec{v}_c \times \hat{n}_i) \times \vec{v}_c \right) \cdot \hat{y}} \quad (4.3)$$

They deduced total acceleration due to the drag and lift accelerations, $\vec{a}_{dt}$, from the STAR accelerometer measurements by excluding the acceleration due to the solar radiation pressure and Earth radiation pressure. The uncertainty in neutral density estimation with this method is generally less than 15%. The details of the method and error analysis are described in Sutton et al. [2007].

Comparison of neutral density models

In order to compare the accuracy of thermospheric neutral density models, three cases were investigated: (1) daily averaged density variations, and (2) sun/geomagnetic quiet and (3) storm time density variations at 400 km along the track of the CHAMP satellite. Figure 4-2
represents daily averaged neutral densities (approximately 1,800 data points a day) from the CHAMP accelerometer, JB2006, JB2008, NRLMSISE-00, and TIE-GCM models at 400 km during the year of 2002 (near solar maximum) through 2007 (near solar minimum). Figure 4-3 shows percent difference from the CHAMP accelerometer to computer models. As seen from the figures, all of the density models have good morphology to the variations of densities from the CHAMP accelerometer. However, the TIE-GCM has relatively larger uncertainties in the RMS (root mean square): 47.1% for the TIE-GCM compared to 15.7%, 10.3%, and 20.3% for the JB2006, JB2008, and NRLMSISE-00 models, respectively. Figure 4-4 shows the curve-fitted daily averaged neutral density ratios of the TIE-GCM to the CHAMP each year using the Fourier analysis. The ratios increase with the decline in the solar activity and the density estimation errors also gradually increase approaching to the solar minimum time. In addition, it shows similar patterns of density estimation errors following the semiannual density variation. Qian et al. [2009] suggested decreasing the semiannual density errors by using the modified eddy

Table 4-1  Surface properties of CHAMP satellite [Lühr et al., 2002]

<table>
<thead>
<tr>
<th>Panel</th>
<th>$A_i$ (m$^2$)</th>
<th>$c_{rs}$ (visible)</th>
<th>$c_{rd}$ (visible)</th>
<th>$c_{rs}$ (IR)</th>
<th>$c_{rd}$ (IR)</th>
<th>Normalized vector normal to plate</th>
</tr>
</thead>
<tbody>
<tr>
<td>top</td>
<td>1.2920</td>
<td>0.05</td>
<td>0.30</td>
<td>0.03</td>
<td>0.16</td>
<td>1.0000 0.0000 0.0000</td>
</tr>
<tr>
<td>bottom</td>
<td>3.6239</td>
<td>0.68</td>
<td>0.20</td>
<td>0.19</td>
<td>0.06</td>
<td>-1.0000 0.0000 0.0000</td>
</tr>
<tr>
<td>left</td>
<td>3.1593</td>
<td>0.05</td>
<td>0.30</td>
<td>0.03</td>
<td>0.16</td>
<td>0.7070 0.0000 0.7070</td>
</tr>
<tr>
<td>left (rear)</td>
<td>0.3020</td>
<td>0.40</td>
<td>0.26</td>
<td>0.23</td>
<td>0.15</td>
<td>-0.7070 0.0000 -0.7070</td>
</tr>
<tr>
<td>right</td>
<td>3.1593</td>
<td>0.05</td>
<td>0.30</td>
<td>0.03</td>
<td>0.16</td>
<td>0.7070 0.0000 -0.7070</td>
</tr>
<tr>
<td>right (rear)</td>
<td>0.3020</td>
<td>0.40</td>
<td>0.26</td>
<td>0.23</td>
<td>0.15</td>
<td>-0.7070 0.0000 0.7070</td>
</tr>
<tr>
<td>aft</td>
<td>0.4902</td>
<td>0.40</td>
<td>0.26</td>
<td>0.23</td>
<td>0.15</td>
<td>-0.3420 -0.9397 0.0000</td>
</tr>
<tr>
<td>front</td>
<td>1.2199</td>
<td>0.20</td>
<td>0.40</td>
<td>0.26</td>
<td>0.51</td>
<td>-0.9397 0.3420 0.0000</td>
</tr>
<tr>
<td>boom (top)</td>
<td>0.9300</td>
<td>0.40</td>
<td>0.26</td>
<td>0.23</td>
<td>0.15</td>
<td>1.0000 0.0000 0.0000</td>
</tr>
<tr>
<td>boom (bottom)</td>
<td>0.9300</td>
<td>0.40</td>
<td>0.26</td>
<td>0.23</td>
<td>0.15</td>
<td>-1.0000 0.0000 0.0000</td>
</tr>
<tr>
<td>boom (left)</td>
<td>0.9300</td>
<td>0.40</td>
<td>0.26</td>
<td>0.23</td>
<td>0.15</td>
<td>0.0000 0.0000 -1.0000</td>
</tr>
<tr>
<td>boom (right)</td>
<td>0.9300</td>
<td>0.40</td>
<td>0.26</td>
<td>0.23</td>
<td>0.15</td>
<td>0.0000 0.0000 1.0000</td>
</tr>
<tr>
<td>boom (front)</td>
<td>0.0529</td>
<td>0.20</td>
<td>0.40</td>
<td>0.26</td>
<td>0.51</td>
<td>0.0000 1.0000 0.0000</td>
</tr>
</tbody>
</table>
diffusion coefficients at the lower boundary with Fourier series of four harmonics per year. However, the semiannual density variation is beyond the scope of this dissertation.

