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Investigating the Climatology of Mesospheric and Thermospheric Gravity Waves at High Northern Latitudes

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INVESTIGATING THE CLIMATOLOGY OF MESOSPHERIC
AND THERMOSPHERIC GRAVITY WAVES AT
HIGH NORTHERN LATITUDES

by

Michael Ray Negale

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

of

DOCTOR OF PHILOSOPHY

in

Physics

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2018
ABSTRACT

Investigating the Climatology of Mesospheric and Thermospheric Gravity Waves at High Northern Latitudes

by

Michael Ray Negale, Doctor of Philosophy
Utah State University, 2018

Major Professor: Michael J. Taylor, Ph.D.
Department: Physics

The primary objective of this research is to investigate the characteristics and propagation of short-period (<1 hr) gravity waves (GWs) in the upper mesosphere (87 km) and thermosphere (100–300 km) at high northern latitudes, where high-quality measurements are few, but have shown strong wave activity and variability in their propagation directions. Knowledge of the spatial and temporal characteristics, and especially propagation directions, is vital to better understand both the dominant and recurrent sources of GWs and the momentum they transport into the upper atmosphere and ionosphere. New data from three proven instruments (an all-sky airglow imager (ASI), an Advanced Mesospheric Temperature Mapper (AMTM), and an incoherent scatter radar (ISR)) strategically located at two well-separated (>160° longitude) sites, have been used to identify and characterize over 1,200 discrete wave events. Concurrent imaging and radar measurements were made over a three-year period (2010–2013) from the Poker Flat Research Range (PFRR) Alaska (65° N, 147° W), while additional AMTM measurements were used from northern Norway (69° N, 16° E) during the winter of 2011–2012. Together these data have enabled a novel comparative study of the longitudinal variability in wave characteristics in the upper mesosphere and the first investigation of GW climatologies at two adjacent regions, in the upper mesosphere and thermosphere. Importantly, the capability of the Poker Flat ISR to observe waves during both the summer and winter seasons has provided essential new knowledge on the full seasonal thermospheric wave climatology. Chance observations of two SSW events
offer further insight on the effects of induced local/regional changes in the winds on the prevailing GW propagation directions observed in the upper mesosphere. Additionally, the discovery of unusual electron density profiles during these analyses may provide new evidence for GW breaking in the mesopause region and seeding of secondary GWs that can propagate to high thermospheric altitudes.
An important property of the Earth’s atmosphere is its ability to support wave motions, and indeed, waves exist throughout the Earth’s atmosphere at all times and all locations. What is the importance of these waves? Imagine standing on the beach as water waves come crashing into you. In this case, the waves transport energy and momentum to you, knocking you off balance. Similarly, waves in the atmosphere crash, known as breaking, but what do they crash into? They crash into the atmosphere knocking the atmosphere off balance in terms of the winds and temperatures. Although the Earth’s atmosphere is full of waves, they cannot be observed directly; however, their effects on the atmosphere can be observed. Waves can be detected in the winds and temperatures, as mentioned above, but also in pressure and density. In this dissertation, three different studies of waves, known as gravity waves, were performed at three different locations.

For these studies, we investigate the size of the waves and in which direction they move. Using specialized cameras, gravity waves were observed in the middle atmosphere (50–70 miles up) over Alaska (for three winter times) and Norway (for one winter time). A third study investigated gravity waves at a much higher altitude (70 miles on up) using radar data from Alaska (for three years). These studies have provided important new information on these waves and how they move through the atmosphere. This in turn helps to understand in which direction these waves are crashing into the atmosphere and therefore, which direction the energy and momentum are going. Studies such as these help to better forecast weather and climate.
I would like to acknowledge all those who have contributed to this dissertation. I would first like to thank Dr. Mike Taylor for his excellent guidance, teaching, devotion, enthusiasm, time, and expertise. Without the help of Mike Taylor, this dissertation would not be possible. Thank you Mike.

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A.1 Copyright permission for use of Figure 3.4.
1.1 The Neutral Atmosphere

The Earth’s neutral atmosphere is a relatively thin gaseous envelope surrounding our planet. It is held in balance by the upward buoyant force due to pressure differences and the downward force due to gravity (weight), known as hydrostatic equilibrium. The Earth’s atmosphere is almost always stably stratified, meaning that atmospheric density decreases with increasing altitude. The structure of the neutral atmosphere is controlled by the absorption and reradiation of solar energy from the various gases, creating a temperature profile that varies significantly with altitude, geographic location, and time. The atmosphere is divided into four layers, defined by regions where the temperature increases/decreases with altitude, as illustrated in Figure 1.1. They are the troposphere, stratosphere, mesosphere, and thermosphere. The boundaries separating these regions correspond to turning points in the temperature profile, and are called the tropopause, stratopause, and mesopause.

The troposphere begins at the surface of the Earth and extends to \(~10–15\) km. It is characterized by an almost uniform decreasing temperature with altitude of \(~6–7\) K/km. Incoming visible and infrared solar radiation are absorbed and reradiated by the Earth’s land masses and oceans causing warm air to rise and cool. The results of these convective processes makes the troposphere a very active region in terms of weather. The tropopause occurs at \(~15\) km at the equator and \(~8\) km at the poles and limits the upward convection of warm air from below, thus restricting weather phenomena mainly to the troposphere.

From the tropopause to \(~50\) km the temperature increases with altitude, making it stable against convection. This region of the atmosphere is known as the stratosphere. The ozone layer (O$_3$), which resides within this region, absorbs incident solar ultraviolet (UV) radiation and is the primary source of heating in the stratosphere.

Above the stratosphere, the temperature reverts back to decreasing with altitude until \(~90\) km. This region of the atmosphere is known as the mesosphere. Ozone concentration
Figure 1.1: Average high-latitude summer (red) and winter (blue) temperature profiles of the Earth’s neutral atmosphere. [Source: NRLMSISE-00 65° N.]

within this region is very low and there is little absorption of solar radiation. The mesopause is the coldest region in the Earth’s atmosphere, particularly during the high-latitude summer months, where typical temperatures are 150 K, but temperatures as low as 100 K have been reported [e.g., Schmidlin, 1992].

Above the mesosphere, the temperature again increases with altitude. This region of the atmosphere is known as the thermosphere. In this region atomic oxygen absorbs extreme solar ultraviolet (EUV) radiation. Due to the very low atmospheric density, the temperature increases rapidly with altitude and can reach up to 2000 K during solar maximum conditions. EUV radiation also provides enough energy to ionize the neutral atmosphere.

1.2 The Ionosphere

As mentioned in the previous section, solar EUV radiation can provide enough energy to strip electrons from atmospheric molecules (i.e., can ionize the Earth’s upper atmosphere). This charged portion of the atmosphere is known as the ionosphere. In addition to EUV radiation, energetic particles penetrating the Earth’s atmosphere can also provide energy to ionize the Earth’s atmosphere [e.g., Rawer, 1957]. The structure of the ionosphere not only
depends on the production of the free electrons and ions, but also on the disappearance of these particles. There are two principle processes for the disappearance of free electrons: recombination, where an electron attaches with a positive ion, and attachment, where an electron attaches to a neutral molecule [e.g., Rawer, 1957].

The structure of the ionosphere is also determined by the neutral atmospheric constituents present. Figure 1.2 shows both the dominant ion densities and the dominant neutral densities above ∼100 km altitude. From the Earth's surface to ∼100 km, the neutral atmosphere is well mixed by turbulence. Above this altitude, called the turbopause, the lack of turbulence allows the atmospheric constituents to separate where the vertical distribution of each neutral gas depends on its molecular weight [e.g., Rishbeth and Garriott, 1969].

The ionosphere is characterized by changes in the electron density profiles with altitude, as seen in Figure 1.2 red line labeled e⁻. The electron density profiles vary with solar zenith angle, time of day, season, solar cycle, and solar activity, as illustrated in Figure 1.3. For example, during the daytime more EUV is incident on the neutral atmosphere (termed atmosphere from now on) causing a dramatic increase in the electron density (Figure 1.3 red) at ∼250 km. This F region peak is reduced during the nighttime (Figure 1.3 blue) due to the lack of absorption of EUV, and the D layer disappears as soon as the absorption of

![Figure 1.2](image_url)

**Figure 1.2:** Atmospheric composition profiles based on spectrometer measurements taken above White Sands, New Mexico. [Modified from Kelly [1989], their Figure 1.2]
EUV is gone. This layer disappear because molecular ions, such as NO$^+$ and O$_2^+$ (which are abundant at these altitudes, see Figure 1.2), recombine with an electron faster than an electron combines with an atomic ion, where energy and momentum are more easily conserved if two bodies are produced [e.g., Hargreaves, 1979].

The ionosphere is divided into three regions (with symbols D, E, and F) using the relative maxima and minima of the daytime electron density profile, as shown by the red line in Figure 1.3. The D region from $\sim$50–100 km consists mostly of N$_2$, O$_2$, Ar, and He (Figure 1.2). The ionized component of the D region consists mostly of NO$^+$ and O$_2^+$ with typical electron densities around $10^4$ el/cm$^3$ (Figure 1.2). The ionosphere from $\sim$100–140 km is called the E region. The ionized constituents of the E region consist mostly of NO$^+$ and O$_2^+$ with typical electron densities around $10^4$–$10^5$ el/cm$^3$. Peak plasma density occurs in the region $\sim$140–500 km. This region is known as the F region, which can have a peak plasma density as high as $10^6$ el/cm$^3$ near noontime and is dominated by O$^+$ [e.g., Kelly, 1989].

Figure 1.3: Typical plasma density daytime (red) and nighttime (blue) profiles. [Source: International Reference Ionosphere for local midnight and local noon.]
Figures 1.1 and 1.3 show modeled atmospheric and ionospheric temperature and density profiles, respectively. These profiles should be taken as idealized. When considering real measured profiles, there is almost always evidence of wave structure in both the neutral and ionized regions of the atmosphere.

