MESOSPHERE AND LOWER THERMOSPHERE NEUTRAL WINDS OBSERVATIONS USING ROCKET-RELEASED CHEMICAL TRAILS AT POKER FLAT, ALASKA

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MESOSPHERE AND LOWER THERMOSPHERE NEUTRAL WINDS
OBSERVATIONS USING ROCKET-RELEASED CHEMICAL TRAILS AT
Poker Flat, Alaska

A Dissertation
Presented to
the Graduate School of
Clemson University

In Partial Fulfillment
of the Requirements for the Degree
Doctor of Philosophy
Physics

by
Tianyu Zhan
August 2007

Accepted by:
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Dr. Bradley S. Meyer
ABSTRACT

Sounding rocket campaigns ARIA I through ARIA IV, CODA 2, HEX 1, JOULE 1 and JOULE 2 all carried out at Poker Flat Research Range at Alaska, covering the geomagnetic condition from quiet to highly disturbed. Trimethyl aluminum (TMA) were released during the rocket flights to study the mesosphere and lower thermosphere neutral wind at high-latitude region. The results of horizontal neutral wind profiles are presented. The comparison shows that under disturbed condition the wind velocity is stronger and the jet feature at the bottom side of wind maximum with unstable wind shear is lifted to a higher altitude. Under the quiet condition, the dominance forcing acting on the neutral atmosphere is the upward propagating tides below 120 km and the Lorentz force and viscosity in the region above 120 km. While under the disturbed condition, the tidal force is disrupted by Hall drag in the region of 105–125 km and the wind profile is a result of complex interplay of tidal force, Lorentz force and Joule heating. Modeling works have also been presented. The comparisons are poor for the global general circulation models and are better for localized non-hydrostatic models. It is also concluded that a detailed high-resolution time-history of auroral forcing and the upward propagating tidal forcing are both important for theoretical model to predict the small scale features of the horizontal neutral wind in the auroral E region and lower F region.
DEDICATION

To my dear parents and my lovely wife.
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I would like to appreciate my advisor, Dr. Miguel Larsen, for his continuous guidance, support and encouragement through all 5 years of my program. I would also like to appreciate the research committee member, Dr. John, Meriwether, Dr. Bradley Meyer and Dr. Gerald Leimacher for their advice and feedback following the progress of my Ph.D. project. Thanks for Dr. Sean Brittain representing Dr. Leimacher for the defense.

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Chapter 1

INTRODUCTION

1.1 Objectives

The mesosphere and lower thermosphere at high latitudes is a region with strong coupling with both the middle and lower neutral atmosphere and the ionosphere and magnetosphere. The momentum and energy sources come from the dissipation of tides, planetary waves and gravity waves propagating upward from the troposphere and stratosphere, and from the ion drag, Joule heating and particle heating. During disturbed conditions, such as substorm events, the enhanced ionization and strong electric fields will further induce stronger coupling and the accelerated neutral wind pattern at $E$ region and lower $F$ region represents a complex interaction between the auroral forcing and the tidal forcing. Mikkelsen et al. [1987] and Larsen et al. [1989] showed large horizontal neutral winds with magnitudes of 100–150 m s$^{-1}$ in the region of 100 to 120 km altitude, observed by the chemical tracers released by sounding rockets under quiet conditions on March 20, 1985, and March 21, 1987, from Søndre Strømfjod, Greenland. Even larger westward horizontal neutral winds were observed during February and March 1978 by rocket-released trimethyl aluminum (TMA) at Poker Flat, Alaska [Mikkelsen et al., 1981]. Yet the dynamics in the mesosphere and lower thermosphere at high latitudes are still not well understood primarily due to the fact that the direct response of the neutral atmosphere is weaker than at higher
altitudes and the fact that the observational database is sparse [Goncharenko et al., 2004]. The large-scale circulation in the thermosphere at high latitudes has been investigated in many modeling studies [e.g., Roble et al., 1982; Killeen and Roble, 1986; Fuller-Rowell et al., 1996], but the small-scale structure in the horizontal neutral winds in the lower thermosphere has been reproduced with mixed results, at best.

The objective of this dissertation is to present a group of mesosphere and lower thermosphere horizontal neutral wind observations and simulations from large-scale general circulation models and localized models and to discuss the different forcing for the neutral winds under both quiet and disturbed conditions. The observations were obtained using TMA trails released by sounding rockets launched at Poker Flat, Alaska, during the ARIA [Larsen et al., 1997], CODA 2, HEX 1 [Wescott et al., 2006], JOULE 1 and JOULE 2 campaigns. While the theoretical models include the NCAR TIME-GCM [Roble, 1993; Roble and Ridley, 1994], the University of Michigan GITM [Ridley et al., 2006], the STGCM [Mikkelsen and Larsen, 1991], and the localized models of Walterscheid and Lyons [1989] and Parish et al. [2003].

1.2 Organization of Dissertation

The objective of the dissertation was described above. In order to give some background of the dynamics in the mesosphere and lower thermosphere at high latitudes, Chapter 2 describes basic features of the atmosphere and ionosphere, and discusses the governing momentum equation and different forcing terms. In Chapter 3, the prevailing experimental methods used currently for measuring the middle/upper atmosphere neutral winds are introduced and compared. The advantage of using chemical tracers released by sounding rockets to measure the horizontal neutral wind velocities, especially during disturbed conditions, is emphasized. Chapter 4 presents the neutral wind profiles obtained from the sounding rocket experiments at Poker Flat, Alaska, during the ARIA, CODA 2, HEX 1, JOULE 1 and JOULE 2 campaigns. The common features from the observations during
both quiet and disturbed condition will be summarized. In Chapter 5, model simulation studies for those sounding rocket campaigns will be presented, including global general circulation models and localized models. Finally, the different forcing responsible for the horizontal neutral wind patterns observed during quiet and disturbed conditions will be discussed and a conclusion will be given.
Chapter 2

NEUTRAL WIND DYNAMICS IN MESOSPHERE AND LOWER THERMOSPHERE

2.1 The Atmosphere

The Earth’s atmosphere is usually considered horizontally stratified owing to the influence of gravity. The layers of the neutral atmosphere are organized according to the vertical structure of the temperature profile. A typical high-latitude profile of temperature is given in Figure 2.1, obtained from the Mass-Spectrometer-Incoherent-Scatter extended model (MSIS-E-90) [Hedin, 1991]. The lowest layer is the troposphere adjacent to the earth’s surface with an upper boundary that varies from 6 to 16 km. The atmospheric temperature decreases with altitude from the Earth’s surface with a lapse rate that is close to the pseudo-adiabatic value $6.5^\circ$K/km in the troposphere. The upper boundary, called the tropopause, is approximately isothermal and is usually lower at the poles. Above the tropopause is the stratosphere where the temperature increases with the height, primarily due to the absorption by ozone of part of the ultraviolet portion of the solar radiation. The top of the stratosphere at the altitude 50 to 60 km is called stratopause. The mesosphere starts at the top of the stratopause where radiative cooling creates a sharp temperature decrease with increasing altitude with a minimum in the range of 130 – 190 K at about 90–95 km altitude. For the region above the altitude of the temperature minimum, the
so-called mesopause, the temperature increases dramatically with height to values that are quite variable, ranging from 600 to 2000 K. As shown in the Figure 2.1, the thermospheric temperature maximizes at about 800 K. The temperature increase in the thermosphere is at least partially explained by absorption of the extreme ultraviolet (EUV) portion of the incident solar radiation spectrum.

The atmosphere is relatively uniform in composition in the region below 100–110 km, called the homosphere, due to mixing effects from turbulence. The turbulence ceases above the turbopause and the atmospheric constituents begin to separate according to their various masses. A typical high-latitude atmospheric constituents profile is shown in Figure 2.2. Below the turbopause, molecular nitrogen $N_2$ and molecular oxygen $O_2$ dominate the gas mixture and have the relatively uniform ratio of about 4:1. Above about 110 km, the density of atomic oxygen $O$ exceeds that of $O_2$, and above 200 km, exceeds that of $N_2$. This trend is due to the photodissociation of $O_2$ by solar EUV radiation coupled with the absence of turbulent mixing effects above the turbopause. In the region above 500–600 km altitude, atomic hydrogen $H$ and atomic helium $He$ start to dominate.

2.2 The Ionosphere

The incident electromagnetic radiation (ultraviolet and shorter wavelengths) and streaming high-energy solar particles modify atmospheric gases by creating ionization that envelopes the earth in the region called the ionosphere. The radiation and particles penetrate the atmosphere and ionize and excite neutral constituents until the energy is exhausted. The collision between the plasma and neutrals further produce ionization. At high latitudes, particle precipitation is also responsible for the visible emission known as the aurora.

The intensity of ionization grows exponentially with altitude, while the neutral atmospheric gas density decays with increasing altitude. Conventionally, the ionosphere is labeled as follows: the $D$, $E$ and $F$ regions, corresponding to the three primary layers
with enhanced electron densities. The F region ionosphere includes the altitude region from about 180 km to a third of the Earth’s radius and has a peak plasma number density around about 400±50 km. The F region peak number density has a typical daytime value of about $10^6 \text{ cm}^{-3}$ for solar maximum conditions. The $E$ region ionosphere extends from about 80 to 150 km altitude with a decline in ionization below about 120±5 km. Typical $E$ region noon-time peak density can be about $10^5 \text{ cm}^{-3}$ at 106 km under solar maximum conditions. The $E$ region ionosphere is characterized by large ionospheric current systems and plasma instabilities. The $D$ region ionosphere extends from about 50 to 90 km. The ionization in the $D$ region is highly collisional and is strongly controlled by neutral motions.

Vertical profiles of electron number density calculated using the International Reference Ionosphere (IRI) model [National Space Science Data Center, 2001], are plotted in Figure 2.3. The figures show typical daytime profiles. The electron density peaks for the $E$ region and $F$ region are evident. The IRI model profiles are imperfect but give a general idea of the characteristics of ionospheric parameters. The IRI model specification is an empirical model encompassing data from ionosondes, incoherent scatter radars, rockets, and satellite measurements [Bilitza, 1990].

Typical plasma composition profiles for the Earth’s ionosphere are shown in Figure 2.4. The composition values were obtained from the IRI model. The figure shows how plasma composition in the ionosphere varies with altitude in percentage of the total plasma density. This inhomogeneity is introduced by different weights of ionospheric species, including electrons, and by the quasi-neutrality condition and the resulting ambipolar electric field. Plasma composition depends also on re-combination rates and radiative chemical processes, as well as the intensity of ionizing radiation and ionization cross-sections.

The composition in the $E$ and lower $F$ region is dominated by molecular ion species $O_2^+$ and $NO^+$. From the lower $F$ region to about 600 km, the ionosphere is mainly populated by atomic oxygen ions, $O^+$. Above about 600 km and extending to the magnetosphere, the ionosphere is largely a proton plasma, $H^+$, along with a small fraction of $He^+$. 
2.3 Equations of Motion of the Neutral Air

The neutral atmospheric dynamics in the mesosphere and lower thermosphere can be described by the momentum equation 2.1 [Rishbeth and Garriott, 1969]. The neutral air velocity is $U$, with its $x$-component being positive toward east, $y$-component positive toward north and $z$-component positive upward. The operator $d/dt$ represents "differentiation following the motion" of the air parcel, and can be expressed by Equation 2.2. The forcing terms are on the right-hand side of Equation 2.1. In the equation, $p$ is pressure and $\rho$ is neutral density, as usual.

$$\frac{dU}{dt} = -2\Omega \times U + g - \left(\frac{1}{\rho}\right)\nabla p - \nabla \Psi + \left(\frac{\mu}{\rho}\right)\nabla^2 U - \nu_{ni}(U - V_i)$$  \hspace{1cm} (2.1)

$$\frac{dU}{dt} = \frac{\partial U}{\partial t} + (U \cdot \nabla)U$$  \hspace{1cm} (2.2)

The first forcing term $-2\Omega \times U$ represents the Coriolis acceleration, where $\Omega$ is the Earth’s angular velocity with the vector direction along the Earth’s axis of rotation toward the geographic north pole. The Coriolis effect is due to the rotation of the Earth when we consider the atmospheric dynamics in the rotating frame of reference where the Earth is stationary. The Coriolis force turns the moving object to the right in the northern hemisphere and to the left in the southern hemisphere. A visual example is the flow around a low pressure system, which spins counter-clockwise due to the balance between the Coriolis force and the pressure gradient force in the northern hemisphere and spins clockwise in the southern hemisphere.

The variable $g$ is the Earth’s gravitational acceleration, usually considered vertical downward toward the center of the Earth. The small centripetal acceleration terms $\Omega \times (\Omega \times r)$ may also be combined with the gravitational term. Here $r$ is the geocentric distance.

The pressure gradient term is given by $-\left(\frac{1}{\rho}\right)\nabla p$, which accelerates a parcel of air from a high pressure region to a low pressure region. The pressure gradient force acts at
right angles to isobars in the direction from high to low pressure.

A simple form of the viscosity term is given by \((\frac{\mu}{\rho})\nabla^2 U\), in which \(\mu\), the coefficient of molecular viscosity, is assumed to be a constant. Actually \(\mu\) depends on temperature and molecular mass [Sutton, 1953].

\(\Psi\) is a scalar potential due to tidal forcing and \(\nu_{ni}(U - V_i)\) represents the ion-drag forcing, in which \(\nu_{ni}\) is the neutral-ion collision frequency and \(V_i\) is the ion drift velocity. These two terms will be discussed in more detail later.

The large-scale winds in the lower atmosphere are usually the result of the balance between the horizontal pressure gradient and the Coriolis force, i.e., the so-called "geostrophic wind". Using \(x\) and \(y\) to denote the local Cartesian coordinates in the northward and eastward directions, respectively, then the zonal and meridional components of the geostrophic wind are given by

\[
2\Omega u_g \sin \phi = \rho^{-1} \frac{\partial p}{\partial y} \tag{2.3}
\]

\[
-2\Omega v_g \sin \phi = \rho^{-1} \frac{\partial p}{\partial x} \tag{2.4}
\]

In the simplest approximation, the prevailing large-scale winds in the mesosphere and lower thermosphere can be considered to be essentially geostrophic and driven by seasonally-varying pressure variations. In the thermosphere above 120 km, the large-scale winds are driven by daily temperature variations.

2.4 Tides

Atmospheric tides are global scale oscillations excited primarily by solar heating, although lunar gravitational forces can also excite much smaller amplitude modes. The tidal perturbations can either move relative to a fixed rotating Earth frame as migrating tides, or stay at the fixed location in the rotating Earth frame as stationary or non-migrating
tides. Gravitational tides are primarily semidiurnal having a 12-hour period and a horizontal wavelength equal to half of the Earth’s circumference. Thermal tides due to the periodic solar thermal energy input can be diurnal with a 24-hour period, semidiurnal with a 12-hour period, or some other period that is an integer fraction of a day. In the mesosphere and lower thermosphere, the diurnal and semidiurnal tides dominate which arise from the heating associated with the absorption of solar radiation by water vapor and ozone. In the thermosphere above 250 km, the dominant tide is diurnal, produced in situ by the absorption of UV and EUV solar radiation by the thermosphere.

Full theoretical treatments of the atmospheric tides have been given by Wikes [1949]. A momentum equation like Equation 2.1 excluding the ion drag term, with continuity equation and thermodynamics equation, can be solved for the possible oscillation modes of the atmospheric tides. The tides excited from lower atmosphere can also propagate vertically upward. The conservation of wave energy leads to growing tidal amplitudes that vary as $\rho^{-1/2}$ as the height increases. The vertical wavelength of the diurnal tide exited by ozone and water vapor heating and propagating upward was found to be approximately 25 km by Lindzen [1967].