For the numerical comparison of the accuracies of thermospheric neutral density models, percent differences \( D_{per,i} \) of computer models from the CHAMP accelerometer were used, calculated by Eq. (4.4). Plotting the percent differences provides each model’s deviation tendency in each day of year. On the other hand, the root mean square (RMS) provides the overall percent deviations of computer models and is calculated by Eq. (4.5).

\[
D_{per,i} \, (\%) = \left( \frac{\rho_{\text{model},i} - \rho_{\text{CHAMP},i}}{\rho_{\text{CHAMP},i}} \right) \times 100 \quad (4.4)
\]

\[
\text{RMS} = \sqrt{\frac{\sum_{i=1}^{N} (D_{per,i})^2}{N}} \quad (4.5)
\]

In order to investigate thermospheric neutral density response to a severe geomagnetic storm, densities at 400 km from accelerometer measurements from the CHAMP satellites were used on days 323-325 in 2003. The peak \( F_{10.7} \) and \( F_{10.7A} \) were 177 and 138, respectively. The peak \( A_p \) was 150 and the peak \( a_p \) was 300 at 15:00 and 18:00 UT on day 324. Figure 4-5 shows neutral densities from the CHAMP accelerometer and computer models at 400 km, and Figure 4-6 shows percent difference from the CHAMP accelerometer to computer models during the geomagnetic storm time. The periodic oscillations of density in Figure 4-5 are due to the orbital period of the CHAMP satellite of around 90 minutes since the densities are derived along the track of the CHAMP satellite. The figures show substantial increases in densities during this isolated and severe geomagnetic storm time. Globally, neutral density increased around 300%~800% in this time period, with about a four-hour delay at the equator and with relatively little time delay at high latitudes [Bruinsma et al., 2006]. While the JB2006 model uses only \( a_p \) index of 6.7 hours earlier, the NRLMSISE-00 model uses history of \( a_p \) indices up to 57 hours.
prior to current time for geomagnetic storm. The JB2008 model developed the geomagnetic storm modeling to determine exospheric temperature responses using the $D_s$ index which is determined hourly measurements of the magnetic field. The $D_s$ indices are used for geomagnetic storm time and $a_p$ indices are used for non-storm period in the JB2008 model. The JB2008 model has the lowest percent difference from the CHAMP accelerometer of 21.1% compared to 27.9%, 44.0%, and 49.7% of the TIE-GCM, JB2006, and NRLMSISE-00 models, respectively, during this particular storm period.

Figure 4-7 shows neutral densities from the CHAMP accelerometer and computer models at 400 km, and Figure 4-8 shows percent difference from the CHAMP accelerometer to computer models during a geomagnetic quiet time on day 188 in 2003 near solar maximum. The $F_{10.7}$ and $F_{10.7A}$ were 133 and 128, respectively. The daily $A_p$ was 11 and the peak 3-hourly $a_p$ was 22 at 06:00 UT. The JB2008 model also has the lowest percent difference from the CHAMP accelerometer of 14.2% compared to 38.1%, 22.0%, and 18.0% of the TIE-GCM, JB2006, and NRLMSISE-00 models, respectively, during this particular geomagnetic quiet time.

Overall, the newest Jacchia-class model, JB2008 model, has the lowest uncertainties in density estimation both long-term and short-term density comparison. The study by Bowman et al. [2008b] supports these results. Figure 4-9 shows the comparisons of empirical models with the standard deviations of each model based on the values from the averaged percent density differences between computer models and both the CHAMP and GRACE accelerometer data as a function of geomagnetic storm ranges from minor to major storms represented by the $a_p$ values [Bowman et al., 2008b]. It shows that the JB2008 model has the lowest standard deviations through all geomagnetic storm ranges, especially during major storms.
Figure 4-2  Daily averaged mass density comparisons from the CHAMP accelerometer to the computer models (JB2006, JB2008, NRLMSISE-00, and TIE-GCM) at 400 km along the track of CHAMP satellite from 2002 to 2007.
Figure 4-3 Percent differences of computer models from the CHAMP accelerometer from 2002 to 2007 shown in Figure 4-2.
Figure 4.4 Curve fitted density ratios of the TIE-GCM to the CHAMP accelerometer from 2002 (near solar maximum) to 2007 (near solar minimum) using the Fourier analysis. The ratios increase with the decline in the solar activity.
Figure 4-5  Neutral density comparisons from the CHAMP accelerometer to the computer models at 400 km during severe geomagnetic storm time on days 323 ~ 325 in 2003.
Figure 4-6  Percent differences of computer models from the CHAMP accelerometer on days 323~325 in 2003 shown in Figure 4-5.
Figure 4-7 Neutral density comparisons from the CHAMP accelerometer to the computer models at 400 km during geomagnetic quiet time on day 188 in 2003.
Figure 4-8  Percent differences of computer models from the CHAMP accelerometer on day 188 in 2003 shown in Figure 4-7.
Figure 4-9  Comparisons of thermospheric empirical neutral density models as a function of geomagnetic storm ranges (ap values). Standard deviations are based on averaged percent density differences between empirical models and both CHAMP and GRACE accelerometer data [Bowman et al., 2008b].
Analysis of helium effects in the upper thermosphere