1.3 Waves in the Atmosphere

One of the most important properties of the atmosphere is its ability to support wave motion, often observed as wave-like features in tropospheric clouds (Figure 1.4). It is now known that the atmosphere contains a broad spectrum of waves that exhibit a large range in their spatial and temporal sizes. The largest scale waves are known as planetary waves, or Rossby waves, which are global in nature and exhibit horizontal wavelengths of thousands of km and periods of typically 2, 5, 10, and 16 days [e.g., Andrews et al., 1987; Forbes, 1995]. These waves are thought to be generated by strong wind flow across large mountain ranges, such as the Rocky Mountains or Andes, or by large-scale heating differences between continents and oceans (predominantly the northern hemisphere due to the larger land mass). Planetary waves are either stationary as observed from ground or zonally propagating with

![Figure 1.4: Example of tropospheric clouds exhibiting extensive wave-like structure (Lee waves) over Utah State University campus. [Courtesy of V. Chambers.]](image-url)
low observed phase speeds [e.g., Forbes, 1995]. The next largest scale waves are atmospheric tides. These waves are generated by absorption of solar UV radiation in the atmosphere (mainly by stratospheric ozone and tropospheric water vapor) and hence, primarily follow the daily heating cycle. Tides are also global in scale and propagate with the motion of the sun (termed migrating tides) with dominant periods of 24, 12, and 8 hr, and exhibiting horizontal wavelengths of several thousand km [e.g., Andrews et al., 1987; Forbes, 1995; Fuller-Rowell, 1995].

Atmospheric gravity waves (GWs) are another type of wave that exists in the Earth’s atmosphere. GWs are known to originate mostly in the tropospheric region and are produced by tropospheric wind flow over mountain ranges, latent heat releases from convective regions, and shear associated with jet streams. Different sources for GWs determine the size, frequency of occurrence, and direction of propagation. Typical periods range from \( \sim 5 \) min (atmospheric buoyancy period) up to \( \sim 12 \) hr, the inertial period.

Short-period GWs (periods <1 hr) are known to play important roles in the dynamics and thermal structure of the neutral upper atmosphere on a local, regional, and global scale [e.g., Fritts and Alexander, 2003, and references therein]. GW studies in the mesosphere and lower thermosphere (MLT) region (~80–110 km) have shown these waves have strong responses in the atmosphere as they dissipate and deposit significant amounts of energy and momentum. Globally, this process is responsible for the closure of the mesospheric jets that leads to a strong mean meridional circulation that significantly cools the summer mesopause and warms the winter mesopause [e.g., Garcia and Solomon, 1985; Holton, 1982, 1983; Lindzen, 1981].

Knowledge of these short-period GW spatial and temporal characteristics, especially propagation directions, is vital to better understand their dominant/recurrent sources, global distributions, and where/how much energy and momentum they deposit as they dissipate. Various instruments are used to observe/study GWs such as radar, lidar, weather balloons, satellites, spectrometers, and all-sky airglow imagers. Studies of the climatology of short-period GWs have now been performed at many observation sites at mid and low
latitudes [e.g., Ejiri et al., 2003; Hecht et al., 2001, 2004; Medeiros et al., 2003; Nakamura et al., 2003; Pautet et al., 2005; Stockwell and Lowe, 2001a; Suzuki et al., 2004, 2011; Taylor et al., 1993, 1997; Walterscheid et al., 1999; Wu and Killeen, 1996]. The majority of these studies used all-sky airglow imagers (discussed in Chapter 3) to observe GWs. The observed spatial and temporal scales from these measurements have similar ranges, despite their geographic origin, suggesting similar global nature. However, a key outstanding result is that the wave propagation headings frequently show considerable site-to-site, latitudinal, and seasonal variation. Moreover, investigations of short-period GW activity at polar latitudes are still few [e.g., Nielsen et al., 2006, 2009; Suzuki et al., 2009] primarily due to difficult observing conditions and the frequent presence of aurora, the intensity of which can over power the relatively faint airglow emissions. As such, there is an important need for multi-year studies of GWs at polar latitudes to establish their climatologies.

1.4 Medium Scale Traveling Ionospheric Disturbances (MSTIDs)

In Section 1.2, the ionosphere was introduced as being a collection of free electrons and ions. Most of our knowledge of the ionosphere has come from the use of radio waves transmitted from the surface of the Earth, then reflected from the plasma in the D, E, and F regions, and then received back on Earth. Currently, we rely heavily on communication with satellites both inside and outside of the Earth’s ionosphere. Ionospheric irregularities are responsible for radio wave scintillation and radar backscatter and could interrupt radio wave communications [e.g., Tsunoda, 1988].

While investigating these irregularities and irregular motions of the ionosphere, regular periodic motions were also observed [e.g., Beynon, 1948; Munro, 1948, 1950; Pierce and Mimno, 1940]. Pierce and Mimno [1940] were the first to hypothesize that these short-period fluctuations indicated wave motions in the atmosphere. Hines [1960] recognized similarities between these periodic ionospheric motions with properties of GWs in the neutral atmosphere, and provided the theoretical understanding of these ionospheric manifestations termed traveling ionospheric disturbances (TIDs). The hypothesis that TIDs are ionospheric manifestations of GWs, with the ionosphere acting as a passive tracer to
display the motions of the neutral atmosphere, has since received complete acceptance [e.g., Francis, 1975].

A broad range of wave observations and modeling studies have shown that a spectrum of GWs generated in the lower atmosphere can penetrate well into the thermosphere [e.g., Bauer, 1958; Bishop et al., 2006; Francis, 1973; Fritts and Vadas, 2008; Georges, 1968; Hocke and Tsuda, 2001; Hung and Smith, 1978; Kelley, 1997; Röttger, 1977; Vadas, 2007; Vadas and Nicolls, 2008, 2009; Wallock and Jones, 1987]. Depending on their characteristics, these waves may play a number of important roles in both neutral thermosphere and plasma processes. For example, direct acceleration due to GW momentum flux divergence at/near dissipation altitudes can result in substantial local forcing to the thermosphere [e.g., Fritts and Lund, 2011; Garcia and Solomon, 1985; Liu et al., 2013; Vadas and Fritts, 2001, 2005]. In addition, GWs capable of penetrating into the bottom side F region (∼150–300 km) may provide the seeding mechanism for the generation of equatorial spread F plasma bubbles [e.g., Huang and Kelley, 1996; Huang et al., 1993; Hysell et al., 1990; Sekar and Kelley, 1998; Sekar et al., 1995; Taylor et al., 1998].

Since the pioneering work of Hines [1960], there have been many observations of TIDs in the high, mid, and low latitudes using a variety of techniques such as (a) electron density measurements using vertical and oblique ionosondes [e.g., Afraimovich et al., 2008; Morgan et al., 1978; Tedd et al., 1984] and incoherent scatter radars [e.g., Djuth et al., 1994, 1997; Fukao et al., 1993; Nicolls and Heinselman, 2007], (b) backscatter radar soundings [e.g., Bristow and Greenwald, 1996; Bristow et al., 1994; Frissell et al., 2014, 2016], (c) Continuous Wave (CW)–Doppler measurements [e.g., Georges, 1967; Hung et al., 1979; Röttger, 1977], (d) total electron content (TEC) measurements using Global Positioning System (GPS) satellite constellations [e.g., Galushko et al., 2016; Nicolls et al., 2004; Onishi et al., 2009], (e) and optical airglow measurements [e.g., Kubota et al., 2011; Paulino et al., 2016].

These numerous studies have identified two general TID categories each with differing characteristic scales driven primarily by their different sources. Georges [1968] labeled these categories as large-scale TIDs (LSTIDs) and medium-scale TIDs (MSTIDs). LSTIDs have
been shown to originate in the high-latitude upper atmosphere and are generated mainly by Joule heating and auroral particle precipitation events [e.g., Chimonas and Hines, 1970; Richmond, 1978; Testud, 1970]. LSTIDs exhibit periods in the range ~0.5–3 hr, horizontal wavelengths of 1000 km to several thousands km, and phase speeds between 400–1000 m/s [e.g., Hocke and Schlegel, 1996]. In contrast, MSTIDs exhibit periods <1 hr, horizontal wavelengths of ~100–1000 km, and typical phase speeds between ~50–400 m/s [e.g., Ogawa et al., 1987; Samson et al., 1990]. MSTIDs are a common occurrence in the thermosphere/F-region ionosphere and are known to originate primarily from weather driven sources in the lower atmosphere [e.g., Crowley et al., 1987; Georges, 1968; Hocke and Schlegel, 1996; Ogawa et al., 1987; Waldock and Jones, 1986].

1.5 Overview of This Work

In this dissertation, short-period gravity waves (GWs) are investigated at two widely spaced (160° longitude) geographic locations: over central Alaska (65° N) and over northern Norway (69° N) to characterize their properties at high latitudes. They are also investigated at two different altitude regions: the MLT region and the thermosphere.

Chapter 2 introduces GWs, their sizes and how they are created in the atmosphere. Using the conservation of momentum, mass, and energy, linear GW theory describes how these waves propagate in a realistic atmosphere and how these waves are affected by the background atmosphere. Wave dispersion in the MLT region is significantly different than wave dispersion in the thermosphere. Chapter 2 covers both dispersion relations: one for the MLT and the other for the thermosphere. Since the atmosphere affects how GWs propagate, knowledge of the background atmosphere (winds and temperatures) are needed to fully utilize the dispersion relations. However, most of the time these data are not available; and therefore, available model atmospheres are used.

Chapters 3 and 4 describe the instrumentation used in three main studies: an all-sky imager, an Advanced Mesospheric Temperature Mapper, and an incoherent scatter radar. The analysis processes for preparing the data, as well as the methods used to extract the wave parameters, are also described.
The following chapters 5, 6, and 7 report the results of each study. All studies include a statistical analysis of the observed wave parameters. Seasonal variation of the parameters are also investigated, revealing important information on their dominant directions of propagation at each site. These variations are analyzed via critical level wind filtering theory.