2.5 Gravity Wave

Atmospheric gravity waves are smaller scale oscillations that arise from the restoring force of buoyancy forces in the atmosphere. Gravity wave periods range from minutes to hours, while the vertical wavelength is a few kilometers and the horizontal wavelength can be up to thousands of kilometers. In the lower atmosphere, the gravity waves can be excited by instabilities, the flow of air over mountains, thunderstorms, volcanic eruptions and earthquakes. In the upper atmosphere, gravity waves can be generated by unstable shears, by the motion of the terminator, or by uneven Joule and particle heating rates and the Lorentz forcing at high latitudes.

Neglecting the viscous effects, thermal conduction, Coriolis and centrifugal accel-
eration terms, the momentum, continuity and energy equations that are applied to the problem of gravity wave propagation are given in Equation 2.5, Equation 2.6 and Equation 2.7, respectively [Shunk and Nagy, 2000]. The solution shows that gravity waves have upward energy flow and downward phase velocity. The amplitude also grows as $\rho^{-1/2}$, similar to what occurred for tides, which indicates that upward propagating gravity waves are likely to break and to be dissipated somewhere in the upper atmosphere. Hines [1960] also pointed out that atmospheric molecular viscosity and thermal conductivity provided important dissipative mechanisms for gravity waves. Gravity waves also set the plasma in the ionosphere in motion via ion drag forces, which leads to electric current and energy loss by Joule heating [Rishbeth and Garriott, 1969].

\[
\frac{d\mathbf{U}}{dt} = \mathbf{g} - \left(\frac{1}{\rho}\right)\nabla p \tag{2.5}
\]

\[
\frac{\partial \rho}{\partial t} + \nabla \times (\rho \mathbf{U}) = 0 \tag{2.6}
\]

\[
(\frac{\partial}{\partial t} + \mathbf{U} \times) p + \gamma p (\nabla \times \mathbf{U} \times) = 0 \tag{2.7}
\]

### 2.6 The Dynamics in High Latitude Region

At high-latitudes, the plasma drifts can be very large due to the forcing of magnetospheric origin, and thus the ion drag term in the momentum equation will be important for the neutral atmospheric dynamics. The ion drag force is arises from the $\mathbf{J} \times \mathbf{B}$ Lorentz force in the momentum equation for the neutral gas where

\[
\mathbf{J} = \sigma \cdot (\mathbf{E} + \mathbf{U} \times \mathbf{B}) \times \mathbf{B} \tag{2.8}
\]

Here $\mathbf{E}$ is the external electric field.

In the high-latitude region, the plasma in the ionosphere is primarily driven by the
electric field mapping down along the magnetic field lines from the magnetosphere where solar wind-driven electric fields are generated. Diagrams that illustrate the phenomenon can be found in the text by Kelley [1989]. When the interplanetary magnetic field (IMF) turns southward, the magnetic field lines in the solar wind reconnect to the Earth’s magnetic field lines and create open field lines that map into the polar ionosphere. The solar wind plasma is collisionless and creates a dynamo effect when traversing magnetic field lines, inducing electric fields $E_{sw} = -V_{sw} \times B_{sw}$. The Earth’s magnetic field lines are usually considered to be equal-potentials and thus the electric potential across the connected magnetic field lines will be mapped down to the ionosphere at the polar cap. Thus this dawn-to-dusk electric field drives the ionospheric plasma in the antisunward direction across the polar cap. The return flow is at lower latitudes and forms a two-cell convection pattern. When the IMF is northward, the ionospheric plasma flow pattern is more complicated, often with more than two convection cells.

The disturbed conditions studied in this dissertation refer to the ionosphere during auroral substorm events. The substorms correspond to the explosive release of energy in the auroral region near midnight magnetic local time [Akasofu, 1964]. A substorm event typical lasts one hour to several hours. Associated with substorms are enhancements in the electric fields, particle precipitation, and field-aligned and electrojet currents in the local region. The electron density increases due to the enhanced ionization due to particle precipitation. As a result, the neutral atmospheric flow will be significantly affected though the Lorentz forcing as a momentum source and the Joule heating as an energy source during the substorm events.
Figure 2.1: Typical neutral atmosphere temperature profile in the high-latitude region (from the MSIS-E-90 model).
Figure 2.2: Typical neutral composition profile for the Earth’s atmosphere (from the MSIS-E-90 model).
Figure 2.3: Typical daytime electron density profile in the ionosphere (from the IRI model).
Figure 2.4: Typical ion density profile in the ionosphere (from the IRI model).
Chapter 3

EXPERIMENTAL METHODS FOR MESOSPHERE AND LOWER THERMOSPHERE NEUTRAL WIND OBSERVATION

Many techniques have been developed to measure the neutral wind velocities in the middle and upper atmosphere. In this Chapter, major techniques used currently will be introduced, including light detection and ranging system, meteor radar, MF radar, incoherent scatter radar, Febry-Perot interferometer, satellite-borne instruments and chemical tracers released by sounding rockets. The advantage and disadvantage of each technique will be discussed, especially for the neutral wind measurements under disturbed condition.

3.1 Light Detection and Ranging System

The Light Detection and Ranging (LIDAR) systems have been constructed and used routinely to study the structure of the middle atmosphere (30–110 km). Different techniques for the LIDAR systems have been developed to obtain measurements of temperature and winds, among which Rayleigh scattering and resonance fluorescence are major techniques.

Rayleigh lidar relies on the Rayleigh scattering from the air molecules in the height
range 25–60 km, measuring the Doppler broadening and shift of the backscattered echo to obtain the temperature and line-of-sight winds. Resonance fluorescence techniques, dependent on the fluorescence cross sections of mesospheric metals such as Na, Fe, and K, have evolved from broadband dye laser systems into sophisticated narrowband systems that measure the temperature and winds in the mesopause region (80–105 km) [Gardner, 2004, and references therein]. The Maui lidar at Haleakala, Hawaii, and the University of Illinois sodium lidar at Starfire Optical Range located near Albuquerque, New Mexico, are sodium lidar systems, and have been used to study the atmospheric stability in the mesopause region by Li et al. [2005] and Zhao et al. [2003], among others. Comparisons have been made between the wind measurements made by the Maui Na lidar and a meteor radar [Franke et al., 2005], the University of Illinois Na lidar and a partial reflections drift radar [Liu et al., 2002], and the University of Illinois Na lidar and TMA trails released by sounding rockets [Larsen et al. 2003]. Good agreement was found within the overlapping altitude ranges.

The lidar system has excellent resolution of a few hundred meters and a few minutes, making it an excellent instrument to study the small-scale variability in the middle atmosphere. However, its upper altitude limit of 105–110 km makes observations of thermospheric dynamics of limited use. The systems that are capable of high-resolution wind measurements also require relatively large telescopes, and only a few such facilities exist. Of those, only the ALOMAR lidar system is located at high latitudes.

3.2 Meteor RADAR, MF RADAR and Incoherent Scatter RADAR

Meteor radars are VHF or HF radars that are used for meteor detection and measurements of middle atmospheric winds obtained by tracking the meteor trail drifts [Roper, 1975; Liu et al., 1995; Cervera and Reid 1995]. For example, the University of Adelaide VHF meteor radar operates at 54.1 MHz and is located at Buckland Park, Australia. The system was built as a collinear, coaxial (COCO) array and is designed to study winds in the altitude range from 2 to 20 km and 60 to 80 km using both Doppler and spaced antenna (SA) techniques [Vincent et al., 1987; Cervera and Reid 1995]. Yet the meteor drift wind measurements are confined to the altitudes where meteor ablation occurs, i.e., the altitude
range from approximately 80 to 100 km.

MF radar uses the spaced antenna (SA) technique to observe atmospheric winds in the mesosphere and lower thermosphere. The spaced antenna technique was originally used for total reflection experiments, in which the incident radio waves were totally reflected from ionization layers, primarily in the mesosphere [Hocking and Thayaparan, 1997], and later was modified for \( D \) region work using partial reflections [Hocking et al., 1989]. Another kind of MF radar uses the full-correlation-analysis (FCA) technique [Meek, 1980] providing Doppler velocity estimates of the middle-atmospheric neutral wind field, including the vertical wind component. During daytime, the radar coverage is essentially continuous from about 60 to 99 km, while it is more limited within the range from 84 to 99 km at night [Franke et al., 2001].

The incoherent scatter radar technique has also been used for monitoring the ionized structures in the ionosphere. The mesosphere and lower thermosphere neutral winds can be derived owing to the strong coupling between the neutral and ionized components in the \( D \) and lower \( E \) regions. A number of incoherent scatter radars have been built, including the European Incoherent Scatter Radar (EISCAT) located close to Tromsø, Norway, at 69.5°N [Folkestad et al., 1983], the 49.92 MHz incoherent scatter radar at Jicamarca Radio Observatory located near Lima, Peru, at 11.95°S and 76.87° [Farley, 1969]. Neutral wind data measured by EISCAT during Common Program One (CP-1) have been used to study the tidal motions by Williams and Virdi [1989] and the diurnal modes were found to vary considerably both in amplitude and phase from day to day. Brekke et al. [1995] and Nozawa and Brekke [1995] used the CP-1 wind data to study the \( E \) region neutral wind in the quiet and disturbed auroral ionosphere respectively, and will be discussed later in Chapter 6. The biggest disadvantage of using radar data to study the \( E \) region neutral wind is that the derivation of the neutral wind from the observed ion velocity requires a solution of the steady state ion momentum equation [Brekke et al., 1974] with a number of assumptions that may or may not be valid, especially at high latitudes during disturbed conditions. The results may misrepresent the real neutral wind during disturbed condition when there is
rapidly varying auroral forcing.

An inter-comparison of the middle-atmospheric neutral winds measured by meteor radar, MF radar and incoherent scatter radar during the Arecibo Initiative in Dynamics of the Atmosphere 1989 campaign has been interpreted by Hines et al. [1993]. Liu et al. [2001] compared the wind measurements at two layers around 86 and 93 km by meteor radar and sodium lidar at Starfire Optical Range near Albuquerque, New Mexico, and found the mean values had no statistically significant differences.

3.3 Febry-Perot Interferometer

Fabry-Perot interferometers (FPI) have also been used extensively to observe the thermospheric winds, especially in the $F$ region. The technique measures the neutral winds by obtaining the Doppler shifts in the Fabry-Perot fringes for night-airglows or auroral emission lines. The atomic oxygen OI 557.7 nm airglow (green line) centered at an altitude of about $\sim 97$ km has been used by the instrument operated at the University of Saskatchewan ($52^\circ$N, $107^\circ$W) [Phillips et al., 1994], for example, but the majority of the instruments make use of the Doppler shift from the OI 630 nm airglow (red line) with peak emission altitudes that vary between 250 km (solar minimum) and 300 km (solar maximum), such as the FPI located at Arequipa, Peru ($16.2^\circ$S, $71.5^\circ$W) [Meriwether and Biondi, 1995]. Using the upper thermospheric horizontal neutral wind measured by Fabry-Perot interferometer over 7 locations in both hemispheres, the climatologies of the nighttime upper thermospheric winds during quiet time ($Kp < 3$) were analyzed as a function of local time, solar cycle, day of year, and the interplanetary magnetic field (IMF) [Emmert et al., 2006].

Fabry-Perot instruments are able to provide spatial structure in the thermospheric horizontal wind [Conde and Smith, 1998], yet the observation is confined at the altitude where the airglow occurs.
3.4 Satellite-borne Instruments

Mesospheric and thermospheric neutral winds can also be measure by satellite-borne instruments, such as the Wind and Temperature Spectrometer (WATS) on the Dynamics Explorer 2 (DE 2) satellite and The Wind Imaging Interferometer (WINDII) on Upper Atmosphere Research Satellite (UARS).

Spencer et al. [1981] described the design of WATS. Two of the three wind components, the temperature and the concentration of the dominant constituent, can be measured to an altitude of approximately 650 km, while the third component can be measured to ∼375 km. Measurements of the horizontal and vertical components of the wind normal to the spacecraft velocity vector are made by interpreting the modulation of the particle stream entering the mass spectrometer. Innis and Conde [2002] used the WATS measurements on DE 2 to study the large-scale distribution of thermospheric vertical velocities within the altitude range from ∼250 to 650 km and found a clear relationship of the vertical wind activity to geomagnetic activity, characterized by the $AE$ index.

WINDII employs Doppler interferometry to measure the small wavelength shifts of the airglow emission over the altitude range from 80 to 300 km induced by atmospheric motion. Zhang et al. [2003] analyzed the neutral wind measurements by WINDII in 1992–1997 over Millstone Hill (42.6°N) and found good agreement with the measurements from the ground-based incoherent scatter radar at Millstone Hill during daytime.

Satellite in situ wind measurements give the large-scale spatial variations of the wind along the satellite track, but the measurements can only be obtained above a certain location when there is an overpass so there is poor temporal coverage.

3.5 Chemical Tracers Released by Sounding Rockets

The idea of using a chemical tracer (sodium) released by sounding rockets at high altitudes to obtain neutral wind velocities was first proposed by Bates [1950]. Numerous experiments with sodium and other chemical tracer releases were carried out thereafter.
Sodium, lithium and trimethyl aluminum (TMA) have been used extensively as chemical tracers during the past four decades, all working well in the altitude range between approximately 80 and 180 km. Other tracers with limited use include nitrous oxide and nickel carbonyl. Sodium trails are visible due to resonant scattering of sunlight at a wavelength of 5890 Å, but it is extinguished quickly after sunset [Bedinger et al., 1958]. Therefore, wind measurements using sodium tracer are limited to dawn or dusk when the tracer is illuminated at high altitudes and the ground site is located where it is dark enough to observe the trails. Lithium tracers are visible due to red resonant emissions and thus have the same limitation as sodium [Larsen 2002]. The exceptionally bright resonant emission produced by lithium makes daytime neutral wind measurements possible however [Rees et al., 1972]. TMA has fewer limitations for nighttime observations since TMA reacts spontaneously with oxygen to produce AlO in the atmosphere and produces chemiluminescence as a by-product of the reaction, which makes it visible without solar illumination [Rosenberg et al., 1963]. TMA also has additional blue resonant emissions that are visible when the trails are released in twilight conditions. TMA can be used as a chemical tracer at any time during the night since it is chemiluminescent. A further advantage of TMA is that it is a liquid, and the flow rate can be adjusted by changing the size of the exit nozzle. The release can also be modulated by using a solenoid valve to turn the flow on and off in a predetermined sequence during the flight [Larsen 2002]. In addition, spectrometric measurements of TMA re-radiation can be used to obtain temperature profiles [Rees et al., 1972]. TMA quickly become popular as a chemical tracer after its first test by Rosenberg et al., [1963]. Gun launchings have been used as a alternative and inexpensive way to release the chemical tracers [Murphys et al., 1966], although the projectiles only reached a height of 160 km which provided a more limited altitude coverage than sounding rockets.