Although helium is a minor neutral constituent in the lower Earth’s atmosphere, it plays an important role as the one of the major neutral constituent at the upper thermosphere. It significantly affects the specific heat and molecular thermal conductivity with subtle effects on the molecular viscosity in a gas mixture and, subsequently, affects the thermospheric neutral temperature and density. The effects of helium becomes more significant during low solar activity condition rather than during high solar activity.

Physics with mechanical and thermal properties of the air in the TIE-GCM

In order to obtain neutral temperature in the TIE-GCM, the thermodynamic equation expressed by Eq. (4.6) is numerically solved [Roble et al., 1988].

\[
\frac{\partial T}{\partial t} = \frac{g e^2}{P_0 C_p} \frac{\partial}{\partial z} \left( K_T \frac{\partial T}{\partial z} + K_r e^2 C_p \rho \left( \frac{g}{C_p} + \frac{1}{H} \frac{\partial T}{\partial z} \right) \right) - \vec{V} \cdot \nabla T
\]

\[= -w \left( \frac{\partial T}{\partial z} + \frac{RT}{C_p \overline{m}} \right) + \frac{(Q - L)}{C_p} \]

(4.6)

The time derivative of the local neutral temperature on left hand side is calculated by the heating and cooling terms on right hand side in Eq. (4.6). Details about right hand side terms are described in Chapter 3. The molecular thermal conductivity \((K_r)\) and the specific heat per unit mass \((C_p)\) are used in the first, third, and last terms of the thermodynamic equation for calculating the neutral temperature at each pressure level and time step in the TIE-GCM. This dissertation focuses on the specific heat, molecular viscosity, and molecular thermal conductivity that are incorporated into the TIE-GCM. In a gas mixture, the TIE-GCM uses only \(O_2\), \(N_2\), and \(O\) for
calculating these variables. The model calculates the specific heat, molecular viscosity, and thermal conductivity using the equations from Banks and Kockarts [1973].

- **Specific heat**

The equations for the specific heat (the amount of heat per unit mass required to raise the temperature by one degree) for individual gas constituent at constant volume and at constant pressure can be expressed by Eqs. (4.7) and (4.8), respectively, where \( k \) is the Boltzmann constant and \( N \) is the number of degrees of freedom of the particles. \( N = 3 \) for mono-atomic gases and \( N > 3 \) for the other gases. For the major neutral constituents \( \text{O}_2 \) and \( \text{N}_2 \) in the TIE-GCM, \( N = 5 \).

\[
C_{vi} = \frac{Nk}{2m_i} \quad \text{(4.7)}
\]

\[
C_{pi} = \left( \frac{k}{m_i} \right) \left[ 1 + \frac{N}{2} \right] \quad \text{(4.8)}
\]

In a gas mixture, the specific heat can be combined with the sum of the contributions of individual species’ specific heats using the mean molecular weight instead of the molecular weight of individual species and the partial pressure of individual species. The TIE-GCM calculates the specific heat of the air at constant pressure using three major neutral constituents’ contributions. The specific heat in the TIE-GCM is determined as Eq. (4.9), where \( R \) is the universal gas constant and \( P \) is the partial pressure of individual species.

\[
C_p = \frac{1}{2} R \left[ \frac{7}{32} P(O_2) + \frac{7}{28} P(N_2) + \frac{5}{16} P(O) \right] \quad \text{(4.9)}
\]
• **Molecular viscosity**

With many efforts to develop the theoretical and semi-empirical formula for the molecular viscosity (quantity describing a fluid’s resistance to flow), the accurate formula of a mixture of \( n \) constituents has been obtained as Eq. (4.10), where subscripts \( i \) and \( j \) represent \( i^{th} \) and \( j^{th} \) neutral constituents.

\[
\mu = \sum_i \frac{1}{n_i} \sum_j n_j \varphi_{ij}
\]  

(4.10)

\[
\varphi_{ij} = \frac{1 + (\mu_i/\mu_j)^{1/2}(m_j/m_i)^{1/4}}{2 \sqrt{2} \left[ 1 + (m_i/m_j)^{1/2} \right]^{1/2}}
\]  

(4.11)

A simple expression for \( \varphi_{ij} \) in Eq. (4.10) can be defined as Eq. (4.11). \( \varphi_{ij} \) can be further simplified to the value of \( \varphi_{ij} = 1 \) if \( \mu_i/\mu_j \approx 1 \) and \( m_i/m_j \approx 1 \) with the assumption of the same order of magnitude on each component. This simplest formula for the molecular viscosity in a gas mixture is used in the TIE-GCM.

Banks and Kockarts developed the useful formula of the molecular viscosity of individual neutral constituent expressed like \( \mu_i = A_i T^s \) for the applications in aeronomy. After analyzing numerous measurements of major neutral constituents, they determined the \( A_i \) and \( s \) for each species between 200 K and 2,000 K with the uncertainty less than 5%.