Chapter 8 compares the results obtained from the three different studies. These are also compared with two other GW studies performed using similar optical instruments in Canada and Svalbard. Our novel seasonal MSTID results are compared with other MSTID studies at high latitudes.

Chapter 9 summarizes key results and presents future work and new data on potential secondary GW generation in the thermosphere.
2.1 Introduction

As mentioned in Chapter 1, waves in the atmosphere have a wide range of spatial and temporal characteristics. These characteristics depend on the sources of the waves and determine how these waves propagate within the atmosphere. This chapter introduces what wave characteristics are important for these studies, and how these waves are created in the atmosphere, how the waves propagate in the mesosphere lower thermosphere (MLT) region (∼80–110 km), as well as in the thermospheric region (∼100–400 km). As will be shown, wave propagations are affected by the background atmosphere; and therefore, this chapter concludes with an introduction to atmospheric models used to create a background atmosphere.

2.2 Buoyancy Waves

The atmospheric conditions to support gravity wave (GW) propagation in the atmosphere, in particular the characteristic frequency of oscillation, are presented in this section. Following Nappo [2002], in general, GWs propagate at an angle and, therefore, consider a parcel of air that is displaced from its equilibrium position along an inclined plane, as shown in Figure 2.1. If the parcel of air is displaced δs from its equilibrium position along the plane inclined at an angle, β, Newton’s second law along the direction of displacement (S) is

\[ m_p \frac{d^2(\delta s)}{dt^2} = (m_a - m_p) g \sin(\beta), \]

(2.1)

where \( m_p \) is the mass of the air parcel, \( m_a \) is the mass of the surrounding air, and \( g \) is the acceleration due to gravity. If the masses, in terms of densities and volumes, are used along with the ideal gas law, Equation 2.1 can be written as

\[ \frac{d^2(\delta s)}{dt^2} = \frac{g}{T_a} (T_p - T_a) \sin(\beta), \]

(2.2)
where $T_a$ and $T_p$ are the temperatures of the surrounding air and the parcel of air, respectively. Performing an expansion around the temperature $T_0$ (the temperature at the equilibrium altitude position, $z_0$) and assuming the surrounding air temperature does not change much as the parcel of air is displaced, Equation 2.2 can be written as

$$\frac{d^2(\delta s)}{dt^2} = -g \frac{T_a}{T_a} \left( \frac{\partial T_a}{\partial z} - \frac{\partial T_p}{\partial z} \right) \sin^2(\beta) \delta s. \quad (2.3)$$

By introducing the lapse rate ($\Gamma$) as the negative temperature gradient with respect to the vertical,

$$\Gamma = -\frac{\partial T}{\partial z}, \quad (2.4)$$

Equation 2.3 can be written as

$$\frac{d^2(\delta s)}{dt^2} = -g \frac{T_a}{T_a} (\Gamma_p - \Gamma_a) \sin^2(\beta) \delta s, \quad (2.5)$$

where $\Gamma_p$ is the adiabatic lapse rate for the parcel of air and $\Gamma_a$ is the ambient (or environmental) lapse rate.

Equation 2.5 has the general solution,

$$\delta s(t) = A e^{i N' t} + B e^{-i N' t}, \quad (2.6)$$
where $N'$ is the frequency of oscillation given by

$$
N' = \sqrt{\frac{g}{T_a}} (\Gamma_p - \Gamma_a) \sin \beta
$$

(2.7)

$$
= N \sin \beta,
$$

where

$$
N = \sqrt{\frac{g}{T_a}} (\Gamma_p - \Gamma_a)
$$

(2.8)

is known as the Brunt-Väisälä frequency, the natural frequency of oscillation of the displaced air parcel [e.g., Hines, 1960]. Figure 2.2a shows the Brunt-Väisälä period ($\tau_B = \frac{2\pi}{N}$) derived using Equation 2.8 with the temperature profile shown in Figure 2.6a and setting the adiabatic lapse rate to $\Gamma_p \approx 10 \text{ K/km}$ (the adiabatic lapse rate for dry air).

If the atmospheric temperature gradient is greater than the adiabatic lapse rate ($\Gamma_a > \Gamma_p$), then $N$ (Equation 2.8) is an imaginary number, and the solution, $B$, (Equation 2.6) represents unbounded growth of the displacement, meaning an instability. If $\Gamma_p > \Gamma_a$, then $N$ is a real number and oscillatory motion is achieved, and the atmosphere is said to be

**Figure 2.2**: Using the NRLMSISE-00 temperature profile in Figure 2.6a, the (a) buoyancy period ($\tau_b$), and (b) scale height ($H_s$) are derived as a function of altitude.
stably stratified. It will be shown in Section 2.3 that $N$ represents the maximum frequency for vertically propagating GWs ($\beta = 0^\circ$).

2.3 Linear Gravity Wave Theory

A simple dispersion relation ($c = \frac{\omega}{K}$) would suffice if the atmospheric medium was uniform (in temperature, density, flow, etc...). However, the atmosphere is a complex medium and the variations in the background temperature, density, and neutral wind field, through which the waves propagate, need to be considered. The fundamental fluid equations are linearized, following Hines [1960] (known as linear GW theory), to gain more information about the wave field.

The fundamental fluid equations in Cartesian coordinates, ignoring the effects of viscosity and friction, [e.g., Fritts and Alexander, 2003; Holton, 1992] are:

$$ \frac{D\vec{U}}{Dt} = -\frac{1}{\rho} \nabla p + \vec{g} - 2\vec{\Omega} \times \vec{U}, $$  \hspace{1cm} (2.9)

$$ \frac{\partial \rho}{\partial t} + \rho \nabla \cdot \vec{U} = 0, $$  \hspace{1cm} (2.10)

$$ \frac{D\theta}{Dt} = 0, $$  \hspace{1cm} (2.11)

where $\vec{U} = (u, v, w)$ is the fluid velocity vector, where $u$, $v$, and $w$ are the zonal, meridional, and vertical winds, respectively. The advective derivative is defined as

$$ \frac{D}{Dt} = \frac{\partial}{\partial t} + \vec{U} \cdot \nabla. $$  \hspace{1cm} (2.12)

Equations 2.9, 2.10, and 2.11 represent the conservation of momentum, the conservation of mass, and the conservation of energy, respectively. The remaining symbols are: $\vec{g}$ is the acceleration of gravity, $p$ is pressure, $\rho$ is density, $\theta$ is potential temperature (defined below), and $\vec{\Omega}$ is the Earth’s rotation rate.
The equation of energy (Equation 2.11) introduces the potential temperature ($\theta$), which is the temperature that a parcel of air at pressure, $p$, and temperature, $T$, would obtain if it expanded/compressed adiabatically to a standard pressure, $p_s$, written as

$$\theta = T \left( \frac{p_s}{p} \right)^{\frac{R}{c_p}}, \quad (2.13)$$

where $R$ is the universal gas constant and $c_p$ is the heat capacity at constant pressure.

The fluid equations are linearized by assuming a steady, slow varying background state ($q_0$), which is allowed to vary in the vertical direction, but is uniform in the horizontal direction and to consider perturbations ($q_1$) to the background. The variables then have the following general form:

$$q(x, z, t) = q_0(z) + q_1(x, z, t). \quad (2.14)$$

These perturbations are assumed to be much smaller than the background values and do not change the background state. Since the perturbations are small, products of perturbations are also small and are, therefore, neglected. The perturbations can be caused by several mechanisms such as turbulence, plumes, etc., however, we only consider perturbations caused by GWs. By neglecting products of perturbations, interactions of waves are neglected, and therefore, the waves cannot create or destroy existing waves [e.g., Nappo, 2002].

It is common to neglect the wind shear terms by assuming the background wind vector, $\vec{U}_0 = (u_0, v_0, 0)$, and buoyancy frequency, $N$, vary slowly over a wave cycle in the vertical [e.g., Fritts and Alexander, 2003; Gill, 1982]. A plane wave solution of the form

$$q_1 = \vec{q} \cdot \exp \left[ i (k_x x + k_y y + k_z z - \omega t) + \frac{z}{2H_s} \right], \quad (2.15)$$

where $\vec{K} = (k_x, k_y, k_z)$ is the wave vector with components along the east, north, and vertical directions, respectively, is assumed for the variables $\vec{u}$, $\vec{v}$, $\vec{w}$, $\vec{\theta}$, $\vec{p}$, and $\vec{\rho}$. Equation 2.15 introduces the scale height, the distance over which a quantity (such as density or pressure)
decreases by a factor of $e$, defined as

$$H_s = \frac{RT}{g}. \quad (2.16)$$

Figure 2.2b shows the scale height derived using the temperature profile in Figure 2.6a. The result, neglecting the imaginary terms, yields the following dispersion relation:

$$\hat{\omega}^2 \left( k_x^2 + k_y^2 + k_z^2 + \frac{1}{4H_s^2} - \frac{\hat{\omega}^2 - f^2}{c_s^2} \right) = N^2 \left( k_x^2 + k_y^2 \right) + f^2 \left( k_z^2 + \frac{1}{4H_s^2} \right), \quad (2.17)$$

where $f = 2|\vec{\Omega}|$ is the Coriolis parameter, $c_s$ is the speed of sound, and

$$\hat{\omega} = \omega - k_x u_0 - k_y v_0 \quad (2.18)$$

is the intrinsic frequency (the frequency that would be observed in a frame of reference that is moving with the background wind $\vec{U}_0$). Here, $u_0$ and $v_0$ are the background zonal and meridional winds, respectively.

This dispersion relation, Equation 2.17, is fourth order in $\hat{\omega}$ and supports both acoustic waves and GWs. For an acoustic wave, the restoring force is due to the change in pressure caused when the gas is compressed. However, as shown in Section 2.2, the restoring force for GWs is due to gravity and the buoyancy force on a displaced parcel of air. The basic properties of acoustic waves are that they are isotropic and nondispersive, meaning the waves propagate equally well in all directions and with a wave speed that is constant and depends only on the properties of the medium. GWs are also isotropic; however, they are dispersive. A dispersion relation to investigate acoustic waves and GWs is developed next.