The chemical tracer released by the sounding rocket along the trajectory forms a trail, which will be highly distorted by the neutral winds and spreads out because of molecular diffusion. The drift of the trail thus gives the atmospheric wind profile [Groves, 1960]. By taking photographs of the chemical trails from two or more geographically separated
ground observation sites, the geographical position (latitude, longitude and altitude) can be determined using triangulation techniques applied to simultaneous trail images from different sites. By matching the background stars in the photographs to the Smithsonian Astrophysical Observatory (SAO) or other star catalogs, the look angle (azimuth and elevation) of the line-of-sight for the chemical trail within the image frame from each ground station can be determined. The three-dimensional geographical position in space of the chemical trail can thus be obtained by a triangulation technique, which calculates the intersection positions for portions of the trail observed from different sites. Having a sequence of the spatial positions of the chemical trail, the motion of the trail, i.e., the horizontal neutral wind velocity can be calculated by least squares fitting the geographical position (latitude and longitude) versus time. It is assumed that the horizontal neutral wind is uniform during the observation period, which is about 3 to 5 minutes, and the vertical wind is small enough to be neglected.

The uncertainty of the triangulation comes from several sources [Larsen et al., 1995, Wescott et al., 2006]. The first arises from the accuracy of the star fitting. Using least squares fitting helps to eliminate lens aberration effects. The second uncertainty source is from identification of features in the chemical trail. The chemical trail is a diffuse, optically thin cloud and usually the midpoints of the chemical trail are picked from photographs to represent the spatial position of the trail, but the line-of-sight to one trail midpoint may not exactly intercept the line-of-sight to the chosen cloud midpoint in the photographs from another site. Therefore, the appearance of the chemical trail in the photograph and the feature identification by the operator become very sensitive, although currently human operators are still far better than any existing numerical algorithm for identifying common features in the trail images. Currently TMA trails are released in puffs, which makes it much easier for the operator to identify common features in the trail in images taken from different sites. Yet another significant problem is associated with the so-called "re-entry bag", which occurs in the TMA trail releases during the downleg portion of the rocket trajectory when there is also a release during the upleg portion. The effect is due to the TMA that freezes
on the rocket skin during the upleg portion of the flight and then is released when the rocket reenters the lower $E$-region, creating a bright blob at the lower end of the TMA trail. This makes it difficult for the operator to identify the midpoint of the TMA trail and thus leads to larger errors in the triangulation. The error can be determined from the minimum separation between the lines-of-sight from two or more different locations. Another uncertainty is from the assumption of the small vertical wind, including the sedimentation of the chemical tracer. Usually the sedimentation can be neglected. Wescott et al. [2006] found that the TMA sedimentation rate is less than 1 m s$^{-1}$ even at 160 km, based on MSIS model atmosphere values for the background densities. However, the vertical wind may affect the movement of the chemical trail, especially at higher altitudes [Groves, 1960]. Typical total uncertainties for current triangulation results using TMA trails are in the range of 5–10 m s$^{-1}$.

The advantage of using chemical tracers released by sounding rockets is that it gives the in-situ measurements of the neutral wind in the mesosphere and lower thermosphere under any geomagnetic conditions. Especially for the disturbed conditions, such as during geomagnetic substorms or storm events, the result derived from the chemical trails would be the best available representation of the local neutral wind velocities. The altitude resolution of the neutral wind can be as small as 0.5 km. However, due to the high expense, strict safety requirements for rocket launch conditions and the requirement for clear sky at different ground sites for photographing the chemical trails, the measurements are much less frequent than those from other instruments, such as radar, lidar, FPI and satellites. Another limitation is that the measurements can only be made at certain geographical locations and for limited time intervals of less than 20 minutes for each chemical trail. Although several chemical trails can be deployed at different locations nearly simultaneously or at the same location in a time sequence, the spatial and temporal gradients of the neutral winds derived from the measurements are very limited.
Chapter 4

NEUTRAL WIND OBSERVATIONS

The chemical tracer technique for wind measurements on sounding rockets has been an important tool for investigations of the horizontal neutral wind in the mesosphere and lower thermosphere during both quiet and disturbed magnetic conditions. Hundreds of sounding rockets experiments have been carried out around the world over the last several decades [Larsen 2002]. Among all the launch sites, Poker Flat Research Range is of particular interest because of its location within the auroral oval. Poker Flat Research Range is a sounding rocket launch facility located at 65°07’ north latitude and 147°28’ west longitude. It is operated by the Geophysical Institute of the University of Alaska at Fairbanks. The horizontal neutral wind results from the sounding rocket experiments carried out at Poker Flat Research Range form a unique data set for analyzing the high-latitude atmosphere neutral dynamics.

In this chapter, the results of horizontal neutral wind measurements from eight sounding rocket experiments carried out at Poker Flat Research Range will be presented, including Atmospheric Response in Aurora (ARIA) I to IV (1992 to 1995) [Larsen et al., 1997], CODA 2 in 2002, Horizontal E-Region Experiment (HEX) in 2003 [Wescott et al., 2006], JOULE I in 2003 and, most recently, JOULE II in 2007. The eight experiments covered geomagnetic conditions ranging from quiet levels to moderately and highly disturbed conditions.
4.1 ARIA

The Atmospheric Response in Aurora (ARIA) campaign was a series of four sounding rocket experiments carried out at Poker Flat Research Range, Alaska. The objective was to measure the response of the E-region neutral winds to substorm activity in the post-midnight sector of the auroral oval [Larsen et al., 1997]. Four experiments were conducted during quiet (ARIA IV), moderate (ARIA I), moderate to high (ARIA III) and high (ARIA II) activity levels. TMA trails were released during the rocket flights and the triangulation technique was used to obtain the neutral wind profiles. The wind measurements were made after one hour of forcing associated with a substorm event during each level of disturbed conditions. The real-time measurements from the Alaska magnetometer chain showed maximum currents located between the launch site and the impact point of the rockets in each case. In the following subsections, the wind measurements from the four ARIA experiments will be presented.

4.1.1 ARIA I (moderate magnetic activity)

The ARIA I experiment was carried out at 1400 UT on March 3, 1992, at Poker Flat Research Range, and the result has been analyzed by Larsen et al. [1995], Brinkman et al. [1995], Anderson et al. [1995] and Hecht et al. [1995]. The value of the 3-hourly Kp index on the launch day ranged from 2 to 3, as shown in Figure 4.1. The planetary 3-hour Kp index is the mean standardized K-index from 13 geomagnetic observatories between 44
degrees and 60 degrees northern or southern geomagnetic latitude, which is scaled from 0 to 9 expressed in thirds of a unit. The K-index is a quasi-logarithmic local index of the 3-hourly range in magnetic activity relative to an assumed quiet-day curve for a single geomagnetic observatory site. First introduced by J. Bartels in 1938, it consists of a single-digit 0 through 9 for each 3-hour interval of the universal time day (UT). The launch of the rockets was about one hour after the commencement of a substorm in the post-midnight sector. The Alaska magnetometer chain showed near 500 nT deflection close to Poker Flat and the strongest current north of the launch site.

A Nike-Tomahawk sounding rocket was launched along an azimuth 5° east of north. The trimethyl (TMA) trail was released on the downleg portion of the trajectory within the altitude range from 190 to 90 km. Photographs of the TMA trail were taken at Poker Flat, at Arctic Village and at Cold Foot. The triangulation technique was then used to determine the motion of the TMA as a function of altitude and thus to calculate the mean horizontal neutral wind by least squares fit. The horizontal neutral wind profile observed during the ARIA I experiment is shown in Figure 4.2 (Figure 5 in Larsen et al., 1997). The solid curve represents the zonal wind, positive to the east, and the dashed curve represents the meridional wind, positive to the north. The wind profile covers the altitude range from 90 to 140 km. Both the zonal and meridional wind oscillated in magnitude and direction below 130 km. There was a wind peak in the lower E-region with a magnitude of \( \sim 125 \) m s\(^{-1}\) near 115 km and a direction toward southwest. The wind shear at the bottom of the peak was 200 m s\(^{-1}\) over 5 km. The neutral wind above 130 km increased with altitude, mainly due to the Pedersen drag. The hodograph corresponding to the wind profile is shown in Figure 4.3 (Figure 6 in Larsen et al., 1997). The hodograph plots zonal wind versus meridional wind and shows the changes in the wind vector as a function of altitude. The curve in the hodograph represents the movement of the wind vector tip. The wind vector observed during the ARIA I experiment rotated clockwise, following a slightly elongated circular trajectory.
Figure 4.2: Wind profile observed during the ARIA 1 experiment on March 2, 1992. After Larsen et al., [1997].

Figure 4.3: Hodograph corresponding to the wind profile observed during the ARIA 1 experiment on March 2, 1992. The horizontal scale represents the zonal velocity and the vertical scale represents the meridional velocity. After Larsen et al., [1997].
4.1.2 ARIA II (high magnetic activity)

The ARIA II experiment was carried out at 1308 UT on February 12, 1994, at Poker Flat Research Range, Alaska. The result was analyzed by Odom et al. [1995]. The magnetic conditions during the ARIA II launches were the most disturbed of the four experiments. The Kp index, shown in Figure 4.4, was mostly greater than 4 during the day of the launch. The maximum deflection detected by the Alaska magnetometer chain exceeded 1000 nT, with the strongest current located between the upleg and downleg portion of the rocket flights [Larsen et al., 1997]. The launch of the chemical release rockets occurred after approximately an hour of substorm activity. Chemical release rockets were launched at 1308 UT and 1312 UT and a total of four TMA trails were deployed on the upleg and downleg portions of the two rocket flights.

The resulting horizontal neutral wind profile from the ARIA II experiment is shown in Figure 4.5 (Figure 3 in Larsen et al., 1997). A jet-like feature is evident in the wind profile in the lower E-region from 95 km to 105 km. One wind peak near 105 km is about 180 m s$^{-1}$ northeastward, and another near 115 km has a magnitude greater than 250 m s$^{-1}$. The wind shear below the 115 km wind peak was more than 300 m s$^{-1}$ over an altitude range of 5 km. The hodograph corresponding to the wind profile is shown in Figure 4.6 (Figure 4 in Larsen et al., 1997). The tip of the wind vector rotated clockwise in a elongated circular fashion.
4.1.3 ARIA III (moderate to high magnetic activity)

The ARIA III experiment was carried out on February 2, 1995, at Poker Flat Research Range, Alaska. The chemical release rockets were launched at 1600 UT providing neutral wind measurements for the height range between 90 and 140 km. The Kp index was nearly 4 all during the day of the launch. The activity level was between that of ARIA I and ARIA II. The magnetic deflection was nearly 900 nT for the H-component just prior to the launch time, as shown in Figure 4.8 for the measurements from Eagle station. The location of Eagle station is listed in Table 4.1 and the spacing between grid lines in the Figure 4.8 is 500 nT.

The horizontal neutral wind profile is plotted in Figure 4.9 and the corresponding hodograph in Figure 4.10 (Figure 7 and Figure 8 in Larsen et al., 1997). A jet feature is again evident in the lower E-region in the altitude range between 100 and 110 km. One wind peak near 104 km was northeastward with a magnitude of 120 m s$^{-1}$. Another wind peak near 110 km was southeastward with a magnitude close to 180 m s$^{-1}$. At the bottom
side of the 110 km wind peak between 105 and 110 km, the large wind shear exceeded 230 m s\(^{-1}\) over 2.5 km. The hodograph corresponding to the wind profile is a significantly elongated circular shape and the wind vector rotated clockwise with increasing altitude.

4.1.4 ARIA IV (low magnetic activity)

The ARIA IV experiment was carried out on November 24, 1995, at Poker Flat Research Range, Alaska. The launch time for the chemical release rocket was 1407 UT under quiet magnetic conditions. The local launch time was within 2 hours of the local time of the launches during the ARIA I, II, and III experiments under disturbed condition. The Kp Index was less than 2 for all of the day on November 24, as shown in Figure 4.11. The measurements of the magnetometer at Gakona Station, shown in Figure 4.12, gave the magnetic deflection close to zero all through the day. No substorm event occured during the
launch night and the magnetic conditions had been quiet for several days prior to November 24.

The horizontal neutral wind profile is shown in Figure 4.13 and the corresponding hodograph in Figure 4.14 (Figure 1 and Figure 2 in Larsen et al., 1997). The neutral wind velocity oscillated as the height increased. The wind peaked near 105 km with the velocity magnitude less than 130 m s\(^{-1}\) and northeastward. The wind direction changed to southwestward between 110 to 120 km altitude with magnitude less than 100 m s\(^{-1}\). The largest shear observed was within the height range between 105 and 100 km but much smaller than the shears observed during disturbed conditions in ARIA I, II, and III. The corresponding hodograph shows that the wind vector tip rotated clockwise in a nearly circular shape as the altitude increased.

### 4.2 CODA 2

In the CODA 2 experiment, three chemical release and one instrumented rocket were launched from Poker Flat Research Range in Alaska on February 21, 2002. The first chemical release rocket was launched along a trajectory close to magnetic north at 09:57 UT with a geographic azimuth of 17.0°. The second launch was along an azimuth further to the west at 09:59 UT with a geographic azimuth of 346.0°. The third launch at 10:01 UT was in the northeast direction with an azimuth of 72.6°. Each rocket released a pair of TMA trails in puffs on both the upleg and downleg sections of the trajectories, beginning
around 74 seconds after each launch. The TMA trails released on the upleg lasted for \( \sim 4 \) minutes, covering the height interval 86–139 km. The TMA trails released on the downleg were visible for \( \sim 10 \) minutes, covering the height interval 84–144 km. The locations of the three pairs of TMA trails are indicated as line segments in Figure 4.15.

On the day of the CODA 2 launch, the geomagnetic conditions were quiet to moderate. The Kp value was less than 2 during most of the day and was 2+ right before the launch time, as shown in Figure 4.16. The ground-based magnetometer chain in Alaska provided real-time measurements of the local magnetic field perturbations. The locations of selected stations are listed in Table 4.1 which, from north to south, include Kaktovik, Arctic Village, Bettles, Fort Yukon, Poker Flat, College, Eagle, Gakona and Talkeetna. The data from the magnetometer chain not only support the launch decision, but also are used for the post-launch analysis, such as being part of the input for the AMIE analysis procedure [see Anderson et al., 1995 and Richmond and Kamide, 1988], which is used to specify the Joule heating and Lorentz forcing in the high-latitude region.
Figure 4.9: Wind profile observed during the ARIA III experiment on February 2, 1995. After Larsen et al., [1997].

Figure 4.10: Hodograph corresponding to the wind profile observed during the ARIA III experiment on February 2, 1995. After Larsen et al., [1997].
Figure 4.11: Kp index during the ARIA IV experiment on November 24, 1995.

Figure 4.12: Magnetometer measurements at Gakona station during the ARIA IV experiment on November 24, 1995. The vertical grid spacing is 250 nT and the value on the right of each curve is the reference value. The H component is positive to the north, the D component is positive to the east, and the Z component is positive for vertical down.
Figure 4.13: Wind profile observed during the ARIA IV experiment on November 24, 1995. After Larsen et al., [1997].