The final formula for calculating molecular viscosity in a gas mixture with major neutral constituents is expressed as Eq. (4.12) and the TIE-GCM adopted this result.

\[
\mu = [4.03 \ P(O_2) + 3.42 \ P(N_2) + 3.9 \ P(O)] \times 10^{-6} \ T^{0.69}
\]  

(4.12)
• **Thermal conductivity**

The thermal conductivity coefficient (property of a material reflecting its ability to conduct heat) of an individual species in a mixed gas is closely related to the molecular viscosity, which can be expressed by Eq. (4.13).

\[
K_{T_i} = f \mu_i C_{vi}
\]  

(4.13)

where, \(f\) is a number depending on the interaction potential. \(f\) can be simply approximated to 2.5 for mono-atomic species. The thermal conductivity coefficients for the diatomic species \(N_2\) and \(O_2\) have been established using different expressions [Chapman and Cowling, 1970; Hirschfelder et al., 1964]. With these values, Eq. (4.13) can be approximated and simplified to Eq. (4.14) for a gas mixture with a neutral temperature between 200 K and 2,000 K, and the TIE-GCM uses that equation for calculating the thermal conductivity,

\[
K_T = [56 P(O_2) + 56 P(N_2) + 75.9 P(O)] T^{0.69}
\]  

(4.14)

For reference, the GITM uses the different form of the thermal conductivity coefficient expressed by Eq. (4.15), which ignores a change in the neutral composition of the thermosphere,

\[
K_T = 56 T^{3/4}
\]  

(4.15)

**Variation of helium in the upper thermosphere**

The TIE-GCM does not include any physics of helium at all altitude ranges, while the other thermospheric empirical models do include helium for calculating the neutral density.
However, helium should be taken into account at the upper thermosphere where helium gradually takes an important role as a major neutral constituent. Differential heating between the summer hemisphere and the winter hemisphere causes the gradient in temperature and pressure in the Earth’s thermosphere. As a result, large scale thermospheric circulation systems are developed from the summer hemisphere of higher pressure to the winter hemisphere of lower pressure in the upper thermosphere and vice versa in the lower thermosphere. The circulation system also carries and redistributes atmospheric constituents, including helium. Since helium has relatively large scale height compared to the other gases, less helium is returned from the winter hemisphere to the summer hemisphere, which causes the helium bulge in the winter hemisphere [Keating et al., 1997].

In addition to the characteristic of the more concentration of helium in the winter hemisphere, the other important characteristic of helium is that the partial pressure of helium significantly varies with solar activity in the upper thermosphere. The partial pressure of helium increases with the decline in the solar activity. Figure 4-10 shows the number densities of N₂, O₂, O, and He calculated from the NRLMSISE-00 model near the spring equinox (day 80) in 2001 (near the solar maximum) and Figure 4-11 shows them near the fall equinox (day 260) in 2008 (near the solar minimum). During the near solar maximum time, the helium number densities are greater than the N₂ number densities above around 450 km (Figure 4-10 (a) and Figure 4-11 (a)). On the other hand, this altitude threshold is lowered to around 370 km during the solar minimum time (Figure 4-10 (b) and Figure 4-11 (b)). Figure 4-12 illustrates the daily averaged number densities of neutral constituents N₂, O₂, O, and He from the NRLMSISE-00 model at 400 km along the track of the CHAMP satellite with solar activity from solar maximum to minimum. As shown in Figure 4-12, the effect of the decline in the solar activity is subtle to the helium number densities, while the number densities of the other neutral constituents dramatically decrease with the decline in the solar activity. As a result, the partial pressure of helium relatively increases
with the decline in the solar activity. Figure 4-13 shows the mean molecular weight with and without helium, daily averaged mass mixing ratios, and partial pressures of neutral constituents at 400 km calculated from the NRLMSISE-00 model along the track of the CHAMP satellite. The mean molecular weight of neutral constituents of N\textsubscript{2}, O\textsubscript{2}, and O excluding He, which is used in the TIE-GCM, is around 17 g/mol regardless of the solar activity since the major neutral constituents have the similar rates of change of number densities with solar activity as shown in Figure 4-10 and Figure 4-11. However, the mean molecular weight including helium varies from around 17 gm/mol to 15 gm/mol at the end of solar cycle 23 (Figure 4-13 (a)). Figure 4-13 (b) represents the mass ratios of each neutral constituent. The mass ratio of helium can be almost negligible during the solar maximum time. However, it should be considered as one of the major species during the solar minimum time, since the mass ratio of helium becomes almost the same as the N\textsubscript{2} at the end of solar cycle. Figure 4-13 (c) shows the partial pressures of neutral constituents along the track of CHAMP satellite which are used to calculate specific heat, molecular viscosity, and thermal conductivity in the TIE-GCM. While the partial pressure of helium is negligible near the solar maximum, it is greater than that of N\textsubscript{2} during the solar minimum. Therefore, excluding helium can cause considerable uncertainties for calculating specific heat, molecular viscosity, and thermal conductivity, subsequently, for calculating neutral temperature and density. The errors should be gradually increased with the decline in the solar activity because helium gradually plays an important role in the upper thermosphere near solar minimum time. Figure 4-4 represents these errors, which can be decreased if helium is incorporated into the TIE-GCM for calculation of the specific heat, molecular viscosity, and thermal conductivity.
Figure 4-10 Number density profiles of $N_2$, $O_2$, $O$, and He from the NRLMSISE-00 model near the spring equinox (day 80) in 2001 (near solar maximum). Dashed lines indicate the altitude which the helium number density equals the molecular nitrogen number density.
Figure 4-11 Number density profiles of N$_2$, O$_2$, O, and He from the NRLMSISE-00 model near the fall equinox (day 260) in 2008 (near solar minimum).
Figure 4-12 (a) Daily averaged number densities of neutral constituents from the NRLMSISE-00 model at 400 km along the track of the CHAMP satellite.  (b) Daily and 81-day averaged 10.7 cm solar flux (F$_{10.7}$ and F$_{10.7A}$, respectively).
Figure 4-13 (a) Mean molecular weight with and without helium, (b) mass mixing ratios, and (c) partial pressures of neutral constituents at 400 km calculated from the NRLMSISE-00 model along the track of the CHAMP satellite.
Helium effects on the mechanical and thermal properties in a gas mixture