The Coriolis parameter ($f$) is important when considering low-frequency (long-period) GWs near the inertial frequency. However, for high-frequency (short-period, <1 hr) GWs the Coriolis parameter is insignificant. Chapters 3–4 present studies of short-period GWs in the MLT region; therefore, the Coriolis parameter is neglected and Equation 2.17 simplifies
to
\[ K^2 = \frac{\omega^2}{c_s^2} \left( \frac{c_s^2 - \hat{\omega}^2}{\frac{c_s^2}{NH^2} \cos^2 \alpha - \hat{\omega}^2} \right), \quad (2.19) \]

where
\[ \cos^2 \alpha = \frac{k_H^2}{K^2}, \quad (2.20) \]

\( k_H = \left( k_x^2 + k_y^2 \right)^{\frac{1}{2}} \) is the horizontal wave number and \( \alpha \) is the angle between lines of constant phase and the vertical. In Section 2.2, an angle \( \beta \) was defined as an angle measured from the horizontal to the direction of the air parcel displacement. Therefore, \( \alpha + \beta = 90^\circ \), and Equation 2.7 can be written as
\[ N' = N \cos \alpha. \quad (2.21) \]

The scale height can be written in terms of the speed of sound \( (c_s) \) as
\[ H_s^2 = \frac{c_s^2}{4 \omega_a^2}, \quad (2.22) \]

where \( \omega_a \) will be investigated below. Using this form of the scale height, the dispersion relation, Equation 2.19, becomes
\[ K^2 = \frac{\omega^2}{c_s^2} \left( \frac{\omega_a^2 - \hat{\omega}^2}{(N')^2 - \hat{\omega}^2} \right). \quad (2.23) \]

This simplified form of the dispersion relation allows us to investigate the acoustic and GW domains. Waves that can propagate, known as freely propagating waves, require \( K^2 > 0 \), which means the numerator and the denominator in Equation 2.23 must have the same sign, yielding two possible cases. The first case is when \( \hat{\omega} > \omega_a \) and \( \hat{\omega} > N' \), known as the acoustic wave branch. The second case is when \( \hat{\omega} < \omega_a \) and \( \hat{\omega} < N' \), known as the internal GW branch (or simply the GW branch). Therefore, \( \omega_a \) is the acoustic wave cut-off frequency and \( N \) (the Brunt-Väisälä frequency) is the maximum frequency for GWs.

If \( K^2 < 0 \), we do not have freely propagating waves. This yields two more possible cases;
however, since $\omega_a > N'$, only a single case is left: $\omega_a < \omega < N'$. This region is a forbidden region, where no waves can have a vertical component of propagation. Figure 2.3 shows the relationship between the acoustic wave and GW branches, as well as the forbidden regions.

For the study of short-period GWs in the MLT region $m^2 = k_z^2 << \frac{1}{4H_s^2}$, can be shown using the mean vertical wavelengths from Chapter 6, $\lambda_z = 14$ km and a scale height $H_s = 7$ km, and the term $\frac{1}{4H_s^2}$ can be neglected yielding a more simplified GW dispersion relation,

$$\omega^2 = N^2 \frac{k_H^2}{K_z^2} = N^2 \cos^2 \phi,$$

where $\phi$ is an angle measured from the horizontal to the wave vector, as illustrated in

**Figure 2.3:** Relationship between the acoustic wave and GW branches, and the forbidden regions. $\omega_B$ represents the Brunt-Väisälä frequency and $\omega_a$ represents the acoustic wave cut-off frequency. [Courtesy Nielsen [2007], Ph.D. Dissertation, their Figure 2.3.]
Figure 2.4. The frequencies can be rewritten in terms of the periods along with trigonometry identities to obtain

$$\tan \phi = \left( \frac{\tau_I^2}{\tau_B^2} - 1 \right)^{\frac{1}{2}},$$  

(2.25)

where $\tau_I$ is the intrinsic period and $\tau_B$ is the Brunt-Väisälä period. This means that the propagation directions for smaller period waves propagate at steeper angles than longer period waves, as shown in Figure 2.5 for three intrinsic periods.

The relationship between the direction of propagation and the group velocity for short-period GWs are now considered. Assuming no background winds ($\vec{U}_0 = 0$) and using the simplified dispersion relation (Equation 2.24), the group velocity is

$$\vec{c}_g = \frac{\partial \omega}{\partial k_x} = (c_{gh}, c_{gz}) = \frac{N}{K^3} (m^2, -m k_H),$$

(2.26)

where $c_{gh}$ is the horizontal group velocity and $c_{gz}$ is the vertical group velocity. If a GW is propagating upwards, then $m > 0$ and $c_{gz} < 0$, meaning the direction of energy propagation

\[\text{Figure 2.4: A wave with wave vector, } \vec{K}, \text{ and wave fronts (dashed lines) propagating at an angle, } \phi, \text{ from the horizontal, illustrating an upward propagating GW. The group velocity, } \vec{c}_g, \text{ shows the direction of energy propagation, in this case, downward energy propagation. [Modified from Nappo [2002], their Figure 2.1.] }\]
is downwards, as illustrated in Figure 2.4. Furthermore, it can be shown that the group velocity is parallel to the phase fronts.

2.4 Wave Dispersion in the Thermosphere

The GW dispersion relation in Equation 2.19 introduced in the previous section was derived by neglecting effects such as molecular viscosity and thermal diffusivity, and has been shown to accurately represent short-period GWs in the stratospheric and mesospheric regions. Observations and modeling studies have shown that GWs with certain characteristics can penetrate well into the thermosphere, >110 km altitude [e.g., Bauer, 1958; Bishop et al., 2006; Francis, 1973; Fritts and Vadas, 2008; Georges, 1968; Heale et al., 2014; Hocke and Tsuda, 2001; Hung and Smith, 1978; Kelley, 1997; Röttger, 1977; Vadas, 2007; Vadas and Liu, 2009; Vadas and Nicolls, 2008, 2009; Vadas et al., 2014; Waldock and Jones, 1987]. Above this altitude molecular viscosity and thermal diffusivity become important due to the higher rate of decreasing atmospheric density (Figure 2.6b). Therefore, this section presents a modified GW dispersion relation developed by Vadas and Fritts [2005] to account for the dissipation effects in the thermospheric region.

Starting with the nonlinear, compressible fluid equations, Equations 2.9–2.11, and now including molecular viscosity and thermal diffusivity to account for the appropriate diffusion
processes for high-frequency GWs:

\[
\frac{D\vec{U}}{Dt} = -\frac{1}{\rho} \vec{\nabla}p + \vec{g} + \frac{\mu}{\rho} \left[ \nabla^2 \vec{U}_1 + \frac{1}{3} \nabla (\nabla \cdot \vec{U}_1) \right],
\]

(2.27)

\[
\frac{\partial \rho}{\partial t} + \rho \vec{\nabla} \cdot \vec{U} = 0,
\]

(2.28)

\[
\frac{D\theta}{Dt} = \frac{\kappa \theta}{c_p \rho T} \nabla^2 T_1,
\]

(2.29)

where \(\mu\) is molecular viscosity, \(\kappa\) is thermal conductivity, \(\vec{U}_1\) is the velocity perturbation and \(T_1\) is the temperature perturbation. The remaining variables are defined in the previous section. In Chapter 7, results of short-period GWs in the thermosphere are presented; and therefore, the Coriolis force is neglected, as in the previous section.

The thermal conductivity can be written as

\[
\kappa = \frac{\nu c_p \rho}{Pr},
\]

(2.30)

where \(\nu = \frac{\mu}{Pr}\) is the kinematic viscosity and \(Pr\) is the Prandtl number. The molecular viscosity contains two pieces, where the \(\frac{1}{3} \nabla (\nabla \cdot \vec{U}_1)\) term is negligible with respect to the term \(\nabla^2 \vec{U}_1\) in an atmosphere with molecular viscosity for GWs with \(m^2 >> \frac{1}{4H_s^2}\) (as was assumed in the previous section), and is neglected in the following derivations.

Linearizing the fluid equations for the variables \(\tilde{u}, \tilde{v}, \tilde{w}, \tilde{\rho},\) and \(\vec{T}\) yields the compressible, complex, dispersion relation of GWs damped by both molecular viscosity and thermal diffusivity:

\[
-\hat{\omega}^2 \left( \hat{\omega} - i \alpha \nu \right)^2 \left( 1 - \frac{i \gamma \alpha \nu}{Pr \hat{\omega}} \right) + \left( \hat{\omega} - i \alpha \nu \right) \left( \hat{\omega} - i \frac{\alpha \nu}{Pr} \right) \left( K^2 + \frac{1}{4H_s^2} \right) = k_H N^2,
\]

(2.31)

where \(\hat{\omega}\) is the intrinsic frequency, \(c_s\), is the speed of sound, \(\alpha = -K^2 + \frac{1}{4H_s^2} + \frac{im}{H_s}\), and \(i = \sqrt{-1}\). In the limit that molecular viscosity and thermal diffusivity are negligible, this dispersion relation reduces to Equation 2.17 (without the Coriolis parameter, \(f\)).
As before, the dispersion relation is restricted to GWs by neglecting the acoustic wave branch:
\[
(\hat{\omega} - i \alpha \nu)(\hat{\omega} - \frac{i \alpha \nu}{Pr}) = \frac{k_H^2 N^2}{K^2 + \frac{1}{4 H_s^2}}.
\] (2.32)

Where molecular viscosity and thermal diffusivity are negligible in the stratosphere and mesosphere, Equation 2.32 reduces to
\[
\hat{\omega}^2 = \frac{k_H^2 N^2}{K^2 + \frac{1}{4 H_s^2}}
\] (2.33)

and is the same as Equation 2.24 (keeping the extra term \(\frac{1}{4 H_s^2}\)).