Figure 4.14: Hodograph corresponding to the wind profile observed during the ARIA IV experiment on November 24, 1995. After Larsen et al., [1997].
Figure 4.15: CODA sites map
<table>
<thead>
<tr>
<th>Location</th>
<th>Longitude</th>
<th>Latitude</th>
</tr>
</thead>
<tbody>
<tr>
<td>Kaktovik</td>
<td>216.35°</td>
<td>70.14°</td>
</tr>
<tr>
<td>Arctic Village</td>
<td>212.14°</td>
<td>68.12°</td>
</tr>
<tr>
<td>Bettles</td>
<td>208.45°</td>
<td>66.90°</td>
</tr>
<tr>
<td>Fort Yukon</td>
<td>214.69°</td>
<td>66.56°</td>
</tr>
<tr>
<td>Poker Flat</td>
<td>212.54°</td>
<td>65.11°</td>
</tr>
<tr>
<td>College</td>
<td>212.14°</td>
<td>64.87°</td>
</tr>
<tr>
<td>Eagle</td>
<td>218.84°</td>
<td>64.78°</td>
</tr>
<tr>
<td>Gakona</td>
<td>214.87°</td>
<td>62.39°</td>
</tr>
</tbody>
</table>

Table 4.1: Magnetometer Chain in Alaska

As indicated by the Alaska magnetometer chain measurements, the CODA 2 experiment followed a long period of quiet conditions prior to the launch time of the first rocket at 0957 UT. The magnetic disturbance was nearly zero prior to 0600 UT. Between 0600 UT and 1000 UT the H component remained small with magnitudes less than \(\sim 120\) nT at all stations. The positive deflection of the H component at 0600 UT to 1000 UT indicates an eastward auroral electrojet current. The Z component had negative peak deflections near 0900 UT at Fort Yukon and Arctic Village, with a magnitude up to \(\sim 170\) nT, which indicates an auroral electrojet south of that location. The magnetometers at College and Gakona showed small positive deflections in the Z component up to \(\sim 50\) nT during the period from 0600 UT to 1000 UT, while the measurements at Poker Flat were close to 0 nT. The Z component measurements imply that the auroral electrojet current was located between Fort Yukon and College and near Poker Flat.

The neutral horizontal wind velocity was obtained from the observations of the motion of the three pairs of TMA chemical release trails. Photographs of the TMA trails were taken from the Poker Flat Research Range, Cold Foot, and Fort Yukon, all located in Alaska. All the trails, except the downleg from the second rocket, were covered by photos from all three sites. The location of the sites and the trajectories of the rockets are shown in Figure 4.15. Each site was equipped with two Hasselblad cameras with 80 mm focal length lenses, two Nikon film cameras with 50 mm lenses and an intensified CCD camera.

The derived wind profiles from the CODA 2 experiment are shown in Figures 4.1837
to 4.22. The Hasselblad cameras only covered the TMA trails released on the downleg portion of the 1st and 3rd rocket trajectories, while the Nikon cameras covered all trails except the one released on the downleg portion of the second rocket’s flight. The analysis was mostly based on the results from the photos taken by the Nikon cameras, in spite of the less accurate calibration for the Nikon camera in the triangulation program. Figures 4.18 to 4.20 show the zonal and meridional components of the neutral wind obtained from the three upleg TMA trails. The three profiles give similar results and are nearly identical below 120-km altitude. There was one wind peak near 97 km, with magnitude greater than \(~90\ \text{m s}^{-1}\) in the southwestward direction, and another peak at 103 km, with magnitude greater than \(~120\ \text{m s}^{-1}\) in the northeastward direction. Above 115 km, the wind was northwestward and increased with height, becoming more uniform above 120 km. Figure 4.21 and Figure 4.22 give the zonal and meridional wind velocity profiles obtained from the center and east downleg TMA trails. The wind profiles are similar to those derived from
Figure 4.17: CODA 2 Alaska magnetometer chain data.
upleg TMA trails. The wind peak at 97 km was more than \( \sim 90 \, \text{m s}^{-1} \) toward southwest in the results from downleg trails. The wind peaked at 105 km with magnitude \( \sim 135 \, \text{m s}^{-1} \) in the center downleg and was at 107 km with magnitude greater than \( \sim 90 \, \text{m s}^{-1} \) from the east downleg trail, both toward northeast. At the bottom side of the wind peak, large wind shears were found to be greater than \( 80 \, \text{m s}^{-1} \, \text{km}^{-1} \) at the altitude near 100 km.

Figure 4.18: CODA 2 wind velocity profile obtained from the TMA trail released on the center upleg part of the trajectory. The solid (dashed) line represents the zonal (meridional) wind component. The error bars show plus or minus the standard deviation.

Figure 4.23 and Figure 4.24 show the hodographs corresponding to the wind profiles obtained from the upleg and downleg TMA trails, respectively, in CODA 2. The tip of the wind vector in CODA 2 rotates clockwise nearly circularly as the height increases, which indicates the dominance of tidal forcing during the CODA 2 launches.
Figure 4.19: CODA 2 wind velocity profile obtained from the TMA trail released on the west upleg. The solid (dashed) line shows the zonal (meridional) component.
Figure 4.20: CODA 2 wind velocity profile obtained from the TMA trail released on the east upleg, the solid (dashed) line shows the zonal (meridional) component.

4.3 HEX

The Horizontal E-Region Experiment (HEX) was carried out at Poker Flat Research Range, Alaska, on March 25, 2003. The HEX experiment was designed to measure the vertical neutral wind near a stable pre-midnight auroral arc, and the results were analyzed by Wescott et al. [2006]. The magnetic activity conditions were quiet at the time of the launch, with the Kp value less than 2 for the whole day of March 25, as shown in Figure 4.25. Figure 4.26 shows the geomagnetic field for the launch day measured by six Alaska magnetometer stations, listed from North (Top) to South (Bottom). The magnetic deflection was small at most southern stations and was mostly smaller than 50 nT at Arctic Village and Fort Yukon prior to the launch time at 0950 UT. The all-sky camera images at the time of the launch show two weak stable arcs located near Arctic Village at the time of the launch.
Figure 4.21: CODA 2 wind velocity profile obtained from the TMA trail released on the center downleg. The solid (dashed) line represents the zonal (meridional) component. The black (red) curve is for the result based on the photographs taken with the Nikon (Hasselblad) cameras.
Figure 4.22: CODA 2 wind velocity profile obtained from the TMA trail released on the east downleg. The solid (dashed) line shows the zonal (meridional) component.
Two chemical release sounding rockets were launched during the HEX experiment [Wescott et al., 2006]. The first rocket was a Black Brant X, launched at 0950 UT on March 25, 2003. It travelled and deployed TMA trail along a nearly horizontal trajectory across the auroral arcs within the altitude range from 137 to 155 km. The second rocket was a Terrier Orion, launched at 1009 UT, following a normal parabolic trajectory. TMA trails were released nearly vertically during both upleg and downleg portions of that flight. The horizontal neutral winds were derived by the triangulation technique from the photographs of the TMA trails taken at Poker Flat, Arctic Village, Toolik Lake and Old Crow. The wind
Figure 4.24: CODA 2 hodograph for all wind profiles obtained from TMA trails released on the downleg portion of the trajectories.

Figure 4.25: HEX experiment Kp index.
Figure 4.26: HEX experiment Alaska magnetometer chain data.
is shown in Figure 4.27 (Figure 7 in Wescott et al., 2006). The horizontal wind was mostly northwestward and the magnitude increased with altitude. The largest wind observed from the horizontal TMA trail was nearly 250 m s$^{-1}$ from the southern end of the trail at about 145 km. The wind observed from the two vertical TMA trails show that the wind turned toward southwest at altitude near 120 km, but with much smaller magnitude.

4.4 JOULE I

The JOULE I Experiment was carried out at the Poker Flat Research Range on March 27, 2002. Two pairs of rockets were launched from Poker Flat Research Range, with two instrumented rockets each followed by a chemical tracer rocket. The two chemical tracer Terrier-Orion rockets were launched at 12:00UT and 12:13UT. Unfortunately the first rocket failed because of a problem with second stage separation, and thus no chemical tracer was released by this rocket. Trimethyl aluminum (TMA) trails were released successfully around 1 minute after the launch of the second chemical tracer rocket, both on the upleg and downleg portions of the trajectory. The TMA trails released on the upleg lasted for $\sim$5 minutes, covering the height interval 87–169 km. The TMA trail released on the downleg was visible for $\sim$20 minutes, covering the height interval from 87–172 km. The TMA chemical was released in puffs, with a duration of 1.0 s and 1.0 s delay between puffs. The locations of TMA trails are shown as line segments in the Figure 4.28. The pair of TMA trails released by the second chemical tracer rocket is on the left.

The JOULE I launches occurred after many hours of active conditions. The Kp index value on the launch day were mostly greater than 3 and up to 5-. Prior to the launch time of 1200 UT, the ground magnetometers detected large H components up to $\sim$500 nT in magnitude. The H component started to fluctuate and a negative bay started to develop after 0800 UT, with the largest deflection nearly $\sim$500 nT at Eagle station. The bay was larger and more continuous at southern stations than at northern stations. This indicates a strong westward auroral electrojet current close to the stations at CIGO and
Figure 4.27: HEX experiment results. After Wescott et al. [2006]
Figure 4.28: JOULE I sites map
Gakona. From 0800UT to 1200UT, the Z component fluctuated and increased, and the increase was larger at northern stations. Following the launch, there was a larger substorm, i.e., a large negative bay in the H component from 1400UT to 1800UT, which indicated a stronger westward auroral electrojet. The Z component was always larger than the quiet level at Kaktovik and Arctic Village, which means that the auroral eletrojet was located south of those stations. The negative deflection at Gakona indicates that the electrojet remained north of that station. The reversal in the direction of the deflections at CIGO, Poker Flat, Eagle, Fort Yukon, and Bettles implies that the electrojet moved from south to north, passing those stations.

The horizontal neutral wind velocity was obtained from the observations of the motion of the TMA chemical releases. Photographs of the TMA trails were taken from Poker Flat Research Range, Coldfoot, and Toolik Lake, all located in Alaska. The two trails were covered by photographs from all three sites. The location of the sites and
Figure 4.30: JOULE I data from the Alaska magnetometer chain.
the trajectories of the rockets are shown in Figure 4.28. Each site had two Hasselblad cameras with 80-mm lenses, two Nikon film cameras with 50-mm lenses and one intensified CCD video camera. The Hasselblad cameras took photographs every 20 seconds and the Nikon cameras every 4 seconds. Again, the triangulation technique was used, based on the photographs taken at the three observation sites, to calculate the neutral wind. The typical uncertainty is 5–10 m s\(^{-1}\).

The TMA trail’s proximity to the observation site at Poker Flat Research Range required camera pointing elevation angles as large as \(\sim 80^\circ\). The TMA trails released on the upleg in the photographs taken at Poker Flat therefore had a very complicated shape, with the top of the trail mostly obscured by the bottom part of the trail. The top of the TMA trail disappeared very quickly, which made it even more difficult to determine the spatial position of the top using triangulation, so that it was difficult to obtain credible results for the horizontal velocities above \(\sim 120\) km from the upleg TMA trail.

The resulting horizontal neutral wind velocity profiles are shown in Figures 4.31 to 4.32. Figure 4.31 gives the zonal and meridional wind velocity obtained from the TMA trail released on the upleg by triangulation based on the photographs taken with the Hasselblad cameras. The zonal and meridional velocities are plotted as a function of height within the interval from 95 to 119 km altitude. The neutral wind had one peak at 98 km with magnitude greater than 110 m s\(^{-1}\) northeastward. The wind at 119 km was as large as 200 m s\(^{-1}\) southwestward. Figure 4.32 shows the zonal and meridional wind velocities for the altitude range from 94 to 159 km obtained from the TMA trail released on the downleg. Triangulation was based on the photographs taken by both Hasselblad and Nikon cameras, which is shown as black and red curves, respectively, in the figure. The wind profiles from two different cameras have consistent results below \(\sim 115\) km but deviate above that altitude, because of the less accurate calibration for the Nikon camera in the triangulation program. The wind also peaked at 98 km with a magnitude around 100 m s\(^{-1}\) northeastward, and another peak at 120 km with a magnitude close to 200 m s\(^{-1}\) southwestward. Large wind shear found at the bottomside of the 120 km wind peak with a magnitude greater than
Figure 4.31: JOULE I wind velocity profile obtained from the TMA trail released on the upleg part of the rocket trajectory. The zonal (meridional) component is shown as a solid (dashed) line. The error bars show plus or minus the standard deviation.

60 m s\(^{-1}\) km\(^{-1}\)

Hodographs corresponding to the wind profiles in the JOULE I experiment are plotted in order to illustrate the variation of the direction and magnitude of the wind as a function of height. Figure 4.33 and Figure 4.34 show the hodographs corresponding to the wind profiles obtained from the upleg and downleg TMA trails, respectively. The shape of the hodographs is complicated, but the tip of the wind vector generally follows an elongated clockwise rotational curve as height increases.

### 4.5 JOULE II

The JOULE II sounding rocket experiment was carried out at the Poker Flat Research Range on January 19, 2007. The location of the launch site is shown in Figure 4.35. Two pairs of rockets were launched from Poker Flat Research Range, with two instrumented
Figure 4.32: JOULE I wind velocity profile obtained from the TMA trail released on the downleg. The zonal (meridional) component is shown as a solid (dashed) line. The black (red) curve is for the result based on the photographs taken by the Hasselblad (Nikon) cameras.
Figure 4.33: JOULE I wind velocity profile obtained from the TMA trail released on the downleg. The zonal (meridional) component is shown as a solid (dashed) line. The black (red) curve is for the result based on the photos taken by Hasselblad (Nikon) cameras.
Figure 4.34: JOULE I wind velocity profile obtained from the TMA trail released on the downleg. The zonal (meridional) component is shown as a solid (dashed) line. The black (red) curve is for the result based on the photos taken by Hasselblad (Nikon) cameras.
rockets each followed by a chemical tracer rocket. All four rockets were designed to have the same trajectory, in order to investigate the temporal changes of the electrodynamics and dynamics of the E-region and lower F-region of the ionosphere.

During the launch day of Jan 19, 2007, the geomagnetic conditions were moderately active. The Kp index was slightly greater than 4 during the period from 0300 UT to 0630 UT and was between 2 and 4 during the rest of the time prior to the launch. Although it was moderately active in a global average sense, a magnetic substorm affected the area around the launch site, as shown by the magnetometers. The magnetometer readings from nine stations in the Alaskan magnetometer chain are shown in Figure 4.37. The station names are listed at the left side of the plot and are sorted from north to south. A negative bay was seen in the H-component from 09 - 16 UT at most northern stations, with largest disturbance of nearly -500 nT at Poker Flat. The D-component did not show large disturbances except for a negative spike at around 10 UT detected at most of the northern stations. Meanwhile, a positive bay was observed in the Z-component at northern stations with the largest disturbance up to ∼250 nT. The magnetometer data indicated that there was a strong southward current in the ionosphere near Poker Flat.

The two chemical tracer rockets were launched at 1231 UT and 1247 UT. Trimethyl aluminum (TMA) trails were released successfully around 1 minute after the launch of the second chemical tracer rocket, both on the upleg and downleg sections of the trajectory, but the release of the downleg trail failed in the first launch. The TMA trails released on the upleg lasted for ∼5 minutes, while the TMA trail released on the downleg was visible for more than 20 minutes, all covering the height interval from 80–170 km. The positions of the upleg and downleg TMA trails are shown in Figure 4.35. All were located north of the launch site at Poker Flat.

The TMA trails were photographed from the three ground camera stations at Poker Flat, Fort Yukon, and Coldfoot, as shown in Figure 4.35. Each station was equipped with two Hasselblad cameras with 80-mm lenses, two Nikon film cameras with 50-mm lenses and two Nikon digital cameras. The Hasselblad cameras took photographs every 20 seconds and
Figure 4.35: JOULE 2 sites map
the Nikon cameras every 5 seconds. Due to the cloudy weather conditions at Coldfoot at the time of the launch, the TMA trails were only successfully observed by the cameras at Poker Flat and Fort Yukon. The positions of the TMA trails were determined by triangulation using the photographs, and the horizontal neutral wind velocity was calculated. The typical error is $\sim 5-10 \text{ m s}^{-1}$.