Banks and Kockarts also proposed the formula of helium for the mechanical and thermal properties. Equations (4.9), (4.12), and (4.14) can be modified and shown as Eqs. (4.16), (4.17), and (4.18) when helium is included for calculating specific heat, molecular viscosity, and thermal conductivity, respectively [Banks and Kockarts, 1973].

\[
C_p = \frac{1}{2} R \left[ \frac{7}{32} P(O_2) + \frac{7}{28} P(N_2) + \frac{5}{16} P(O) + \frac{5}{4} P(He) \right] \tag{4.16}
\]

\[
\mu = [4.03 P(O_2) + 3.42 P(N_2) + 3.9 P(O) + 3.84 P(He)] \times 10^{-6} T^{0.69} \tag{4.17}
\]

\[
K_T = [56 P(O_2) + 56 P(N_2) + 75.9 P(O) + 299 P(He)] T^{0.69} \tag{4.18}
\]

Figure 4-14 shows the specific heat calculated from the equations with and without helium (Eqs. (4.9) and (4.16), respectively) using partial pressures of neutral constituents from the NRLMSISE-00 model on day 80 in (a) 2001 near the solar maximum and (b) 2008 near the solar minimum. Compared to the solar maximum period, helium greatly affects the specific heat during the solar minimum period, especially above about 300 km. Excluding helium in the upper thermosphere can cause significant error in calculating specific heat in a gas mixture and, therefore, helium should be included in the physics-based model for better estimating neutral density. On the other hand, the effect of helium on the molecular viscosity can be negligible regardless of solar activity up to 500 km. Figure 4-15 shows that the variation of molecular viscosity with and without helium can be negligible during both (a) the solar maximum and (b) the solar minimum period. Figure 4-16 represents the variation of thermal conductivity with and without helium according to solar activity. Similar to the specific heat, while the effect of helium to thermal conductivity is negligible below around 300 km, it shows remarkable variations by
helium above around 300 km where helium gradually plays a role as one of the major neutral constituents. The variations during the solar minimum time are much greater than those during the solar maximum time, which can cause the TIE-GCM to have increasing errors with the decline in the solar activity due to the exclusion of helium. Therefore, the physics of helium should be incorporated into physics-based thermospheric neutral density models, such as the TIE-GCM, in order to increase the accuracy in estimating neutral density in the upper thermosphere, especially during the solar minimum periods.
Figure 4-14 Specific heat with and without helium using number densities from the NRLMSISE-00 model, shown in Figure 4-10, on day 80 in (a) 2001 near the solar maximum and (b) 2008 near the solar minimum.
Figure 4-15 Molecular viscosity with and without helium using number densities from the NRLMSISE-00 model, shown in Figure 4-10, on day 80 in (a) 2001 near the solar maximum and (b) 2008 near the solar minimum.
Figure 4-16  Thermal conductivity with and without helium using number densities from the NRLMSISE-00 model, shown in Figure 4-10, on day 80 in (a) 2001 near the solar maximum and (b) 2008 near the solar minimum.
Improvement of the TIE-GCM

Incorporation of helium into the TIE-GCM

In order to incorporate helium effects into the TIE-GCM for the upper thermosphere, the empirical NRLMSISE-00 model was used to calculate the partial pressures of helium. Number densities of N₂, O₂, O, and He were calculated along the track of the CHAMP satellite. Then, daily averaged number densities (shown in Figure 4-12 (a)) of neutral constituents, subsequently, daily averaged partial pressures of helium (shown in Figure 4-13 (c)) were obtained to be incorporated into the TIE-GCM for calculating the molecular viscosity, thermal conductivity, and specific heat. Since the NRLMSISE-00 model has its own uncertainty for calculating neutral constituents, four cases were simulated for the effects of helium in the TIE-GCM from near solar maximum period (2002) to near solar minimum period (2007) in solar cycle 23:

- **Case 1: TIE-GCM**
  Case 1 is the simulation with the original TIE-GCM. Therefore, there are no helium effects in this simulation. The results are the same as the bottom panels in Figure 4-2 and Figure 4-3.