For this derivation, a complex intrinsic frequency and a real vertical wave number are input into the dispersion relation. This results in a time-decaying wave amplitude, and means we can determine the solution for any value of viscosity and thermal diffusivity, and so describes the GW before, after, and during dissipation. The intrinsic frequency is written as the sum of real and imaginary parts:
\[
\hat{\omega} = \omega_{Ir} + i \omega_{Ii},
\] (2.34)

where \(\omega_{Ir}\) is real and relates to the intrinsic frequency of the GW, to the wave structure, buoyancy frequency, and damping due to kinematic viscosity and thermal diffusivity. \(\omega_{Ii}\) expresses the inverse decay rate of the wave amplitude with time due to kinematic viscosity and thermal diffusivity [e.g., Vadas and Fritts, 2005].

Substituting Equation 2.34 into Equation 2.32 yields the GW dispersion relation:
\[
\omega_{Ir}^2 + \frac{\nu^2}{4} \left( K^2 - \frac{1}{4 H_s^2} \right)^2 \left( 1 - Pr^{-1} \right)^2 \frac{(1 + \delta_+ + \frac{\delta^2}{Pr})}{(1 + \frac{\delta_+}{2})^2} + \frac{\nu \omega_{Ir}}{H_s} + \frac{\nu^2 m^2}{Pr H_s^2} = \frac{k_H^2 N^2}{K^2 + \frac{1}{4 H_s^2}},
\] (2.35)

where \(\delta = \nu m / (H_s \omega_{Ir})\), \(\delta_+ = \delta(1 + Pr^{-1})\), and \(\nu_+ = \nu(1 + Pr^{-1})\). In the limit that molecular viscosity and thermal diffusivity are negligible in the stratosphere and mesosphere, the damping rate goes to zero \((\omega_{Ii} = 0)\), \(|\delta| << 1\) and \(|\delta_+| << 1\), and Equation 2.35 reduces to Equation 2.33.
2.5 Utilizing Background Atmospheric Models

As we saw in the previous two sections, the short-period GW dispersion relations in the MLT and thermospheric regions depend on the background winds, neutral temperatures, and densities. When making real observations, these data are not always available, as is the case in a majority of observations in the following chapters. To gain a better idea of how these waves propagate within the atmosphere, this section introduces well-known climatological models for the background winds, temperatures, and densities to be used along with the dispersion relations.

2.5.1 Background Temperatures and Densities

NRLMSISE-00 (where NRL stands for Naval Research Lab) is a major upgrade to the Mass Spectrometer and Incoherent Scatter (MSIS) class of models that provides neutral temperatures and number densities for He, O, N$_2$, O$_2$, Ar, H, N, and anomalous oxygen (which accounts for appreciable O$^+$ and hot atomic oxygen contributions to the total mass density at high altitudes) [e.g., Hedin, 1987, 1991; Hedin et al., 1977; Picone et al., 2002]. The model includes variations in latitude, longitude, annual, semiannual, local time, and solar cycle. Figure 2.6 shows an example of the daily averaged (a) neutral temperature, and (b) total mass density for 1 December 2011.

2.5.2 Background Winds

The horizontal wind model (HWM) is the most comprehensive empirical model available that provides the atmosphere’s vector wind fields [e.g., Drob et al., 2015; Emmert et al., 2008; Hedin et al., 1996; Hedin, 1991; Hedin et al., 1988]. HWM provides the background winds and a geomagnetic storm time component (DWM07) [e.g., Emmert et al., 2008] provides storm induced winds. Together, they provide horizontal winds from the surface of the Earth to ∼450 km altitude as a function of latitude, longitude, altitude, day of year, and time of day.

Figure 2.7 shows HWM14 quiet time wind profiles every three hours (gray lines) on 1 December 2011, as well as the daily averaged quiet winds (solid black). The dashed black
Figure 2.6: NRLMSISE-00 daily-averaged (a) neutral temperatures, and (b) total mass densities for 1 December 2011.

line corresponds to the quiet HWM14 winds plus disturbed DWM07 winds set with Ap=80.

2.5.3 MERRA

The models described in the previous sections, NRLMSISE-00 and HWM14, provide background temperatures, densities, and winds from the ground to well in the thermosphere. Initially, these models described the background atmosphere in the thermospheric region and were extended to the ground. In order to extend these models to the lower atmosphere, they used tabulated data such as from CIRA-86 and the MAP Handbooks [e.g., Drob et al., 2008; Picone et al., 2002].

The Modern-Era Retrospective for Research and Applications (MERRA) is a reanalysis that combines model fields with observations distributed irregularly in space and time into a spatially complete gridded meteorological dataset, with an unchanging model and analysis system spanning the historical data record [e.g., Rienecker et al., 2011].

MERRA data used in the studies in Chapters 3 and 4 utilized the MERRA instantaneous six hourly three-dimensional direct analysis on a native resolution of $0.5^\circ \times 0.66^\circ$.
grid and 72 levels extending from 985-0.01 hPa. To convert from pressure levels to altitude (in km), the following was used

\[ z = H_s \log \left( \frac{L_0}{L} \right), \tag{2.36} \]

where \( H_s \) is the scale height, \( L_0 \) is the surface level pressure, and \( L \) is the pressure level.

Figure 2.8a–b shows the daily averaged MERRA horizontal winds (black lines) and compares with HWM14 daily averaged winds (gray lines). Figure 2.8c shows the daily averaged MERRA temperatures (black line) and compares with NRLMSISE-00 daily averaged temperatures (gray line) for 1 December 2011. The MERRA temperatures agree a lot better than the HWM14 winds, although the shapes of the wind profiles are similar.

2.6 Critical Level Wind Filtering

Consider a wave with horizontal wave vector, \( \vec{k}_H \), propagating at an angle, \( \phi \), (as measured counterclockwise from East) in a uniform horizontal wind field, \( \vec{U} = (V_{z_0}, V_{m_0}) \),
where \( V_{z0} \) and \( V_{m0} \) are the zonal and meridional winds, respectively, as illustrated in Figure 2.9. The intrinsic phase speed of the wave is obtained by dividing the intrinsic frequency (Equation 2.18) by the horizontal wavenumber to get

\[
\hat{c} = c_H - U_0,
\]

(2.37)

where \( c_H \) is the observed (ground-based) horizontal phase speed, and

\[
U_0 = V_{z0} \cos \phi + V_{m0} \sin \phi,
\]

(2.38)

is the wind in the direction of wave propagation [e.g., Taylor et al., 1993; Wang and Tuan, 1988]. The dispersion relation (Equation 2.33) in terms of the vertical wavenumber squared, can be written as

\[
m^2 = \frac{N^2}{(c_H - U_0)^2} - k_H^2 - \frac{1}{4H_s^2}.
\]

(2.39)

If the intrinsic wave frequency is Doppler shifted to zero, then the intrinsic phase speed

\[
\text{Figure 2.8: Daily-averaged MERRA winds for 1 December 2011. (a) Meridional (black), and (b) zonal winds are compared with HWM14 (gray) winds. (c) Daily-averaged MERRA neutral temperatures (black) are compared with NRLMSISE-00 neutral temperatures (gray).}
\]
Figure 2.9: Illustration of a horizontal wave vector, \( \vec{k}_H \), in a horizontal wind field, \( \vec{U}_0 = (V_{z0}, V_{m0}) \), where \( \phi \) is an angle measured counterclockwise from East.

\( \hat{c} \rightarrow 0 \) as well, and Equation 2.39 will have a singularity where \( m^2 \rightarrow \pm \infty \). Hines and Reddy [1967] showed that under these conditions the wave energy can be strongly absorbed into the background medium with no significant transmission. This way of selecting waves is called critical layer filtering and can cause significant anisotropy in GW propagation headings in the upper atmosphere [e.g., Medeiros et al., 2003; Stockwell and Lowe, 2001b; Taylor et al., 1993; Wang and Tuan, 1988].

Waves that are filtered in this way can be investigated by considering when the intrinsic phase speed is zero (\( \hat{c} = 0 \)). This occurs when the observed phase speed (\( c_H \)) is equal to the wind in the direction of the wave (\( U_0 \)),

\[
c_H = V_{z0} \cos \phi + V_{m0} \sin \phi.
\]  
\tag{2.40}

The regions when \( c_H = 0 \) are when the wave has reached a critical layer and will not be able to propagate upwards. Equation 2.40 can be used to look at forbidden regions, also known as blocked regions, by taking \( V_{z0} \) and \( V_{m0} \) at each altitude and for each angle (0–360°) to calculate \( c_H \). Then for each angle, \( \phi \), take the maximum of \( c_H \), as a function of altitude, and this will correspond to the minimum observed phase speed the wave can have at that
angle before $\dot{c}$ goes to zero.

The example in Figure 2.10a considers three GWs with observed phase speeds of 10, 30, and 40 m/s propagating directly eastward being affected by the background winds corresponding to the total daily averaged HWM14 zonal winds (Figure 2.7b) and assuming the meridional winds are zero. The waves with phase speeds of 10 and 30 m/s will be absorbed into the background flow at altitudes of $\sim$11 and 30 km, respectively, while the wave with a phase speed of 40 m/s will not be affected by the background winds and will continue to propagate upwards. Figure 2.10b is the blocking diagrams for each altitude (gray lines) and the total blocking diagram (black line), illustrating that the wind blocking is towards the east. Also shown are the three phase velocities of each wave, again illustrating that the waves with phase speeds 10 and 30 m/s are well within the blocking region while the wave with phase speed 40 m/s is outside the blocking region.

2.7 Summary of Important Equations

This section lists equations from this chapter that are used throughout this dissertation.