The resulting horizontal neutral wind velocity profiles are shown in Figures 4.38 to 4.41. Figure 4.38 gives the zonal and meridional wind velocity obtained from the TMA trail released on the upleg of the first launch by triangulation based on the photographs taken with the Hasselblad cameras. The zonal and meridional velocities are plotted as a function of height within the interval from 94–140 km altitude. The zonal wind is plotted as a solid curve and the meridional wind as a dashed curve. The horizontal error bars indicate the standard deviation of the calculated wind velocities. The zonal wind has wave-like oscillations as the altitude increases and a peak of $100 \text{ m s}^{-1}$ at 100 km. The meridional wind has one peak of $\sim 120 \text{ m s}^{-1}$ at 95 km and another of $-140 \text{ m s}^{-1}$ at 125 km. The wind maximum near 100 km altitude was northeastward with magnitude around $100 \text{ m s}^{-1}$, and the one near 120 km altitude was southeastward with a magnitude greater than $150 \text{ m s}^{-1}$. Figure 4.39 gives the zonal and meridional wind velocity obtained from the TMA trail released on the upleg of the second launch obtained by triangulation based on the photographs taken with the Hasselblad cameras. The wind profile is quite similar to the one obtained from the first upleg release, except for the positive meridional component between 110 km to 115 km.
Figure 4.37: JOULE 2 Alaska magnetometer chain data
The zonal wind has a peak of $\sim 100$ m s$^{-1}$ at 100 km. The meridional wind has one peak of $\sim 120$ m s$^{-1}$ at 95 km and the other of $-140$ m s$^{-1}$ at 125 km. The wind maximum near 100 km altitude was northeastward with magnitude around 100 m s$^{-1}$, and the one near 120 km altitude was southeastward with magnitude greater than 150 m s$^{-1}$. Figure 4.40 gives the zonal and meridional wind velocity obtained from the TMA trail released on the upleg of the first launch by triangulation based on the photographs taken with the Hasselblad cameras. The zonal and meridional velocities are plotted as a function of height within the interval from 90–125 km altitude. The wind profile is similar to those observed from the two upleg releases. The northeast wind peak near $\sim 100$ km altitude was 80 m s$^{-1}$, and the southeast wind maximum near $\sim 120$ km was greater than 150 m s$^{-1}$. Since the upper part of the downleg release dissipated very quickly, the Hasselblad cameras were not able to provide enough photographs for triangulation. The Nikon cameras took photographs more frequently and thus those photographs of the TMA trail released on the downleg part
Figure 4.39: JOULE 2 wind profile derived from the second upleg TMA trail. The solid line is the zonal wind, positive to the east. The dashed line is the meridional wind, positive to the north.
Figure 4.40: JOULE 2 wind profile derived from the second downleg TMA trail using the photos taken with the Hasselblad cameras. The solid line is the zonal wind, positive to the east. The dashed line is the meridional wind, positive to the north.
Figure 4.41: JOULE 2 wind profile derived from the second downleg TMA trail using the photos taken with the Nikon cameras. The solid line is the zonal wind, positive to the east. The dashed line is the meridional wind, positive to the north.
of the second launch were used to obtain the wind profiles above an altitude of 105 km. Although the triangulation program has a less accurate calibration for the Nikon cameras, the resulting wind profile shown in Figure 4.41 shows good consistency with the results from the Hasselblad photographs within the overlapping altitude range, especially the southeast wind peak near 120 km.

The hodographs of the neutral wind profiles are shown in Figure 4.42 to Figure 4.44. The hodograph plots the zonal wind versus the meridional wind, with the zonal wind on the x-axis and the meridional wind on the y-axis. The corresponding altitudes are marked to the right of the points on the curve. The hodograph shows the curve traced out by the tip of the wind vector as altitude increases. The hodographs corresponding to the upleg wind profiles are shown in Figure 4.42 to Figure 4.43. In both hodographs, the tip of the wind vector rotates counter-clockwise in a nearly circular fashion as the altitude increases in the region below 110 km, while the wind vector tip moves in a more linear fashion as the height increases in the region above 110 km. The hodographs corresponding to the wind profiles derived from the TMA trails released on the downleg of the second launch are shown in Figure 4.44. The movement of the tip of the wind vector looks more random as altitude increases.

4.6 Summary of Neutral Wind Observation

The mesosphere and lower thermosphere horizontal neutral wind profiles obtained from a total of eight sounding rockets campaigns have been presented. All the neutral wind velocities were derived by triangulation of TMA trails released at Poker Flat, Alaska.

Large horizontal winds were observed within the altitude range from 100 to 120 km with large wind shears below the wind maximum as a common feature in all the wind profiles presented here. Larsen [2002] has published statistical results for the wind and wind shear maxima for a data set of more than 400 mesosphere and lower thermosphere wind profiles measured with the chemical release technique since the late 1950’s. His analysis included
Figure 4.42: JOULE 2 hodograph of the wind profile derived from the TMA trail released on the upleg of the first launch. The zonal wind component is plotted on the x-axis and the meridional wind is plotted on the y-axis. The corresponding altitudes are marked to the right of the curve. The hodograph shows the curve traced out by the tip of the wind vector as a function of altitude.
Figure 4.43: JOULE 2 hodograph of the wind profile derived from the TMA trail released on the upleg of the second launch.
Figure 4.44: JOULE 2 hodograph of the wind profile derived from the TMA trail released on the downleg of the second launch.
only the wind profiles that were obtained at sites that were equatorward of 60° latitude, i.e., excluding wind profiles that were likely to be influenced by direct auroral forcing. It was found that the wind profiles showed a wind maximum in the altitude range between 100 and 110 km with over 60% wind velocities exceeding 100 m s$^{-1}$. There were associated large wind shears at the bottom side of the wind maximum. The wind profiles presented in this chapter show some of the same qualitative features, i.e., large winds near 100-km altitude and large wind shears near the bottom side of the wind maximum, but the wind profiles are distinctly different from the mid- and low-latitude profiles discussed in Larsen’s [2002] study. For example, in all but the wind profile measurement that was made in quiet conditions, the wind maximum was found to be as large as, and generally exceeded, the largest wind speeds found for the mid- and low-latitude profiles. The location of the wind maximum was also found to vary significantly with activity level. As the strength of the forcing increased, the wind maximum moved to higher altitudes near 115 to 120 km, which is consistent with a response to enhanced Pedersen conductivity and ion drag response to electric field forcing of magnetospheric origin.

To analyze all the wind profiles, we first group them according to the observation time. All the wind measurements during the ARIA series were made during post-midnight time (1308 UT – 1600 UT). Poker Flat was located in the dawn cell of the $F$ region plasma convection pattern. The prevailing plasma drift was eastward and had been accelerating the neutral atmosphere for several hours. Wind profiles from the ARIA I to ARIA IV experiments have consistent features in the $E$ region with respect to the wind direction. A northeastward wind peak in the region from 100 to 110 km altitude and another peak in the southwestward direction in the region between 110 and 120 km. The narrow altitude range between the two wind peaks has large wind shear and are highly unstable with Richardson number near or below the critical value of 0.25. The CODA 2 and HEX 1 experiments were carried out near 1000 UT and Poker Flat was in the dusk cell but near the transition from dusk to dawn in the plasma convection pattern. The neutral winds measured during CODA 2 and HEX 1 show consistent northwestward winds in the region above 120 km.
The chemical release sounding rockets in the JOULE 1 and JOULE 2 experiments were launched soon after local magnetic midnight (1200 UT to 1246 UT). The atmosphere above Poker Flat was experiencing the transition from the dusk to the dawn cell. The prevailing $F$ region plasma flow was anti-sunward, being southward over Poker Flat in meridonal direction but could be varying rapidly in the zonal direction. Comparing the neutral wind measured during the JOULE 1 and JOULE 2 experiments, the meridional wind profiles are very similar to each other with the wind changing from northward below 110 km to southward above 110 km. However, the zonal winds are quite different, probably owing to the $F$ region plasma drift changing rapidly in the zonal direction.

We can also compare the horizontal neutral wind measurements under different magnetic activity levels. The geomagnetic activity was quiet during ARIA IV, moderate during ARIA I, moderate to high during ARIA III and high during ARIA II, according to the Kp index value and magnetometer readings. The wind velocity at the southwestward peak, which was between 110 to 120 km, increases with the geomagnetic activity levels, from $100 \, \text{m s}^{-1}$ during ARIA IV to nearly $250 \, \text{m s}^{-1}$ during ARIA II. Also the wind maximum and the unstable jet tend to be at a higher altitude during the higher magnetic activity conditions than during quiet conditions. The CODA 2 observations were made during quiet to moderate magnetic activity conditions, while JOULE 1 and JOULE 2 were carried out during moderate and moderate to high magnetic activity levels, respectively. The same conclusions can be drawn from the comparison of the wind profiles from the CODA 2 experiment with those from the JOULE 1 and JOULE 2 experiments. The wind velocity at the maximum was smaller during CODA 2 with magnitude of $120 \, \text{m s}^{-1}$ at 115 km altitude, while the wind velocity at the maximum was nearly $200 \, \text{m s}^{-1}$ during JOULE 1 and over $150 \, \text{m s}^{-1}$ during JOULE 2, both at altitudes near 120 km.

As a summary of the wind profile features, we find that all the wind profiles present a wind maximum in the 110–120-km altitude region and another comparatively smaller peak in the 100–110 km altitude region, with an unstable wind shear between the two peaks. The wind profiles obtained within the same time sector (pre-midnight, local magnetic midnight
or post-midnight) have similar features, especially with respect to the flow direction at the wind maximum. The wind velocity tends to be larger during more active geomagnetic conditions than during quiet conditions and the unstable large wind shears were lifted about 5 km higher.
Chapter 5

MODEL SIMULATIONS FOR THE WIND IN MESOSPHERE AND LOWER THERMOSPHERE

The large-scale circulation in the thermosphere at high latitudes has been investigated in many modeling studies [e.g., Roble et al., 1982; Killeen and Roble, 1986; Fuller-Rowell et al., 1996]. In this chapter, modeling studies for the periods corresponding to the sounding rockets experiments discussed in the previous chapter will be presented. The large-scale general circulation models include the NCAR TIME-GCM for the JOULE 1 experiment period, the GITM developed by the University of Michigan for both the JOULE 1 and JOULE 2 periods. A global spectral model, the STGCM [Mikkelsen and Larsen, 1991], is also discussed. Localized models include the one developed by Walterscheid and Lyons [1989] and the one developed at UCLA [Parish et al., 2003], both of which were used to simulate the ARIA I experiment winds.

5.1 The Thermosphere Ionosphere Mesosphere Electrodynamics General Circulation Model

The Thermosphere General Circulation Models (TGCM’s) developed by NCAR are three-dimensional, time-dependent numerical models of the earth’s neutral upper atmosphere. The models use a finite difference technique to obtain a self-consistent solution for the coupled, nonlinear equations of hydrodynamics, thermodynamics, and continuity of the neutral gas and for the coupling between the dynamics and the composition. The original TGCM was developed by Dickinson et al. [1981], and was improved by including coupling of dynamics and composition [Dickinson et al., 1984]. Roble et al. [1988] incorporated a self-consistent scheme for the coupled thermosphere and ionosphere and developed a new
Thermosphere Ionosphere General Circulation Model (TIGCM). Both the TGCM and the TIGCM have spatial resolutions of $5^{\circ}$ in latitude and longitude and 25 constant pressure levels in the vertical direction (from approximately 95 km to 500 km) with a resolution of 0.5 scale height. The time step is 5 minutes. By replacing the empirical models of the electric fields and ion drifts with self-consistent electrodynamic interactions between the thermosphere and ionosphere, Richmond et al. [1992] extended the TIGCM to be a thermosphere-ionosphere-electrodynamics GCM (TIE-GCM).

The most recent model in the series is the Thermosphere Ionosphere Mesosphere Electrodynamics General Circulation Model (TIME-GCM), described by Roble [1993] and Roble and Ridley, 1994. It includes the coupled thermosphere and ionosphere system and the physical and chemical processes appropriate for the mesosphere and upper stratosphere. The physical and chemical processes included in TIME-GCM were described by Roble [1993]. The output consists of as many as 30 fields on a three-dimensional grid, including the neutral temperature and winds, chemical composition, ion and electron temperature, electron density, electrostatic potential and other relevant parameters. The geographic longitude and latitude resolution is 5 degrees, with the range from -180 degrees west to +180 degrees east and from -87.5 south to +87.5 north. The vertical coordinate is expressed in a log pressure scale ($\ln(p_0/p)$) from -17.0 at the bottom to +5.0 at the top (approximately from 30 km to 500 km), with a resolution of 0.5, i.e., half a scale height. The TIME-GCM calculates the global distribution of neutral temperature (K), neutral wind including zonal, meridional and vertical components (m/s), mass mixing ratios of $O_2$, $N_2$, $OX$ ($O+O_3$), $N(^4S)$, $NOZ$ ($NO+NO_2$), $CO$, $CO_2$, water vapor, $H_2$, $HOX$ ($OH+HO_2+H$), O+ ionized atomic oxygen, $CH_4$, $O_2(^1D)$, $NO_2$, $NO$, $O_3$, $O$, $OH$ Hydroxyl, $HO_2$, $H$, $N(^2D)$, ion temperature (K), electron temperature (K) and density, height (km) and the electrostatic potential (volts).

The TIME-GCM simulation was run by Dr. Geoffrey Crowley at Southwest Research Institute (SwRI) for an extended period leading up to and including the time of 12:19:48 UT on March 27, 2003, when the JOULE experiment was carried out. The electric
field was defined as the output of the AMIE procedure with the magnetic disturbance measurements from 155 ground-based magnetometers and the DMSP satellite measurements. The Assimilative Mapping of Ionospheric Electrodynamics (AMIE) procedure, developed at the National Center for Atmospheric Research (NCAR) and described by Richmond and Kamide [1988], is a technique for mapping high-latitude electric fields and currents and their associated magnetic variations in order to produce a consistent picture of the electrodynamics across the polar cap and auroral zone. The output result of the neutral horizontal wind velocity is plotted against altitude within the range from 80 to 180 km in Figures 5.1.

In order to cover the region around N65°, W147° where the TMA appeared, output data at four locations, i.e., the model grid points with latitude N62.5° and N67.5° and longitude W145° and W150°, are shown in the plots.

Neutral zonal and meridional wind velocities are plotted with solid and dashed curves, respectively, in Figure 5.1, with positive values to the east for the zonal component and to the north for the meridional component. Compared to the horizontal wind velocity obtained from the TMA tracer motion, large discrepancies are evident. In the TIME-GCM modeling result, the zonal wind velocity is small below 120-km altitude with a magnitude smaller than ∼20 m/s, and above 120-km, the zonal wind increased toward west as the height increased, reaching a maximum at 160 km with a magnitude of ∼40 m s⁻¹ westward, while the result from the JOULE experiment showed a large westward peak in the zonal wind at ∼120 km with a magnitude of 120–140 m s⁻¹, two eastward peaks below 120 km with magnitudes of 60–97 m s⁻¹, and above 120 km, the zonal wind decreased with increasing altitude, becoming nearly zero at 160 km. The meridional component, according to the simulation, was almost all northward from 80 to 120 km, with a magnitude smaller than 20 m s⁻¹, and increasing in the southward direction above 120 km, having a maximum of 170–200 m s⁻¹ southward at 160 km, while from the JOULE results, the meridional component had a similar profile to the zonal component, having a large southward peak near 120 km with a magnitude of 150–160 m s⁻¹, a northward peak of 60–97 m s⁻¹ below 120 km, decreasing above 120 km altitude, and was ∼60 m s⁻¹ at 160 km. The magnitudes
of the wind velocities given by the TIME-GCM simulation are much smaller than those from the experimental results, except for the meridional component above \( \sim 130 \) km. Also, the large wind shears due to the auroral forcing in the altitude range from 97.5–112.5 km and the instabilities with Richardson numbers less than 0.25 occurred near 100 km and 112.5 km in the experimental result but did not appear in the model simulations.