- **Case 2: TIE-GCM with helium from the NRLMSISE-00 model**
  Case 2 is the simulation with 100% daily averaged partial pressures of helium calculated from the NRLMSISE-00 model along the track of the CHAMP satellite. With this simulation, the general effects of helium in the upper thermosphere are incorporated into the TIE-GCM.

- **Case 3: TIE-GCM with 71% helium from the NRLMSISE-00 model**
  Case 3 is the simulation using the optimal weighted factor of the partial pressures of helium based on the results from the cases 1 and 2. Case 1 is for 0% and case 2 is for
100% of the partial pressures of helium incorporated into the TIE-GCM. An optimal weighted factor was determined in order to minimize the RMS of percent differences of daily averaged neutral densities between case 1 and case 2. Figure 4-17 shows the RMS between case 1 and case 2, and shows that 71% of helium is the optimal weighted factor for minimizing the RMS of percent differences.

- Case 4: TIE-GCM with 50% helium from the NRLMSISE-00 model

Case 4 is the simulation with 50% partial pressures of helium in order to see the sensitivity of helium.
Figure 4-17 RMS variation of percent differences with the weighted portion of helium from 0 to 1. The RMS has the minimum value when 71% helium from the NRLMSISE-00 model is incorporated into the TIE-GCM.
Simulation results

Figure 4-18 shows daily averaged neutral densities from the CHAMP accelerometer data and the TIE-GCM with cases 1 through 4 at 400 km along the track of the CHAMP satellite from near solar maximum (2002) through near solar minimum (2007) of solar cycle 23. Figure 4-19 illustrates percent differences of the four cases of the TIE-GCM from the CHAMP accelerometer shown in Figure 4-18. Figure 4-20 and Figure 4-21 show the results from the GRACE-A accelerometer data. The GRACE-B also carries the identical accelerometer to the GRACE-A on board. In this dissertation, only the GRACE-A accelerometer data were considered since the density data from the GRACE-A and -B were very similar. The neutral density data from the GRACE accelerometer data are available since day 213 in 2002. The RMS’s of the percent difference for the cases are listed in each panel in Figure 4-19 and Figure 4-21.

The percent differences of the TIE-GCM (Case 1, first panels in Figure 4-19 and Figure 4-21) gradually increase as approaching the solar minimum time appeared in the end of solar cycle because the TIE-GCM does not include the helium physics in the model causing the uncertainties to increase during low solar activity condition. As a result, the RMS’s of percent differences in the TIE-GCM show large uncertainties with the values 47.1% and 62.3% for the CHAMP and GRACE-A accelerometer data, respectively. The secular increases with the decline in the solar activity were eliminated in the Case 2 (second panels in Figure 4-19 and Figure 4-21) by including daily averaged helium from the NRLMSISE-00 model. The RMS’s of percent differences of the TIE-GCM with helium were remarkably lowered to 27.2% and 25.5% for the CHAMP and GRACE-A accelerometer data, respectively. These results show the effects of helium in the upper thermosphere during especially low solar activity conditions and show the necessity of incorporation of helium physics into the physics-based thermospheric neutral density models for better estimation in the upper thermosphere. The accuracy of the TIE-GCM can be
further increased when the optimal weighted factor was applied to the partial pressures of helium from the NRLMSISE-00 model. The RMS’s of percent differences of the TIE-GCM with 71% helium from the NRLMSISE-00 model (Case 3) are 21.4% and 22.8% for the CHAMP and GRACE-A accelerometer data, respectively. With 50% helium from the NRLMSISE-00 model (Case 4), the RMS’s of percent differences are slightly increased compared to the Case 3 (23.8% and 29.4% for the CHAMP and GRACE-A accelerometer data, respectively).

Consequently, incorporating the helium data from the empirical thermospheric neutral density model into the TIE-GCM remarkably improved the accuracy in neutral density estimation of the TIE-GCM, especially during the solar minimum time. Yearly results of these simulations are in Appendix A for the CHAMP and GRACE-A accelerometer data.
Figure 4-18 Daily averaged mass density comparisons from the CHAMP accelerometer to the four cases of the TIE-GCM at 400 km height along the track of the CHAMP satellite from 2002 to 2007: TIE-GCM, TIE-GCM with helium, TIE-GCM with 71% helium, and TIE-GCM with 50% helium from the NRLMSISE-00 model.
Figure 4-19  Percent differences of the four cases of the TIE-GCM compared to the CHAMP accelerometer from 2002 to 2007 shown in Figure 4-18.
Figure 4-20 Daily averaged mass density comparisons from the GRACE-A accelerometer to the four cases of the TIE-GCM at 400 km height along the track of the GRACE-A satellite from 2002 to 2007: TIE-GCM, TIE-GCM with helium, TIE-GCM with 71% helium, and TIE-GCM with 50% helium from the NRLMSISE-00 model.
Figure 4-21 Percent differences of the four cases of the TIE-GCM compared to the GRACE-A accelerometer from 2002 to 2007 shown in Figure 4-20.
Chapter 5