![Figure 2.10](image)

**Figure 2.10:** Example wind blocking using HWM14 winds and three waves with different observed periods. (a) Daily-averaged HWM14 zonal winds, and (b) blocking regions for each altitude (gray lines) and total blocking region (black line), illustrating how the three waves are blocked.
The wave number in terms of wavelength ($\lambda$):

$$k = \frac{2\pi}{\lambda}.$$  \hfill (2.41)

The horizontal wavelength in terms of the zonal ($k_x$) and meridional ($k_y$) wave numbers:

$$\lambda_H = \frac{2\pi}{\sqrt{k_x^2 + k_y^2}}.$$  \hfill (2.42)

The ground-based phase speed in terms of the ground-based period ($\tau$):

$$c_H = \frac{\lambda_H}{\tau}.$$  \hfill (2.43)

Dispersion relation appropriate for the MLT region:

$$m^2 = \frac{N^2}{(c_H - u_0)^2} - k_H^2 \frac{1}{4H_s^2}.$$  \hfill (2.44)

Dispersion relation appropriate for the thermospheric region:

$$m^2 = \frac{k_H^2 N^2}{\omega_T^2 \left( 1 + \delta_+ + \delta^2 \frac{\delta_+}{\tau} \right)} \left[ 1 + \frac{\nu^2}{4\omega_T^2} \left( K^2 - \frac{1}{4H_s^2} \right)^2 \left( 1 - P \frac{1}{\frac{1}{2} + \frac{1}{4} \frac{\delta_+}{\tau} \left( 1 - \frac{\delta_+}{\tau} \right)^2} \right) \right]^{-1} - k_H^2 \frac{1}{4H_s^2}. \hfill (2.45)$$
CHAPTER 3
GRAVITY WAVE IMAGING INSTRUMENTATION AND ANALYSIS

3.1 Introduction

This chapter introduces one technique used to observe gravity waves (GWs) in the mesosphere lower thermosphere (MLT) region (∼80–110 km). The chapter begins with an introduction to the airglow emission layers that act as a tracer for GWs, followed by two optical instruments used in these studies. The chapter concludes with a description of how to prepare the image data for analysis and the analysis method to obtain horizontal GW parameters in the MLT region.

3.2 Airglow Emissions

Solar radiation incident on the Earth’s atmosphere is capable of exciting atmospheric atoms and molecules. At lower altitudes the atmosphere is very dense and the excited atoms are de-activated through collisions with other atoms or molecules before they can reradiate. However, in the upper atmosphere the density is much lower and the time between collisions is longer, therefore the excited atoms/molecules are able to radiate producing airglow emissions [e.g., Chamberlain, 1961].

Airglow emissions are categorized according to the processes that produce them. When sunlight is shining from above onto the emitting region, the airglow is termed dayglow. During the day, the atmospheric molecules are photoionized by the incoming solar radiation. It is difficult to observe dayglow because of the direct sunlight and strong diffuse background of Rayleigh scattered solar radiation. When sunlight is shining from below, or at an angle, onto the emitting region, the airglow is termed twilightglow. Twilightglow are easier to detect than dayglow; however, they have relatively short observing times [e.g., Taylor, 1986].

Nightglow are nighttime emissions that are produced indirectly by solar radiation. During the night, those atoms/molecules excited by direct solar radiation recombine, creating
optical emissions termed nightglow. Therefore, nightglow emissions are present during the day and twilight; however, they are overpowered by the other emissions. To the naked eye, nightglow emissions are faint in comparison to the moon and stars. However, there have been reports of bright nights where airglow emissions appeared visible to the naked eye with the earliest observation on record dating back to 1788 [e.g., Chamberlain, 1961].

The spectrum of visible and NIR nightglow emissions recorded by a spectrometer [e.g., Broadfoot and Kendall, 1968] are shown in Figure 3.1. There are several prominent emissions at visible wavelengths such as the OI (green line) emission (557.7 nm), and the Na emission (589.2 nm). The NIR range (>720 nm) is dominated by several strong OH molecular band emissions, which were first identified by Meinel [1950a,b]. In general, the dissociated atoms and molecules creating the nightglow emissions are created in relatively restricted altitude regions and appear as layers enveloping the Earth. Figure 3.2 shows a picture taken from the International Space Station looking back at the Earth’s limb. A thin nightglow layer is clearly evident due to the long line-of-sight integration through the emitting layer.

Sounding rocket experiments have determined the altitudes of these emission layers. The OH nightglow emission originates in a thin layer located near the mesopause region at

![Figure 3.1: Visible and NIR spectrum of the Earth’s nightglow emissions showing dominant atomic and molecular emission lines and bands. [Adapted from Broadfoot and Kendall [1968].]](image-url)
Figure 3.2: Image of the Earth’s limb and foreground taken from the International Space Station showing airglow emission layers. [NASA ISS photograph.]

an altitude of $\sim 87$ km. The thickness of this layer is $\sim 10$ km [e.g., *Baker and Jr*, 1988]. The Na emission originates at an altitude of $\sim 90$ km, the O$_2$ at $\sim 94$ km, and atomic OI at $\sim 96$ km, with similar layer thicknesses as the OH layer. Figure 3.3 shows the normalized volume emission rate as a function of altitude for these four nightglow emissions.

In Chapters 5 and 6, we focus on the bright NIR OH airglow emissions. We therefore briefly introduce the processes that create these emissions. *Bates and Nicolet* [1950] and *Herzberg* [1951] independently found the primary mechanism for the formation of OH in an excited state. This mechanism involves hydrogen and ozone, both minor species at mesospheric heights:

$$H + O_3 \rightarrow OH^* + O_2,$$

(3.1)

where $*$ denotes that OH is formed in an excited state. This mechanism is termed the ozone mechanism and forms part of the catalytic cycle for the destruction of ozone in the upper atmosphere. The hydrogen atom acts as a catalyst and is recycled through a reaction
The amount of energy released in the ozone reaction is enough to excite the OH molecule into the $v = 9$ vibrational state. The OH nightglow emissions results from the release of this excess energy as photons by cascading down through to the lower vibrational levels creating a series of overlapping molecular band emissions spanning the spectral range 140–4000 nm. The OH emission is the most intense nightglow emission and can be easily detected from the ground.

3.3 Airglow Imaging

As mentioned in Chapter 1, there are waves that exist in the atmosphere. These waves cannot be observed directly; however, their effects on the atmosphere, such as winds, temperatures, and densities (as in Chapter 2), can be observed. These induced oscillations
on the atmosphere can affect the reaction rates of the airglow emissions [e.g., Snively and Pasko, 2005], which result in variations in airglow intensity. Figure 3.4 shows an early example of wave-like structure photographed in the OH emission using sensitive infrared film [e.g., Peterson and Kieffaber, 1973a].

The ability to photograph wave structure in the airglow emissions has allowed numerous studies of GWs. The earliest measurements were made using photographic film [e.g., Armstrong, 1982; Moreels and Herse, 1977; Peterson and Kieffaber, 1973a,b], low-light television cameras [e.g., Crawford, 1978; Taylor et al., 1987; Taylor, 1986], and more recently charged couple device (CCD) detectors [e.g., Taylor and Hill, 1991]. With the development of all-sky imaging systems using telecentric optics, wave structure has been observed over the whole sky [e.g., Peterson and Adams, 1983; Swenson and Mende, 1994]. These GW studies rely heavily on the identification of relatively high-contrast wave patterns. The following two sections introduce two optical instruments used to study GWs in the MLT region.

![Figure 3.4: Early photograph taken from Capilla Peak, New Mexico (35° N, 106° W), showing wave-like structure in the NIR OH emission. [From Peterson and Kieffaber [1973a], with permission.]](image-url)
3.3.1 All-Sky Airglow Imager

The Mesospheric Airglow Imaging and Dynamics (MAID) project was initiated in January 2011 to investigate short-period GW dynamics over central Alaska. MAID is a collaborative project between Utah Valley University (UVU), Utah State University (USU), and the University of Alaska, Fairbanks (UAF). One of the goals of MAID was to establish a long-term climatology of short-period GWs in the high northern latitudes during the winter season.

In Chapter 5, measurements of short-period GW events from Poker Flat Research Range (PFRR), Alaska (65° N, 147° W) using an all-sky, multi-wavelength CCD imaging system will be discussed. Figure 3.5 shows the location of PFRR, labeled PF, represented as the red dot and Figure 3.6 shows the T. Niel Davis science building along with the all-sky imager. The imager is a state of the art system designed to remotely sense several airglow emissions primarily in the MLT region by utilizing a fast three-inch telecentric all-sky optical system mapped onto a sensitive 1024×1024 pixels, back-illuminated CCD. The system is equipped with several three-inch, two nm full width half max (FWHM) narrow

![Figure 3.5: Map of the Northern Hemisphere showing the locations of PFRR (labeled PF, red dot), ALOMAR (labeled AL, blue dot), Resolute Bay (labeled RB, black dot), and Svalbard (labeled SV, orange dot).](image-url)
band interference filters to sequentially monitor the \( \text{O}_2 \) (865.5 nm), Na (589.2 nm), and the OI (630.0 nm) line emissions, while a broadband filter is used to observe the NIR OH (715–930 nm) emission. The bright OH emission is captured using a fast exposure time of 15 seconds while the fainter \( \text{O}_2 \), Na, and OI emissions require exposure times of 120, 90, and 120 seconds, respectively [e.g., Taylor et al., 1993]. The acquisition sequence used in this study was OH–\( \text{O}_2 \)–OH–Na–OH–OI yielding a fast \( \sim 2 \) min cadence for the OH data and \( \sim 6 \) min cadence for the remaining emissions. The image data are processed on chip by \( 2 \times 2 \) binning resulting in a \( 512 \times 512 \) data product. Figure 3.7 illustrates a sequence of imager data from the PFRR MAID all-sky imager (ASI from now on), observed in the four emissions monitored on 14 January 2012. On this particular night, a prominent wave event filled the majority of the field of view (FOV), and persisted throughout the entire observation period (>15 hours). The wave is clearly evident in the OH (Figure 3.7a) and Na emissions (Figure 3.7b).