Possible reasons for the large discrepancies between the experimental results and the model simulations are the small characteristic vertical scale of the jet feature and the instability. The vertical scale of the jet is less than \( \sim 10 \) km, which approaches the resolution limit of the TIME-GCM model. The instability and turbulence caused by the unstable shears are not incorporated in the model. By improving the vertical spatial resolution and using an empirical scheme for the instability, a more realistic simulation result should be expected. Another difference of the meridional wind velocity above 120 km could be explained by the smaller electron density observed by the in situ instrument on the sounding rocket in the JOULE experiment than the TIME-GCM output, as shown in Figure 5.2 and Figure 5.3. The Pedersen conductivity was dominant at heights greater than 120 km, and thus the Pedersen drag forcing. The larger electron density given by the model simulation implies a shorter time constant for the Pedersen drag forcing in the model. Prior to the launch time of the JOULE rockets, the plasma drift in the vicinity of the Poker Flat range experienced the shift from the dusk cell to the dawn cell, therefore having large southward drift velocities. The large southward meridional wind derived from the model was due to the ion drag from the fast southward plasma drift. Since the plasma convection pattern changed rapidly, a more detailed analysis should consult the plasma convection pattern within one hour prior to the launch (1100 UT–1200 UT), and discover the reason for the difference in the electron density profile between the experiment and the model results.
Figure 5.1: TIME-GCM simulation output for the JOULE I experiment. The solid (dashed) line shows the zonal (meridional) wind.
Figure 5.2: TIME-GCM simulation output of the electron density profile for the JOULE I experiment launch time.
Figure 5.3: Electron density profile measured by the instrumented rocket during the JOULE I experiment.
The global ionosphere-thermosphere model (GITM) was recently developed at the University of Michigan [Ridley et al., 2006]. GITM is a three-dimensional general circulation model. The major difference between the GITM and other general circulation models, such as TIME-GCM, is the use of an altitude grid instead of a pressure grid for the vertical coordinate. The altitude spacing is done automatically using scale-heights and specifying a lower and upper boundary and the number of grid points. In addition, GITM uses a three-dimensional spherical grid that can be stretched in both latitude and altitude while having a fixed resolution in longitude, which makes the resolution extremely flexible. The model is fully parallelized using a block-based two-dimensional domain decomposition with latitudinal and longitudinal host-cells bordering the blocks [Oehmke and Stout, 2001]. It has been run on up to 80 processors with a resolution as fine as 1.25° latitude by 5° longitude uniform over the globe and 40 vertical grids. The model can use a dipole or the IGRF magnetic field with the APEX coordinate system [Richmond, 1995] and can be coupled to a large number of models of the high-latitude ionospheric electrodynamics such as AMIE. The outputs of the GITM include the global distribution of neural temperature (K), neutral wind (m/s), ion temperature (K), plasma drift velocity (m/s), electrostatic potential (volts), electron density (cm$^{-3}$), heating rate (K/s), neutral densities of O, O$_2$, N($^2$D), N($^2$P), N($^4$S), N$_2$, NO, H and He and ion densities of O$^+$($^4$S), O$^+$($^2$D), O$^+$($^2$P), O$_2^+$, N$^+$, N$_2^+$, NO$^+$, H$^+$ and He$^+$.

GITM was run for the period of the JOULE 1 and JOULE 2 experiments. Because the JOULE 2 experiment was just carried out in January 2007, there are not enough measurements at this time to obtain an improved high latitude electrostatic potential pattern from the AMIE procedure, therefore, Weimer’s empirical model [Weimer, 1995] was used to provide the background electrostatic potential pattern. The resulting horizontal neutral winds from the simulation are plotted in Figure 5.4. The red solid (dash) curve represents the zonal (meridional) wind velocity using uniform horizontal grids in the GITM model,
Figure 5.4: GITM simulation output for the JOULE 2 experiment. The solid (dashed) line shows the zonal (meridional) wind. The red solid (dashed) curve represents the zonal (meridional) wind velocity using a uniform horizontal grid in the GITM model, the blue solid (dashed) curve represents the zonal (meridional) wind velocity using stretched horizontal grids, and the black solid (dashed) curve represents the zonal (meridional) wind velocity obtained from the TMA released during the upleg of the second chemical tracer rocket.
the blue solid (dash) curve represents the zonal (meridional) wind velocity using stretched horizontal grids, and the black solid (dash) curve represents the zonal (meridional) wind velocity obtained from the TMA released during the upleg of the second chemical tracer rocket. The consistency is poor between the observed and model simulated neutral winds. GITM predicts southwestward neutral winds, with a meridional component that increases in magnitude with altitude and a zonal wind with a peak near 110 km and changing to westward above 150 km. The oscillations in both the observed zonal and meridional wind velocities were not shown by the model. The lower boundary at 100 km certainly limits the ability of the model to produce realistic neutral winds and not including the tidal modes propagating upward from the lower atmosphere makes it more difficult for the model to reproduce the observed neutral winds near the model’s lower boundary. Another possible reason for the discrepancies is the electron density profile used in the model run, which has larger electron density above 120 km compared to the rocket-borne instrument measurements, as shown in Figure 5.6 and Figure 5.7, and thus overestimates the forcing due to the Pedersen drag, leading to a much larger meridional wind component compared to the observations. The JOULE experiments were designed to investigate the JOULE heating rate as a function of horizontal scale size. In particular, the observational results suggest that the Joule heating may be significantly underestimated when the heating rate is calculated using the large-scale electric fields and the smaller-scale electric field fluctuations are ignored. The heating rate obtained by the GITM, shown in Figure 5.5, does give a larger rate when the stretched grid is used, which has finer resolution in the region near Poker Flat. However, JOULE heating is part of the thermodynamic equation, and its effect on the momentum equations is thus implicit. It is therefore difficult to quantify the effects of the increased JOULE heating due to the small-scale fluctuations in the electrodynamic parameter.

The GITM simulation for the JOULE 1 experiment used AMIE output for the background high latitude electrostatic potential pattern. The resulting horizontal wind velocity is shown in Figure 5.8, in which the black solid (dash) curve represents the zonal
(meridional) wind velocity using uniform horizontal grids in the GITM model with the AMIE electrostatic potential, the blue solid (dash) curve represents the zonal (meridional) wind velocity with Weimer’s electrostatic potential and the red solid (dash) curve represents the zonal (meridional) wind velocity obtained from the downleg TMA release of the second rocket flight. The agreement was not improved significantly, although the more realistic electrostatic potential pattern from the AMIE procedure was used. The model simulation shows southwestward neutral winds with magnitude increasing with altitude. The wind direction matches above 110 km, yet the GITM gives a much larger wind magnitude above 120 km. Also, the oscillation of the wind profile below 110 km is not reproduced by the model. In addition to the possible reasons listed above, the AMIE output has a coarser resolution than GITM, especially within the stretched grids in the Poker Flat region, which might underestimate the enhanced electron density and electric fields in the aurora oval, and thus the auroral forcing in the altitude range below 120 km.

5.3 The Spectral Thermospheric General Circulation Model

The spectral thermospheric general circulation model (STGCM) developed at the Danish Meteorological Institute uses an integration scheme in the horizontal plane in the spectral domain rather than the grid point domain used by the NCAR General Circulation Models [e.g., Mikkelsen and Larsen, 1991]. The effect of the tides generated in the lower atmosphere are simulated by forcing the geopotential height at the lower boundary with a semidiurnal period oscillation. The magnetospheric forcing is specified, and in situ solar heating effects are included. The model is discrete in the vertical with 27 layers spaced by half a scale height. With the assumption of hydrostatic balance in the thermosphere, the two horizontal momentum equations, the hydrostatic equation, the mass continuity equation and the thermodynamic equation in isobaric coordinates are solved for the two horizontal wind components, the pressure tendency, the geopotential and the temperature. The fields are expanded in a series of spherical harmonics using a triangular truncation at
wave number 31, equivalent to a homogeneous global resolution with a minimum wavelength of 1270 km. The neutral composition, an initial global reference temperature profile and the electron density are specified externally. The electron density is obtained by combining the Ching-Chui model [Chui, 1975] and an auroral ionization which is computed using the statistical patterns of electron precipitation constructed from the Defense Meteorological Satellite Program (DMSP) observations [Hardy et al., 1987]. And the effect of the upward propagating tides was simulated by imposing a displacement consistent with the (2,2) and (2,4) semidiurnal modes in the geopotential at the lower boundary at 81 km.

The horizontal neutral winds simulated by the STGCM model were compared to $E$ region and lower $F$ region neutral wind observations obtained by rocket-released chemical tracer in the auroral oval in February and March 1978, in March 1985 and in March 1987 [Mikkelsen and Larsen, 1993]. More realistic $E$ region winds and the rotation of the wind vector with altitude can be simulated by the STGCM model, but relative wind maxima from the model appears at higher altitudes than the observed wind maxima in general [Mikkelsen and Larsen, 1993; Larsen et al. 1997]. But the main obstacle to a detailed comparison between model and data is the lack of information about both auroral forcing and the type and amount of energy propagating upward through the lower model boundary. The comparison showed a detailed time history of high-latitude forcing is needed in order to successfully simulate the winds at a given location.

The STGCM model has been run by Dr. Ib Steen Mikkelsen to simulate the neutral winds during the HEX 1 and JOULE 1 period. The simulation results are shown in Figure 5.9 and Figure 5.10. For the HEX 1 period, the STGCM predicted the correct wind direction between 120 and 180 km, but with much smaller wind magnitudes, while for the simulation of the JOULE 1 winds, the STGCM was unable to reproduce the structures in the neutral wind profile observed during the experiment and gave an incorrect meridional wind direction. The STGCM gave comparatively better simulation results for quiet conditions but was not able to reproduce the large neutral winds with unstable shear observed with the TMA trails.
5.4 Other Localized Model

Walterscheid and Lyons [1989] used a two-dimensional time-dependent non-hydrostatic themospheric model to simulate neutral response to the postmidnight diffuse aurora. The model is fully-compressible with the governing equation on a plane-rotating atmosphere. It is a one-gas model but composition dependent quantities such as the gas constant and specific heats vary with altitude. The model uses geometric height coordinates instead of pressure levels, includes the acceleration term in the vertical momentum equation and is not quasi-static. The energy sources are the ion drag momentum and the Joule and particle precipitation heating sources. To simulate the aurora arc, the divergence of the horizontal electric field is negative, $\nabla \times \mathbf{E} < 0$, with the electric field varying in the direction normal to the arc. The model simulation results indicated strong unstable jets in neutral E region zonal wind were possible. Brinkman et al. [1995] also used this model to simulated the E region neutral winds in the postmidnight diffuse aurora during ARIA I campaign. It was the first time that the time history of the detailed specification s of the magnitude and latitudinal variation of the auroral forcing based on measurements have been used and the results were compared to the observations. Parameters includes ground-based airglow measurements used to derived the electron density profile, and the precipitating electron fluxes and electron densities measured on the rocket, electric fields derived from magnetometer and satellite ion drift measurements and the large-scale background neutral wind patterns from a thermospheric general circulation model. The result is shown in Figure 5.11. The model predicted winds with correct sense above 135 km but with overpredicted magnitude of east wind and failed to capture the meridional wind structure and magnitude, while the agreement under 135 km was poor between the model and observed wind because of a large unmodeled jet-like feature in the observed wind.

Parish et al. [2003] used a three-dimensional high-resolution time-dependent model developed at University of California, Los Angeles (UCLA) based on the model developed by Sun et al. [1995] to simulate themospheric response during the ARIA I experiment. The
UCLA model is non-hydrostatic, nonlinear and fully compressible, which allows features with relatively short time and spatial scales. The model treat the neutral atmosphere as a three component gas consisting of \( N_2, O_2 \) and \( O \). And the model solves the coupled nonlinear time-dependent momentum energy and continuity equations in Cartesian frame. The molecular viscosity and thermal conductivity of the mixture of gases determine by functions of temperature and composition given by Banks and Kockarts [1973]. The eddy viscosity and eddy thermal conductivity are estimated according to the relations given by Waltersheid et al. [1985]. The postmidnight auroral forcing was included using the ion drag forcing terms in the momentum equation and the Joule heating terms in the temperature equations. And the 3-D global coupled Thermosphere-Ionosphere Plasmasphere model (CTIP) [Fuller-Rowell et al., 1987] was used to provide tidal variation of large-scale winds and temperatures. The simulation results given by Parish et al. [2003] is much closer to the observations than the results of previous modeling studies, as shown in Figure 5.12 and 5.13. The model successfully simulated the wind maximum large shear in the 110 km to 120 km region with auroral forcing only, and was able to reproduce unstable shears below the peak with Richardson number < 0.25 when sufficiently high vertical resolution is used for auroral forcing parameters. When combined with the tidal forcing, the model predicted the wind with correct sense of vertical structure but underestimated magnitude of the peak at 115 km, Parish et al. [2003] also argued that the jet feature of the horizontal wind peak near 115 km was likely produced by a combination of auroral and tidal forcing processes.
Figure 5.5: GITM simulation output of the Joule heating rate for the JOULE 2 experiment.
Figure 5.6: GITM simulation output of the electron density for the JOULE 2 experiment.
Figure 5.7: Electron density profile derived from the rocket-borne instrument measurements during the JOULE 2 experiment.
Figure 5.8: The GITM simulation output for the JOULE I experiment. The solid (dashed) line represents the zonal (meridional) wind. The black solid (dashed) curve represents the zonal (meridional) wind velocity using uniform horizontal grids in the GITM model with AMIE electrostatic potentials, the blue solid (dashed) curve represents the zonal (meridional) wind velocity with a Weimer electrostatic potential, and the red solid (dashed) curve represents the zonal (meridional) wind velocity obtained from the TMA released on the downleg portion of the rocket flight.
Figure 5.9: STGCM simulation for the HEX 1 experiment. The bold solid blue (green) curve is for the zonal (meridional) wind measured with TMA, and the other blue (green) curves are the STGCM model simulations at the grid points surrounding the location where the TMA was deployed.
Figure 5.10: The STGCM simulation for the JOULE 1 experiment. The bold solid blue (green) curve is for the zonal (meridional) wind measured with TMA, and the other blue (green) curves are the STGCM model simulations at the grid points surrounding the location where the TMA was deployed.
Figure 5.11: Neutral wind simulation result using Walterscheid and Lyons [1989] model for ARIA I period. The left penal is zonal wind component and the right penal is meridional component.

Figure 5.12: Neutral wind simulation (zonal component) result using UCLA model for ARIA I period.
Figure 5.13: Neutral wind simulation (meridional component) result using UCLA model for ARIA I period.
6.1 Neutral Wind at Quiet Condition

In order to study the mesospheric and lower thermospheric neutral winds in the high-latitude region, the winds during quiet conditions should be investigated before looking into the winds under disturbed conditions.