CONCLUSIONS AND FUTURE WORK

In this dissertation, the thermospheric empirical neutral density models (JB2006, JB2008, and NRLMSISE-00) and the NCAR’s physics-based model (TIE-GCM) were used to compare the accuracy in thermospheric neutral density estimation with respect to the neutral densities derived from the CHAMP and the GRACE accelerometer data during the solar maximum through the solar minimum. Helium data calculated from the NRLMSISE-00 model were incorporated into the TIE-GCM in order to improve the accuracy in density estimation since that model does not include any physics of helium. Ignoring this constituent causes errors in the upper thermosphere, especially during the solar minimum time.

Comparison of thermospheric neutral density models

The thermospheric neutral density data derived from the very accurate accelerometer data equipped in the CHAMP and the GRACE satellites were available with the sub-decimeter accuracy in satellite tracking data. Those neutral density data help to better understand the responses of the upper atmosphere to the variations of solar-terrestrial activity. The neutral density values calculated from the JB2006, JB2008, NRLMSISE-00, and TIE-GCM models were compared to those from the accelerometer data for three cases: the long-term accuracy during the year of 2002 (near solar maximum) through 2007 (near solar minimum), the geomagnetic storm time accuracy on days 323–325 in 2003, and the geomagnetic quiet time accuracy on day 188 in 2003.
Daily averaged neutral density values at 400 km calculated from the JB2008 model during the solar maximum through the solar minimum time showed the smallest RMS of percent differences among the neutral density models. The RMS values are 10.3% for the JB2008 model, 15.7% for the JB2006 model, 20.3% for the NRLMSISE-00 model, and 47.1% for the TIE-GCM. During the severe geomagnetic storm time and the geomagnetic quiet time, the JB2008 model showed the most accurate neutral density estimation with the RMS of 21.1% and 14.2%, respectively. Overall, the newest model, JB2008, has the lowest uncertainties in density estimation since it includes the new solar index for the heating in the lower thermosphere, the new semiannual density corrections, and the newly adopted $D_s$ index.

**Contribution of helium to the physics in the upper thermosphere**

Unlike in the lower atmosphere, helium plays an important role as the one of the major neutral constituent in the upper thermosphere. The contribution of helium to the physics of the gas mixture increases with the decline in the solar activity. The specific heat and the molecular thermal conductivity in a gas mixture are significantly affected by helium in the upper thermosphere, especially during the solar minimum time, and they affect the calculation of neutral temperature and density in TIE-GCM.

The JB2006, JB2008, and NRLMSISE-00 models include the helium number density for calculating neutral density, while the TIE-GCM does not include any helium physics. That causes the TIE-GCM to have increased errors of neutral density estimation in the upper thermosphere above approximately 300 km. Therefore, it is concluded that the physics of helium should be incorporated into the physics-based thermospheric neutral density models such as the TIE-GCM in order to increase the accuracy in estimating neutral density in the upper thermosphere, especially during the solar minimum period.
Incorporation of helium data into the TIE-GCM

The helium data calculated from the NRLMSISE-00 model were incorporated into the TIE-GCM so as to represent the physics of helium. As a result, the secular increases of percent differences with respect to the decline in the solar activity were eliminated. The RMS’s of percent differences in daily averaged neutral density estimation of the TIE-GCM were remarkably lowered to 27.2% and 25.5% from 47.1% and 62.3% for the CHAMP and GRACE-A accelerometer data, respectively, during near solar maximum through near solar minimum of the solar cycle 23. The accuracy of the TIE-GCM was further improved by applying an optimal weighted factor for helium data from the NRLMSISE-00 model since the NRLMSISE-00 model has its own uncertainty in helium calculation. By applying the optimal weighted factor 0.71 for the partial pressure of helium from the NRLMSISE-00 model, the RMS’s of percent differences of the TIE-GCM were lowered to 21.4% and 22.8% for the CHAMP and GRACE-A accelerometer data, respectively. It can be concluded that the accuracy of the TIE-GCM in thermospheric neutral density estimation was improved significantly by incorporating the helium data from the empirical NRLMSISE-00 model into the model, especially during the solar minimum time. This improvement can be incorporated into the next version of the TIE-GCM and motivate the NCAR to incorporate the behavior of helium into the TIE-GCM. This improvement can also be investigated by the comparisons of neutral densities from the GRACE accelerometer data and other sources during the next solar cycle 24.

Future work

In this dissertation, the helium data were calculated from the empirical NRLMSISE-00 model and these data were incorporated into the TIE-GCM for the calculation of the specific heat,
the molecular viscosity, and the molecular thermal conductivity. The chemical reactions of helium can be incorporated into the TIE-GCM in order to represent the effects of helium in the upper thermosphere.