Optical measurements of GWs in the Polar MLT region are difficult to accomplish due to the frequent presence of aurora that can be orders-of-magnitude brighter than the fainter nightglow emissions. Furthermore, presence of tropospheric clouds often obscure the viewing conditions.

Figure 3.6: (a) T. Niel Davis science building at the Poker Flat Research Range (PFRR), and (b) the PFRR MAID all-sky airglow imager.
3.3.2 Advanced Mesospheric Temperature Mapper

In Chapter 6, observations of short-period GWs using an Advanced Mesospheric Temperature Mapper (AMTM) located at the Arctic Lidar Observatory for Middle Atmospheric Research (ALOMAR) in northern Norway ($69^\circ$ N, $16^\circ$ E) will be discussed. Figure 3.5 shows the location of ALOMAR, blue dot labeled AL, while Figure 3.8 shows ALOMAR and the AMTM. The AMTM, like the ASI described in Section 3.3.1, ran continuously during the winter nights for one winter season, October 2011 to March 2012. However, unlike the ASI, the AMTM was designed for high-latitude research and utilizes recent progress in IR imaging capability to observe the OH (3,1) band emission at $\sim$1.55 $\mu$m, which is over 70 times stronger than the submicron OH emissions typically observed by CCD imaging systems making the auroral contamination more limited. This enabled the AMTM to operate even under strong auroral conditions with only limited auroral contamination.
Figure 3.8: (a) Artic Lidar Observatory for Middle Atmosphere Research (ALOMAR) (https://commons.wikimedia.org), and (b) the Utah State University Advanced Mesospheric Temperature Mapper infrared imaging system.

The AMTM was developed at USU Space Dynamics Laboratory and uses a 120° FOV telecentric lens system, and three 4 inch diameter, 2.5–3 nm interference filters centered at 1523 nm \( P_1(2) \), and 1542 nm \( P_1(4) \) and at a nearby background region (BG) [e.g., Pautet et al., 2014]. The exposure time for each filter is 10 s. Figure 3.9 shows examples of AMTM observations of the \( P_1(2) \) (Figure 3.9a), and \( P_1(4) \) (Figure 3.9b) emissions on 16 December 2011. These data are combined to make OH temperature maps (Figure 3.9c) and an OH (3,1) band radiance map (Figure 3.9d) every \( \sim 30 \) s. The AMTM operates automatically from dusk to dawn taking data in the presence of aurora and also under full-moon conditions, which significantly increases the observing time available for investigating GWs in the MLT at high latitudes.

3.4 Airglow GW Data Analysis

For the PFRR MAID ASI and ALOMAR AMTM studies, extensive analysis will focus on distinctive wave events measured in the strong NIR OH emissions. Wave events were identified through visual inspection of the OH images. Band structures (Figure 3.7a) are identified, and are seen to propagate across an image as time progresses. An event starts
Figure 3.9: Raw images obtained from the ALOMAR AMTM on 16 December 2011. (a) $P_1(2)$ band emission, (b) $P_1(4)$ band emission, (c) temperature map, and (d) OH band intensity.

when the band structure appears in the image, and ends either when the band structure is no longer visible, is obscured by tropospheric clouds, or if the aurora becomes too intense (an issue only for the PFRR ASI).

The images are prepared for analysis through a three-step process as illustrated in Figure 3.10. First, the raw image (Figure 3.10a) is rotated to properly align the cardinal directions using the background star field and the observation site information (Figure 3.10b). Second, the star field is removed by assuming an individual star to be a point spread function and approximate the star pixel value to the average of the neighboring pixel values (Figure 3.10c). Third, the image is transformed to uniformly spaced $500 \times 500$ km$^2$ geographic coordinate system, for the PFRR ASI, through a process commonly referred to as unwarping (Figure 3.10d). Image preparation for the AMTM data were similar to the ASI except the images were unwarped to $250 \times 250$ km$^2$ geographic coordinates.
The identified wave events have been analyzed to determine their observed (ground-referenced) horizontal wave parameters: wavelengths, phase speeds, periods, and directions of propagation using well-established two-dimensional Fourier analysis methods and spatial mapping techniques [e.g., Coble et al., 1998; Garcia et al., 1997]. First, the unwarped images are analyzed to determine the observed (ground-based) horizontal phase speeds, $c_H$, by tracing prominent wave fronts sequentially in time. To determine the wavelength of the GW event, a region is selected that contains multiple wave fronts, as illustrated in Figure 3.11a as a dashed red line. A 2-D Fourier analysis on the selected region yields a 2-D horizontal wavenumber spectrum as illustrated in Figure 3.11b. Dominant peaks in wavenumber spectrum yield the horizontal wavenumber, as well as the wave azimuth. Figure 3.11b shows the ambiguous wavenumber spectrum, that exhibits two peaks off of
the center. In order to choose the correct peak, visual inspection of a series of successive images are used to determine the direction of propagation, in this case the phase fronts are observed to be moving towards the East (as indicated by the red arrow in Figure 3.11a). This means the peak with the yellow arrow in Figure 3.11b is chosen as the peak for this particular event. Here the horizontal wavelength was $\lambda_H \sim 22$ km. The period of the wave is derived using $\tau = \frac{\lambda_H}{c_H}$, and was found to be $\sim 10$ min. This analysis was performed at least three times on a set of images before the next set of images were analyzed.

Using the example event on 14 January 2012, observed from $\sim 3$–8:30 UT, and performing the analysis method described above every $\sim 30$ min, yields the horizontal wave parameters as a function of time, as illustrated in Figure 3.12. Here the observed periods ranged from $\sim 12$–26 min and exhibited a time averaged period of $\sim 17$ min (Figure 3.12a). The horizontal wavelengths ranged from $\sim 25$–53 km, with a time average of $\sim 36$ km. The time averaged phase speed was found to be $\sim 35$ m/s and propagated with an azimuth of $\sim 83^\circ$ (eastward), and remained constant throughout. The time-averaged values for this particular event categorize this as a short-period GW event.
The Saura Doppler radar was installed at ALOMAR in July 2002 to complement the many experiments available at ALOMAR. The radar operates between 2–3 MHz allowing continuous monitoring of the mesosphere between 50–95 km altitude throughout the year, enabling wind and turbulence measurements, meteor observations (during nighttime only), and electron density measurements [e.g., Singer et al., 2008]. For our ALOMAR AMTM study, we have used Saura meteor radar observations to obtain the horizontal winds in the MLT region (80–110 km) encompassing the OH layer, during the night times. Figure 3.13a-
show example zonal and meridional winds, respectively, for the night of 8 December 2011. The meridional and zonal winds were separately smoothed in time by applying a three-hour running average, as shown in the figure (positive zonal winds are eastward and positive meridional winds northward). A strong \( \sim 12 \) hr wave exhibiting downward phase progression was evident in both the radar wind and lidar temperature data at this time, probably due to the semi-diurnal tide.

The neutral winds are used to determine the intrinsic periods and phase speeds, using Equations 2.37 and 2.18 in Chapter 2. To illustrate the analysis method, we present data from a single GW event observed over two hours on this occasion (15:20 to 17:44 UT). The average observed wave parameters for this short-period wave event are summarized in Table 3.1. The Saura neutral winds for the intrinsic wave analysis (Figure 3.13a,b) were limited to the time of the occurrence of the event, and were averaged in time to yield a single zonal (Figure 3.14 solid black) and meridional wind profile (Figure 3.14 dashed black). The data were then limited in altitude to 85–95 km to encompass the OH layer. Using the wave azimuth, \( \phi \), the wind in the direction of the wave, \( u_0 \), (Figure 3.14a blue) was calculated via

\[
u_0 = u_z \cos \phi + u_m \sin \phi,
\]

where \( u_z \) is the zonal wind and \( u_m \) is the meridional wind. Once the wind in the direction of the wave was obtained, the intrinsic phase speed, \( c_I \), with altitude (Figure 3.14a red line)

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Mean</th>
</tr>
</thead>
<tbody>
<tr>
<td>( \lambda_H ) [km]</td>
<td>19 ± 1</td>
</tr>
<tr>
<td>( c_H ) [m/s]</td>
<td>40 ± 1</td>
</tr>
<tr>
<td>( \tau ) [min]</td>
<td>8.0 ± 2</td>
</tr>
<tr>
<td>( \phi ) [deg]</td>
<td>51° ± 2°</td>
</tr>
<tr>
<td>( c_I ) [m/s]</td>
<td>36 ± 1</td>
</tr>
<tr>
<td>( \tau_I ) [min]</td>
<td>9 ± 1</td>
</tr>
<tr>
<td>( \lambda_z ) [km]</td>
<td>15 ± 1</td>
</tr>
</tbody>
</table>
Figure 3.13: ALOMAR Saura radar winds observed on 8 December 2011. (a) Zonal, and (b) meridional winds obtained every 30 min with a 3-hr running average applied.

was calculated via

\[ c_I = c_H - u_0, \]  
\[ (3.4) \]

where \( c_H \) is the ground-based phase speed. The intrinsic phase speed over the altitude range 85–95 km capturing the OH layer was then given by the altitude averaged phase speed, and determined to be \( \sim 36 \) m/s for this particular event. The intrinsic period, \( \tau_I \), was then derived using

\[ \tau_I = \frac{\lambda_H}{c_I}, \]
\[ (3.5) \]

where \( \lambda_H \) is the observed horizontal wavelength. For this event, the average intrinsic period was \( \sim 9 \) min, as listed in Table 3.1.