Mikkelsen et al. [1981b] used a two-dimensional numerical model to study the thermospheric neutral winds and found that Lorentz forcing and viscosity provide the dominant effects on the zonal wind component, while the meridional wind component was the result of the interaction of the Lorentz force, Joule heating, and the background pressure gradient force. A theoretical analysis of the high-latitude thermospheric wind patterns by Killeen and Roble [1986] using a Thermospheric General Circulation Model also gave similar results. Their study of different forces at \(E\)-region heights showed the dominance of ion drag and the pressure gradient forces, while the Coriolis force, vertical and horizontal advection terms and viscous forces were less important. Mikkelsen and Larsen [1983] gave an analytic solution for a two-dimensionial shallow-water model for the high-latitude thermosphere and showed that the rotational wind pattern in the \(F\) region follows the plasma convection pattern but will be rotated to the east relative to the plasma convection pattern. The phase difference between the neutral wind and plasma convection depends on the mean electron
The tidal influence on the auroral $E$ region wind under quiet conditions was studied by Brekke et al. [1994] using the neutral wind data measured with the European Incoherent Scatter Radar (EISCAT). The spectral decomposition method was used to investigate the influence of the diurnal, semidiurnal, 8-hour period and 6-hour period tides, separately. Their results showed a strong and persistent eastward wind in the auroral $E$ region below 120 km with largest magnitude of $\sim 60 \, m \, s^{-1}$ in summer, while there was a westward zonal wind above 120 km with largest magnitude of $\sim 70 \, m \, s^{-1}$ in winter. The diurnal tide was found to be dominant in the influence of the horizontal neutral wind in the auroral $E$ region, and the semidiurnal tide was found to have a maximum at 110 km.

During ARIA IV, CODA 2 and the HEX 1 experiments, the magnetic activity could be considered as quiet, with Kp index values less than 2 and the ground-based magnetometer readings less than 200 nT. The hodographs of the horizontal neutral wind profiles obtained during these experiments showed that, below around 120 km, the tip of the wind vector rotated following a nearly circular path as the height increased. The shape of the hodographs is in agreement with the simulation by the STGCM model [Mikkelsen and Larsen, 1991], which analyzed the interaction between the high-latitude $E$ region forcing and the upward propagating tidal modes. The vertical wavelength of the oscillation below 120 km is about 25 km, which can be roughly estimated from the wind profiles and corresponding hodographs. These indicate that the wind is strongly influenced by the tidal modes propagating upward from the lower mesosphere. However, above 120 km, the neutral wind variation became more monotonic and tended to follow the direction of the $F$ region plasma convection in all three experiments. The tides did not penetrate into the higher altitude regions thus must be dominated or attenuated by the auroral forcing effects, even in relatively quiet conditions. The large wind shears at the bottom side of the wind maximum in the layer from 100–110 km may be unstable and thus could produce attenuation of the tides. Above 120 km, viscosity becomes increasingly important and begins to affect the small-scale oscillations. The Pedersen conductivity is dominant in the region above 120 km.
and the Pedersen drag exerted on the neutral atmosphere produces forcing in the direction of the plasma drift.

As a summary, for the high-latitude horizontal neutral wind in the $E$ region and lower $F$ region under quiet condition, the upward propagating tides provided the dominant influence below 120 km, while Lorentz forcing and viscosity were of major importance in the region above 120 km.

### 6.2 Neutral Wind at Disturbed Condition

The period during the ARIA I, ARIA II, ARIA III, JOULE 1 and JOULE 2 experiments can be considered to be disturbed with $K_p$ values exceeding 3 and the ground-based magnetometer readings greater than 500 nT. The mesosphere and lower thermosphere horizontal neutral wind profiles obtained during these experiments show larger wind magnitudes and the unstable wind shear at a higher altitude compared to quiet condition. Their corresponding hodographs showed a more linear or narrow elliptical pattern. Again, the results are consistent with the STGCM results of Mikkelsen and Larsen [1991]. The patterns indicate that the influence of the tidal forcing was interrupted under disturbed conditions by the enhanced Lorentz forcing and Joule heating. The results are in agreement with Nozawa and Brekke [1995], who analyzed the EISCAT wind measurements during disturbed conditions using a method similar to that of Brekke et al. [1994] for quiet conditions, and found the amplitude of the diurnal tide increased above 109 km and the semidiurnal tide was enhanced below 109 km. They also concluded that part of the disturbance in the $E$ region was a result of ion drag, Joule heating and particle precipitation.

Larsen and Walterschid [1995] presented an analysis of the implications of simple balance conditions for the dominant terms in the momentum equation. Unlike the well-known ion drag in the $F$ region associated with the Pedersen conductivity, which is called Pedersen drag, the lower $E$ region heights, where the Hall conductivity dominates, produce a more complicated balanced wind condition due to the Hall drag effects. The Hall drag
term has the effect of decreasing the magnitude of the effective Coriolis parameter and changing the neutral flow configuration.

In order to examine the effects of both the Pedersen drag and Hall drag, the JOULE 2 case was tested using the formulas of modified geostrophy given by Larsen and Walterschid [1995]. Starting from the momentum equation for the thermospheric flow given by

\[
\frac{d\mathbf{U}}{dt} = -2\Omega \times \mathbf{U} - \left(\frac{1}{\rho}\right)\nabla p + \frac{1}{\rho} \mathbf{J} \times \mathbf{B} \tag{6.1}
\]

where \(\mathbf{v}\) is the horizontal flow velocity, \(\mathbf{J}\) is the current density and \(\mathbf{B}\) is the magnetic field. The viscosity term and heating effects are neglected here. Since we are considering region within the auroral oval, the magnetic field can be assumed to be vertical and downward everywhere, and thus \(\mathbf{B} = -B_0 \hat{z}\), where \(\hat{z}\) is the unit vector in the direction of vertical upward. The Lorentz force term can be rewritten as

\[
\frac{1}{\rho} \mathbf{J} \times \mathbf{B} = \frac{1}{\rho} \sigma \cdot \mathbf{E}' \times \mathbf{B} = \frac{1}{\rho} \sigma \cdot (\mathbf{E} + \mathbf{U} \times \mathbf{B}) \times \mathbf{B} \tag{6.2}
\]

where \(\mathbf{E}'\) is the electric field in the neutral frame and \(\sigma\) is the conductivity tensor given by

\[
\sigma = \begin{pmatrix}
\sigma_p & \sigma_h & 0 \\
-\sigma_h & \sigma_p & 0 \\
0 & 0 & \sigma_0
\end{pmatrix} \tag{6.3}
\]

with \(\sigma_p\) is the Pederson conductivity, \(\sigma_h\) is the hall conductivity and \(\sigma_0\) the direct conductivity. Then the Lorentz force can be expanded assuming the plasma drift \(\mathbf{U}_p\) is in the direction of \(\mathbf{E} \times \mathbf{B}\), which is the plasma flow in the \(F\) region.

\[
\mathbf{v}_p = \frac{\mathbf{E} \times \mathbf{B}}{B^2} \tag{6.4}
\]
\[
\frac{1}{\rho} \mathbf{J} \times \mathbf{B} = \left[ \frac{\sigma_h B_0^2}{\rho} (v_p - v) + \frac{\sigma_p B_0^2}{\rho} (u_p - u) \right] \hat{x} + \left[ \frac{\sigma_p B_0^2}{\rho} (v_p - v) - \frac{\sigma_h B_0^2}{\rho} (u_p - u) \right] \hat{y}
\] (6.5)

\(u_p\) and \(v_p\) are the eastward and northward components of the plasma drift, while \(u\) and \(v\) are the eastward and northward components of the neutral flow. \(\hat{x}\) and \(\hat{y}\) are unit vectors in the eastward and northward direction. The terms \(\frac{\sigma_p B_0^2}{\rho}\) and \(\frac{\sigma_h B_0^2}{\rho}\) have units of inverse time and are typically interpreted as collision frequencies. In the \(F\) region only the Pederson drag term is considered and treated as the ion-neutral collision frequency \(\nu_{ni}\) in the Equation 2.1. However, in the lower \(E\) region, we have to consider both Pederson and Hall drag terms. By normalized by the Coriolis parameter \(f = 2\Omega \sin \phi\), we defined the dimensionless coefficients

\[
\alpha = \frac{\sigma_p B_0^2}{\rho f} \quad (6.6)
\]

\[
\beta = \frac{\sigma_h B_0^2}{\rho f} \quad (6.7)
\]

Replacing the pressure gradient terms with geostrophic wind components

\[
u_g = \frac{1}{f \rho} \frac{\partial p}{\partial y} \quad (6.8)
\]

\[
u_g = -\frac{1}{f \rho} \frac{\partial p}{\partial x} \quad (6.9)
\]

we have the two equations under steady state by setting \(du/dt\) and \(dv/dt\) equal to zero

\[-v_g + (1 - \beta)v + \beta v_p + \alpha (u_p - u) = 0 \quad (6.10)\]
\[ u_g - (1 - \beta)u - \beta u_p + \alpha(v_p - v) = 0 \quad (6.11) \]

Solving the two equations above and we have the expression for \( u \) and \( v \) giving by

\[
\begin{align*}
u &= \frac{(1 - \beta)u_g + [\alpha^2 - \beta(1 - \beta)]u_p + \alpha(v_p - v_g)}{\alpha^2 + (1 - \beta)^2} \\
v &= \frac{(1 - \beta)u_g + [\alpha^2 - \beta(1 - \beta)]v_p - \alpha(u_p - u_g)}{\alpha^2 + (1 - \beta)^2}
\end{align*}
\]

The electron density profile used for the analysis is from the measurements by the instrumented rocket flown during the JOULE 2 experiment. The profile versus altitude is shown in Figure 6.1, which was measured during the upleg portion of the rocket flight when the rocket flew though the auroral region. The enhancement of the electron density within the altitude range from 110 km to 130 km occurred within the aura. The neutral atmosphere parameters were obtained from the Mass-Spectrometer-Incoherent-Scatter extended model (MSIS-E-90) [Hedin, 1991], while the magnetic field values are from the International Geomagnetic Reference Field model (IGRF) [Langel, 1992]. Expressions for the conductivity and collision frequencies used here are given by Roble and Ridley [1987]. The profile of Pedersen and Hall conductivities are shown in Figure 6.2. In the region below \( \sim 120 \) km altitude, Hall conductivity was dominant and Pedersen conductivity decreased rapidly, while in the region above \( \sim 120 \) km, Pedersen conductivity was dominant.

As shown in Figure 6.3, the electric field measured during the upleg portion of the rocket flight was mostly southward and the calculated \( \mathbf{E} \times \mathbf{B} \) was mostly eastward, thus here we can just consider a positive \( u_p \) and set \( v_p = 0 \). Just looking at the forcing from ion drag and ignoring the equivalent geostrophic wind by setting \( u_g = v_g = 0 \), the calculated neutral wind is plotted in Figure 6.4. An enhancement of the wind occurs between 110 km to 120 km altitude and a large wind shear is expected below the wind maximum.

The equivalent geostrophic wind was much more difficult to estimate, so in order
Figure 6.1: Electron density profile measured by instrument rocket during JOULE 2 experiment.
Figure 6.2: Pederson and Hall conductivities during JOULE 2 experiment.
Figure 6.3: Electric field measured by instrument rocket and calculated plasma drift during JOULE 2 experiment.
Figure 6.4: Horizontal neutral wind velocity of modified geostrophic wind in units of the nominal plasma drift velocity for the period during JOULE 2 experiment. The equivalent geostrophic wind was wet to zero.
Figure 6.5: Horizontal neutral wind velocity of modified geostrophic wind in units of the nominal plasma drift velocity for the period during JOULE 2 experiment. The ratio between the plasma drift and the equivalent geostrophic wind was set to $u_p/u_g = 3$ in the calculation.
to give a qualitative analysis, we assume an eastward geostrophic wind with positive $u_g$ and $v_g = 0$. The ratio $u_p/u_g$ is chosen to be 3, considering that the largest plasma drift was close to 300 m s$^{-1}$ and the neutral wind at quiet time could also be as large as nearly 100 m s$^{-1}$. The calculated modified geostrophic wind velocity is shown in Figure 6.5. Again, the profile displays an enhancement of the wind velocity between 110 km to 120 km altitude and a large wind shear below this wind maximum. Compared to the neutral wind obtained from the triangulation of the rocket-released TMA trails, shown in Figure 4.39, this wind enhancement due to the modified geostrophic wind balance could qualitative explain the observed wind between 105 km and 125 km altitude. The observed zonal wind had a westward wind peak between 110 km to 115 km and a rapid change to an eastward wind peak at 120 km, while the meridional wind had a wind peak between 120 km and 125 km. The calculated modified geostrophic wind profile had nearly the same shape within this altitude range except that the meridional wind peak is slight lower than the observed wind. The wind direction above 120 km were basically the same as well.

The assumptions used in the calculation may be questionable. First, under disturbed conditions the strong auroral forcing will change the pressure distribution and thus the equivalent geostrophic wind may be quite different from the one used in the calculation. However, the equivalent geostrophic wind is usually much smaller than the plasma drift. Second, the steady state used in the calculation may overestimate the effects of ion drag since the electric field may change rapidly during substorm events and it usually takes many tens of minutes to accelerate the neutral flow due to Lorentz forcing. In spite of these questionable assumptions, the calculated modified geostrophic wind gives qualitative agreement with a number of features of the observed winds, including the large winds observed within the 105 km to 125 km altitude range and the large wind shear below the wind maximum.

As a summary for the high-latitude horizontal neutral wind in the $E$ region and lower $F$ region under disturbed conditions, the wind is a result of complex interplay of upward propagating tidal modes, Lorentz force and Joule heating. Pedersen drag dominates the
region above 125 km, while the enhancement of the wind magnitude and wind shear are primarily due to the effect of the Hall drag in the region from 105 to 125 km where the Hall conductivity is dominant and the interaction with the diurnal and semidiurnal tides is most important.

6.3 From Model Simulation Results

In Chapter 5, the simulation results from the general circulation models and localized models have been presented for the ARIA 1, HEX 1, JOULE 1 and JOULE 2 experiments. The comparison between the model simulations and the observations would help us to understand the forcing of the high latitude neutral wind at $E$ region heights.

One reason that the localized models produced more realistic wind simulations is that the detailed time history of auroral forcing parameters, with sufficient vertical resolution, was used. For the large-scale general circulation models, namely TIME-GCM and GITM, the auroral forcing was specified by the high latitude electrostatic potential pattern produced by the AMIE procedure. Using the AMIE procedure to specify the forcing however did not improve the model simulation results of the $E$ region neutral wind significantly under disturbed conditions, compared to the results obtained with the statistical Weimer electrostatic potential pattern. Neither TIME-GCM nor GITM were able to reproduce the observed winds during the JOULE 1 experiment, using the AMIE electrostatic potential pattern. While comparing the GITM simulations using the AMIE forcing and the Weimer statistical pattern forcing, there was only a small difference in the wind magnitude, but no significant improvement in reproducing the wind variations with height and the large wind near 120 km with unstable shear at the bottom side. This is possibly due to the limited resolution in AMIE which does not have sufficiently high resolution to reveal the enhanced auroral forcing. Also a large portion of the measurements assimilated by AMIE are from ground-based magnetometer measurements of magnetic field disturbance. Yet the inversion method using the magnetometer data only maps the rotational component of the
ionospheric current, but does not include the divergent component. This may misrepresent the electric field patterns when there are strong field-aligned currents under disturbed condition, unless there are also sufficient satellite measurements of the electric or magnetic field assimilated. By improving the AMIE resolution and ability to assimilate more types of measurements, more realistic high latitude electrostatic potential patterns could be expected, and thus better model simulation results.