The accuracy of the TIE-GCM can be improved with ground-based measurements imported from several sources such as airglow data from the Second Generation Optimized Fabry-Perot Doppler Imager (SOFDI) instrument (currently deployed in Peru), and the ISR data from the Arecibo Observatory (AO). For example, the analysis of the AO’s ISR archival data over the last 30 to 40 years and ISR inversion software to extract ion-neutral collision and neutral density can be performed. These data can be used to investigate the atmospheric responses between 100 km and 130 km to solar flux variability. The AO is a unique facility to conduct this research since it does not suffer from the equatorial electrojet contamination.
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Appendix A

FIGURES FOR COMPARING NEUTRAL DENSITY MODELS

This appendix provides the comparisons of yearly results in daily averaged neutral densities and percent differences at 400 km from the TIE-GCM with/without helium shown in Figure 4-18/Figure 4-19 and Figure 4-20/Figure 4-21 comparing with the CHAMP and GRACE-A accelerometer data, respectively. Figure A-1 through Figure A-6 are the comparisons between the TIE-GCM and CHAMP accelerometer data, and Figure A-7 through Figure A-12 are between the TIE-GCM and GRACE-A accelerometer data from 2002 through 2007. As shown in the figures, the effect of helium increases with the decline in the solar activity and, as a result, the errors in the TIE-GCM gradually increase from the near solar maximum to the near solar minimum period. By incorporating the helium data calculated from the NRLMSISE-00 model into the TIE-GCM, the errors during the near solar minimum period can be significantly decreased.
Figure A-1 Daily averaged neutral density comparisons from the CHAMP accelerometer to the TIE-GCM model results at 400 km in 2002. (a) Daily averaged densities at 400 km, (b) Percent differences from the CHAMP
Figure A-2 Daily averaged neutral density comparisons from the CHAMP accelerometer to the TIE-GCM model results at 400 km in 2003. (a) Daily averaged densities at 400 km, (b) Percent differences from the CHAMP
Figure A-3 Daily averaged neutral density comparisons from the CHAMP accelerometer to the TIE-GCM model results at 400 km in 2004. (a) Daily averaged densities at 400 km, (b) Percent differences from the CHAMP
Figure A-4  Daily averaged neutral density comparisons from the CHAMP accelerometer to the TIE-GCM model results at 400 km in 2005. (a) Daily averaged densities at 400 km, (b) Percent differences from the CHAMP
Figure A-5 Daily averaged neutral density comparisons from the CHAMP accelerometer to the TIE-GCM model results at 400 km in 2006. (a) Daily averaged densities at 400 km, (b) Percent differences from the CHAMP.
Figure A-6 Daily averaged neutral density comparisons from the CHAMP accelerometer to the TIE-GCM model results at 400 km in 2007. (a) Daily averaged densities at 400 km, (b) Percent differences from the CHAMP
Figure A-7  Daily averaged neutral density comparisons from the GRACE accelerometer to the TIE-GCM model results at 400 km in 2002. (a) Daily averaged densities at 400 km, (b) Percent differences from the CHAMP
Figure A-8  Daily averaged neutral density comparisons from the GRACE accelerometer to the TIE-GCM model results at 400 km in 2003.  (a) Daily averaged densities at 400 km, (b) Percent differences from the CHAMP
Figure A-9 Daily averaged neutral density comparisons from the GRACE accelerometer to the TIE-GCM model results at 400 km in 2004. (a) Daily averaged densities at 400 km, (b) Percent differences from the CHAMP
Figure A-10 Daily averaged neutral density comparisons from the GRACE accelerometer to the TIE-GCM model results at 400 km in 2005. (a) Daily averaged densities at 400 km, (b) Percent differences from the CHAMP
Figure A-11  Daily averaged neutral density comparisons from the GRACE accelerometer to the TIE-GCM model results at 400 km in 2006.  (a) Daily averaged densities at 400 km, (b) Percent differences from the CHAMP
Figure A-12  Daily averaged neutral density comparisons from the GRACE accelerometer to the TIE-GCM model results at 400 km in 2007.  (a) Daily averaged densities at 400 km, (b) Percent differences from the CHAMP
Appendix B

DIURNAL DENSITY VARIATIONS IN THE LOWER ATMOSPHERE

This appendix shows the comparisons of diurnal density variations in the lower atmosphere during the solar maximum (Figure B-1) and the solar minimum time (Figure B-2). The maximum neutral density usually occurs at 2 p.m. and the minimum density occurs at 4 a.m. local solar time. The NRLMSISE-00 model was used for calculating neutral densities between 40 km and 300 km in the location of 45° latitude and 0° longitude on day 100 in 2001 and 2008 for the solar maximum and the solar minimum time, respectively. The amplitude of the maximum density compared to the minimum density is the order of about 70% at 300 km. On the other hand, the amplitude is the order of less than 10% below about 160 km. Below around 60 km, the diurnal density variation is negligible.
Figure B-1  Diurnal neutral density variations in the lower atmosphere calculated from the NRLMSISE-00 model near the solar maximum time (day 100 in 2001).  (a) Neutral densities at 2 pm vs. 4 am and (b) Diurnal neutral density variations in percent.
Figure B-2  Diurnal neutral density variations in the lower atmosphere calculated from the NRLMSISE-00 model near the solar minimum time (day 100 in 2008). (a) Neutral densities at 2 pm vs. 4 am and (b) Diurnal neutral density variations in percent.
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