The resolution of the model itself certainly is another important factor. TIME-GCM has a resolution of 5° × 5° horizontally, which might be too coarse to simulate the small structures observed in the wind profiles. GITM has a resolution of 1.25° in latitude by 5° in longitude, and can be stretched to achieve higher resolution at locations of interest, yet the higher resolution is constrained by the AMIE high latitude electrostatic potential pattern which has coarser resolution. Localized models gave more realistic neutral wind simulations with their high grid resolutions, although the specification at the local boundary might cause other problems. One major advantage of the STGCM as a global circulation model is that the integration in the spectral domain allows the existence of perturbation with much smaller horizontal wavelength, while for the models that use grid domains, perturbations with small wavelengths were attenuated and suppressed to keep the solution stable, resulting in a minimum perturbation wavelength of about 5 grid points, which is usually much larger than the scale of auroral oval.

As discussed in the previous two sections, tidal forcing is dominant for the region below 120 km under quiet conditions but is disrupted under disturbed condition. Thus, including tidal forcing is necessary for the modeling simulation of the neutral winds. GITM failed to reproduce the oscillating wind structure below 120 km, primarily due to the fact that the upward propagating tidal modes were not included at the lower boundary. However, TIME-GCM and STGCM did not successfully reproduce the wind profiles either with tidal forcing included, and even the UCLA localized model simulation had large discrepancies in the meridional component. This implies that the tidal forcing is necessary but probably not crucial for the model simulations of mesosphere and lower thermosphere horizontal neutral
winds at high latitudes, especially under disturbed conditions. Also, even under quiet conditions, such as during the ARIA IV experiment, the observed neutral wind exceeded 100 m s\(^{-1}\), which would be much larger than the prediction of models including tidal forcing, such as the simulation results of STGCM and UCLA’s localized model. This indicates that there may be an amplification effect for the tidal winds or even a physical mechanism that we have not included in the governing equations for the high-latitude mesospheric and lower thermospheric flow. Thus the tidal forcing is necessary for the model simulations of the high-latitude mesosphere and thermosphere neutral wind, but does not play a significant role in producing the large neutral winds and unstable shears observed during the sounding rocket experiments described here.

6.4 Conclusion

In this dissertation high-latitude horizontal neutral wind profiles in the mesosphere and lower thermosphere obtained with TMA trails from Poker Flat, Alaska have been presented. The sounding rocket experiments included the ARIA I to IV, CODA 2, HEX 1, JOULE 1 and JOULE 2 experiments, covering the geomagnetic condition from quiet to highly disturbed and different magnetic local times. Large horizontal neutral winds were observed between 100 to 120 km altitude with wind speeds in the range from 100 to 250 m s\(^{-1}\). The altitude and the magnitude of the wind peaks and the large wind shears at the bottom side of the peaks that were observed show important differences from the statistical study by Larsen [2002] of more than 400 mesosphere and lower thermosphere winds and shears measurements since 1958 covering from low-latitude to mid-latitude regions. This indicates that the large neutral winds and wind shears are features in the lower \(E\) region at Poker Flat that are primarily driven by auroral forcing, although the global features noted in the large statistical study of low- and mid-latitude winds undoubtedly also contribute to the detailed structure observed at the auroral zone site. Specific evidence of this is the ARIA IV quiet-time wind measurement, which showed much smaller winds than in the disturbed
cases with a more regular oscillatory variation with height, although still much larger winds than would be predicted by the various tidal theories, for example. In that case, the auroral forcing was essentially negligible and had been very small for a period of several days prior to the launch. The quiet-time winds were within the bounds that were expected, based on the low- and mid-latitude statistical study, but smaller in magnitude and with a different vertical structure than in the disturbed cases.

The \( E \) region neutral wind profiles observed within the same general local time sector (pre-midnight, magnetic midnight or post-midnight) have similar features with respect to the wind direction. The ARIA I to IV experiments were all carried out during the post-midnight period from 1308 UT – 1600 UT. Poker Flat was located in the dawn cell of the plasma convection pattern and was dominated by an eastward plasma drift in the auroral oval. The four wind profiles observed all have a northeastward wind peaks in the 100 to 110 km altitude range and another peak in the southwestward direction in the altitude range between 110 and 120 km. The wind reversal between the two peaks was accompanied by large wind shears. The CODA 2 and HEX 1 experiments were both carried out closer to magnetic midnight near 1000 UT but at a time when Poker Flat was located in the dusk cell of the plasma convection pattern and had been driven by westward plasma flow for a number of hours prior to the launch. The neutral winds measured during CODA 2 and HEX 1 show consistent northwestward wind in the region above 120 km. The JOULE 1 and JOULE 2 experiments, on the other hand, were carried out near local magnetic midnight (1200 UT to 1246 UT). The ionosphere above Poker Flat was experiencing the transition from the dusk cell to the dawn cell. The \( F \) region plasma flow was anti-sunward between the two cells and thus southward above Poker Flat, but was varying rapidly in the east/west direction. The meridional component of the neutral wind profiles obtained during the JOULE 1 and JOULE 2 experiments were very similar to each other, with the wind changing from northward below 110 km to southward above 110 km. The zonal winds, however, were quite different during the two experiments. These comparisons indicate the influence of the \( F \) region plasma flow through the Lorentz force, i.e. ion drag force.
Comparing the neutral wind profiles under different geomagnetic conditions, the horizontal neutral wind is generally stronger under disturbed conditions compared to the wind under quiet conditions. The ARIA experiments give a useful data set of neutral wind profiles covering geomagnetic condition from quiet to highly active. Categorized according to the Kp index value and ground-based magnetometer readings, the geomagnetic activity was quiet during ARIA IV, moderate during ARIA I, moderate to high during ARIA III and high during ARIA II. The wind velocity maximum in the southwestward direction, which was the higher altitude peak between 110 and 120 km, increases with the geomagnetic activity levels, from 100 m s\(^{-1}\) during ARIA IV to 125 m s\(^{-1}\) during ARIA I, to 180 m s\(^{-1}\) during ARIA III, and to nearly 250 m s\(^{-1}\) during ARIA II. Meanwhile, with the increasing geomagnetic activity level, the wind shear at the bottom side of this wind maximum grew larger and more unstable, with Richardson numbers near or below the critical value of 0.25. The comparison of the neutral wind profiles from CODA 2, JOULE 1 and JOULE 2 give the same result. CODA 2 was carried out during quiet to moderate geomagnetic activity, while JOULE 1 and JOULE 2 were during moderate to high and moderate activity levels, respectively. The peak wind magnitude observed during CODA 2 was 120 m s\(^{-1}\) at 115 km altitude, while the wind velocity at the maximum was nearly 200 m s\(^{-1}\) during JOULE 1 and over 150 m s\(^{-1}\) during JOULE 2, both at altitudes around 120 km. Again, the wind shear right below the wind maximum was larger and more unstable during JOULE 1 and JOULE 2 than during the CODA 2 experiment. Another interesting feature is that the unstable wind shears at the bottom side of the wind maxima associated with the narrow jet features were lifted to a higher altitude under higher geomagnetic activity. The comparison of the wind peaks in the wind profiles obtained during all the experiments presented here indicate the influence of the Lorentz force, which is strengthened under higher geomagnetic activity due to the enhanced ionospheric currents, electron density and energetic particle precipitation.

The comparisons of hodographs corresponding to the horizontal neutral wind profiles observed during the ARIA I to IV, CODA 2, JOULE 1 and JOULE 2 experiments further
support the conclusions drawn above. The hodographs show the variation of the horizontal wind vector with height in the horizontal plane. The hodographs corresponding to the wind profiles obtained during the ARIA IV and CODA 2 experiments under quiet or moderate geomagnetic conditions show nearly elliptical curves in the region below 120 km altitude. The nearly elliptical shape of the hodographs indicates the dominance of the forcing of upward propagating waves and tides. While the hodographs corresponding to the wind profiles of the ARIA I, ARIA II, ARIA III, JOULE 1 and JOULE 2 experiments, which were characterized by more disturbed conditions, showed a more elongated or nearly linear hodograph shape. In these cases, the auroral forcing dominated the tidal forcing.

Modeling studies for the ARIA I, JOULE 1 and JOULE 2 periods have also been presented. Large scale general circulation models, such as NCAR’s TIME-GCM and the University of Michigan’s GITM have been used to run the simulation for the period during both JOULE 1 and JOULE 2. The AMIE procedure provided high latitude electrostatic potential patterns as input for TIME-GCM and GITM, based on the assimilated data from the Alaska and Greenland magnetometer chains and DMSP satellite electric field measurements for the JOULE 1 experiment, but only with Weimer’s statistical electric potential pattern and limited magnetometer measurements for the JOULE 2 experiment. The simulation results for the E region neutral wind profiles from TIME-GCM and GITM both fail to reproduce the structures observed with the TMA trails. The simulated wind profiles do not have the wind peak with large wind velocity at around 120 km with unstable wind shears below the wind maxima observed during the JOULE 1 and JOULE 2 experiments. Generally the simulated wind magnitudes are much smaller than the observed winds, although the simulation with TIME-GCM for JOULE 1 did show some oscillations. The GITM simulations were not successful either, with its major problem being the lack of tidal modes propagating upward from the lower atmosphere at the lower boundary of the model at 100 km. The simulations were also limited to the grid resolution of the TIME-GCM and GITM models themselves and the resolution of the input high-latitude electric fields from the AMIE procedure. The TIME-GCM runs for JOULE 1 and JOULE 2 were made with
a resolution of 5° by 5°. Although the GITM model can stretch the grid to produce local high resolution grids over Poker Flat, Alaska, the AMIE input of the electrostatic potential pattern with coarser resolution averaged out the locally intensified electrodynamic features related to the auroral substorms. Differences also exist between the electron density profiles produced by the TIME-GCM and GITM models and measured in situ with the sounding rocket instruments. The comparison of the electron density profiles during the JOULE 1 experiment produced by TIME-GCM and measured by the rockets indicates an overestimate of the electron density, especially in the region above 120 km, which would lead to a larger effect from Pedersen drag in the model simulation. The result was larger winds in the simulation in the direction of the plasma flow. The comparison of the electron density profiles during the JOULE 2 experiment produced by GITM and measured by rocket-borne instruments give the same conclusion.

The spectral thermospheric general circulation model (STGCM) developed at the Danish Meteorological Institute was used to study the effects of the tides propagating upward from the lower atmosphere in the high-latitude region, using integration in the horizontal plane in the spectral domain [Mikkelsen and Larsen, 1991]. Although there were no neutral wind profiles generated by STGCM for the period of the sounding rockets experiments presented in this dissertation available for comparison, the shape of the hodographs corresponding to the neutral wind profiles observed during the experiments agree with the general simulation results of the STGCM. This supports the idea that the tidal forcing plays a significant role in driving the horizontal neutral winds in the E region both in quiet and disturbed geomagnetic condition. More realistic results were obtained from localized non-hydrostatic models with both tides and auroral forcing, simulated for the ARIA I experiments [Brinkman et al., 1995; Parish et al., 2003]. The time history of the detailed specification of the magnitude and latitudinal variation of the auroral forcing based on measurements were used in both Brinkman’s model and Parish’s UCLA model. Parish’s model even successfully reproduced the large wind with unstable shear in the 110 to 120 km region, especially when sufficiently high vertical resolution was used for the auroral forcing.
parameters. Including the non-hydrostatic terms in the model may be another important factor in explaining the improved results of the local models, as compared to the simulations by the hydrostatic global circulation models.

Spectral decomposition was used by Brekke et al. [1994] to study the contribution of the various tidal modes to the mesospheric and lower thermospheric neutral winds at high latitudes, using the measurements from the European Incoherent Scatter Radar (EISCAT) located in Northern Scandinavia. They found that the diurnal tide is dominant in the influence on the horizontal neutral wind in the auroral E region, and that the semidiurnal tide has a maximum at 110 km under quiet geomagnetic condition. Nozawa and Brekke [1995] analyzed the EISCAT wind measurements during disturbed conditions using a similar technique and found that the amplitude of the diurnal tide increased above 109 km and that the semidiurnal tide was enhanced below 109 km, with the conclusion that part of the disturbance in the E region was a result of ion drag, Joule heating and particle precipitation. A further study was conducted for the period of the JOULE 2 experiment under disturbed conditions using the modified geostrophy theory proposed by Larsen and Walterscheid [1995]. The modified geostrophy theory points out that, in addition to the ion drag associated with the Pedersen conductivity in the upper E region and F region, the ion drag associated to the dominant Hall conductivity in the lower E region has the effect of decreasing the magnitude of the effective Coriolis parameter and changing the neutral flow configuration. Using the electron density profiles measured by the instrumented rocket during the JOULE 2 experiment, the MSIS-E-90 neutral atmosphere parameters, IGRF magnetic field and the formula for the conductivity and collision frequencies given by Roble and Ridley [1987], the horizontal neutral winds were calculated in units of the plasma drift velocity. An enhancement of the wind was found between 110 to 120 km altitude with a large wind shear at the bottom side of the wind maximum. Although the steady-state assumption used to obtain the modified geostrophic wind solution is questionable, the modified geostrophic analysis did give qualitative agreement with the general features of the wind profile observed during the JOULE 2 experiment.
As a summary, mesosphere and lower thermosphere horizontal neutral winds at Poker Flat were measured during the ARIA I though IV, CODA 2, HEX 1, JOULE 1 and JOULE 2 sounding rockets experiments, using triangulation of the TMA tracers released by rockets. Large horizontal neutral winds were observed between 100 to 120 km altitude with the magnitude range from 100 to 250 m s$^{-1}$ during the experiments. In general the horizontal neutral wind is stronger under disturbed condition compared to the quiet condition, and the wind shear below the wind maximum is more unstable with smaller Richardson number close to or smaller than the critical value of 0.25. The unstable wind shear was also lifted to a higher altitude under disturbed condition. Along with the corresponding hodographs of these wind profiles, we conclude that under quiet condition the upward propagating tide is dominant below 120 km and the Lorentz force and viscosity are most important in the region above 120 km. While under disturbed condition, the tidal force is disrupted by the enhanced auroral forcing, i.e. Lorentz force and Joule heating. More specifically, Hall drag is responsible for the enhancement of wind magnitude and wind shear in region of 105 – 125 km, while the Pederson drag is dominant in the region above 125 km. The comparison between the model simulation by TIME-GCM, GITM and localized non-hydrostatic models and observations indicates that including non-hydrostatic status, upward propagating tidal modes, more realistic high latitude electrostatic potential pattern and detailed time history of auroral forcing parameters with sufficiently high vertical resolution would be crucial for models to reproduce the horizontal neutral winds observed in lower $E$ region at high latitudes.

For future work, a data set of mesosphere and thermosphere neutral wind measurements at Poker Flat can be accumulated with the help of the AMISR radar. Study of the tidal influence can be conducted using Fast Fourier transform (FFT) method, to help determine more realistic tidal modes in model simulations as background large-scale wind or forcing exerted at the lower boundary. Improving AMIE procedure to produce the high-latitude electrostatic potential pattern with higher resolution, in order to reveal the small structures. A more detailed time history of electric field, electron profile and other elec-
trodynamic parameters should be available with the help of AMISR at Poker Flat research range. These would then help both global circulation models and localized non-hydrostatic models to successfully reproduce the neutral winds observed during experiments.
BIBLIOGRAPHY


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