3.6 Vertical Wavelengths

The vertical wavelengths for the individual events were calculated using the intrinsic wave parameters applied to the dispersion Equation 2.24,

\[ m^2 = \frac{N^2}{c_I^2} - k_H^2 - \frac{1}{4H_s^2}, \]
\[ (3.6) \]

which depends on the horizontal wave number, \( k_H \), and the intrinsic phase speed. This equation also requires knowledge of the scale height, \( H_s \), and buoyancy frequency, \( \tau_B \), in
Figure 3.14: Illustration of the steps used to determine the intrinsic phase speed (red line) using the ALOMAR Saura radar wind data (blue line) from the background meridional and zonal wind components (dashed and solid black). Data were averaged over 85–95 km to encompass data in the OH emission layer.

In addition to Saura radar data, some observation nights had sodium (Na) lidar temperature data available. The Weber sodium lidar is a resonance fluorescence lidar that measures winds, temperatures, and sodium densities from \( \sim 80–105 \) km [e.g., She et al., 2002]. Figure 3.15a shows an example of the 3-hr smoothed Na lidar temperatures from \( \sim 80–105 \) km for our example event, and plot b shows the nightly average.

The Saura neutral winds for the intrinsic wave analysis (Figure 3.13a,b) were limited to the time of the occurrence of the event, and were averaged in time to yield a single zonal (Figure 3.14 solid black) and meridional wind profile (Figure 3.14 dashed black). The data were then limited in altitude to 85–95 km to encompass the OH layer.

The Brunt-Väisälä period, \( \tau_B \), and scale height, \( H_s \), were calculated using Equations 2.8 and 2.16 developed in Chapter 2. Figure 3.16a and b, plot the derived Brunt period and scale height as a function of altitude (85–95 km). The Brunt periods varied in altitude from 4.5 to 5.5 min with a mean value of 5.2 min. Note, the typical value for \( \tau_B = \frac{2\pi}{N} \) assumed for dry air is \( \sim 5 \) min period \( (N = 0.02 \text{ s}^{-1}) \) for the MLT region, which agrees very well.
Figure 3.15: ALOMAR Na lidar temperatures observed on 8 December 2011. (a) Temperatures with a 3-hr smoothing applied, and (b) nightly averaged temperatures.

with our results. The altitude averaged scale height was \(\sim 6.3\) km, and exhibited a maxima of \(\sim 6.4\) km at \(\sim 87\) km. Typical values of the scale height for the MLT region are assumed to 7 km, which agrees well with our example event derivations.

Using these data, the derived vertical wavenumber squared and vertical wavelength as a function of altitude for this event are shown in Figure 3.16c,d respectively (solid lines). The vertical wavelengths were observed to decrease in magnitude with altitude, from \(\sim 14\) km at 85 km to \(\sim 8\) km at 95 km altitude with an altitude averaged wavelength of \(\sim 15\) km, which is a good representative value encompassing the OH layer [e.g., Hecht et al., 2001; Nielsen et al., 2012; Taylor et al., 1995; Walterscheid et al., 1999].

Since Na lidar temperatures were not available for every night of observation, we have investigated the effects of using a constant value for the Brunt-Väisälä frequency \((N=0.02\) s\(^{-1}\)) and scale height \((H_s = 7\) km) on the resultant vertical wavenumber squared (Figure 3.16c dashed black line) and vertical wavelength, as shown in Figure 3.16c,d, respectively. In each case the dashed black line plots the results for these constant values. The results clearly show that using fixed values of \(N\) and \(H_s\) provide a good representation, in this case yielding an altitude average \(\sim 15\) km, which is the same as using altitude varying \(N\) and \(H_s\).

We further investigate the effects of using constant values of \(N\) and \(H_s\) by employing the horizontal wave parameters in Table 3.1 and varying \(N\) from \(\pm 70\%\) of 0.02 s\(^{-1}\) and \(H_s\)
Figure 3.16: Derived profiles of the (a) Brunt-Väisälä period, (b) scale height, (c) vertical wavenumber squared, and (d) vertical wavelength using Saura radar winds and Na lidar temperatures. The dashed lines in (c–d) are the results of utilizing constant values of $N=0.02 \text{ s}^{-1}$ and $H_s=7 \text{ km}$, revealing good agreement.

from $\pm 70\%$ of 7 km. The results are illustrated in Figure 3.17, which plots $\tau_B$ along the y-axis, $H_s$ along the x-axis, and $\lambda_z$ as color contours. This figure shows that the vertical wavelength calculations are relatively insensitive to realistic variations in the scale height and Brunt period. We therefore conclude that using constant values of $N$ and $H_s$ will allow us to gain a good estimate of the distribution of vertical wavelengths at ALOMAR.
Figure 3.17: Testing the sensitivity of using constant values of the Brunt-Väisälä frequency and scale height by changing their values from \( \pm 70\% \) using wave parameters in Table 3.1 and calculating the altitude averaged vertical wavelengths.
CHAPTER 4  
RADAR INSTRUMENTATION AND MSTID ANALYSIS

4.1 Introduction

In Chapter 3, two optical instruments used to observe gravity waves (GWs) in the mesosphere lower thermosphere (MLT) region (∼80–110 km altitude) were described. The imagers used airglow emissions located in the MLT as a tracer for GWs, and the ground-based horizontal wave parameters were extracted using an example event.

The thermosphere does contain an airglow emission layer (atomic oxygen) at an altitude of ∼250 km, and studies of GWs using airglow imagers have been performed previously [e.g., Kubota et al., 2011; Paulino et al., 2016]. However, as mentioned earlier, airglow imagers are restricted to nighttime observations, and thus limited to winter nighttime studies in the high-latitude regions. Radars provide another technique to observe GWs in thermosphere, by using the ionosphere as a passive tracer for GWs in the thermosphere.

In this chapter, incoherent scatter radars are introduced, as they provide a wealth of information about the ionosphere. In particular, the Poker Flat Incoherent Scatter Radar (PFISR) is used to observe the ionosphere over central Alaska, providing observations of medium-scale traveling ionospheric disturbances (MSTIDs). The chapter concludes with a technique developed to extract horizontal and vertical wave parameters using the multi-beam measurements of PFISR.

4.2 Doppler Radar

To better understand ISRs, we first need to understand the basics of Doppler radar. Radio detection and ranging (RADAR) uses electromagnetic waves to detect remote targets. A radar is capable of determining the position and velocity of the target using the target range, angles, and Doppler shift. The range of a target is the distance from the target to the radar. Target angles can be found using the elevation (angle from the horizon up to the direction of the radar beam) and azimuth (measured from the direction of geographic
north to the direction of the radar beam) angles of the directed radar beam. The position of the target can be written as

\[
\vec{r} = R [(\cos \theta \sin \phi) \hat{x} + (\cos \theta \cos \phi) \hat{y} + \sin \phi \hat{z}],
\]  

(4.1)

where \( R \) is the range, \( \theta \) is the azimuth angle and \( \phi \) is the elevation angle. Radars transmit waves, modulated by an on-off action of a pulse envelope allowing the radar to survey various ranges. The Doppler shift is used to obtain the line-of-sight (LOS) velocity of the target.

As an example of how to resolve the Doppler frequency, suppose a simple cosine wave is transmitted with a frequency, \( f_0 \). The returned signal will be Doppler shifted with a Doppler frequency \( f_D \), written as

\[
\cos(2\pi(f_0 + f_D)t).
\]  

(4.2)

To resolve \( f_D \), we can multiply the original signal by the Doppler shifted signal (Equation 4.2), known as mixing, and filter out the original frequency \( f_0 \). Performing a Fourier transform on the mixed filtered signal and looking at the power spectrum, there will be two peaks, one peak at \(-f_D\) and one at \(+f_D\). By mixing the original signal with Equation 4.2, information about the sign of \( f_D \) is lost. Therefore, we would like the mixed filtered signal to look like

\[
\exp(i2\pi f_D t) = \cos(2\pi f_D t) + i \sin(2\pi f_D t),
\]  

(4.3)

because the Fourier transform of this equation is a Dirac delta function centered at a single frequency \( f_D \). Signals like Equation 4.3 cannot be measured directly, but can be created by mixing the returned signal with cosine and sine waves using the original frequency \( f_0 \). The real part of Equation 4.3 is known as the in-phase component (I) and the imaginary part is known as quadrature (Q) or the out-of-phase component. Thus, the fundamental output of a pulsed Doppler radar is a time series of complex numbers. If there is a distribution of targets moving at different velocities, such as electrons in the ionosphere, then there is no
single Doppler shift, but rather, a Doppler spectrum.

4.3 Incoherent Scatter Radar

The ionosphere was introduced in Chapter 1 and consists of free electrons and ions. There are many observational techniques used to observe the ionosphere (see Chapter 1, Section 1.4). Incoherent scatter radars (ISR), which are Doppler radars, are an excellent tool to measure properties of the ionosphere. Gordon [1958] is credited with the idea of the first ISR. He predicted that if a powerful beam of radio waves with a frequency well above the penetration frequency were sent vertically through the ionosphere, a very small, but still measurable amount of power would be scattered back to the ground from the randomly distributed free electrons in the ionosphere. The basic physical mechanism underlying the operation of ISRs is Thomson scattering of electromagnetic waves by ionospheric free electrons [e.g., Kudeki and Milla, 2012]. The electric field of the incident waves from the ISR causes the electrons to accelerate, which in turn radiate at a signal at almost the same frequency. The electrons are in constant thermal motion in the ionosphere, and hence, the radiation is Doppler shifted from the incident frequency [e.g., Kelly, 1989].

Gordon [1958] also pointed out that a study of the incoherent scatter signal would provide information about the density and temperature of the electrons not only below the level of maximum density in the F region, as was with conventional sounding methods, but also above this level, to heights of several thousand kilometers or more [e.g., Dougherty and Farley, 1960]. Figure 4.1a shows an example of the expected power spectrum if indeed the transmitted signal was scattered off free electrons.

Bowles [1958] executed the Gordon [1958] idea using a large transmitter and antenna and took a few measurements. He was able to find the variation of electron density over a wide range of heights using a pulse technique. Bowles [1958] did not observe the broadening predicted by Gordon [1958], instead he observed a much narrower double-humped spectrum, as illustrated in Figure 4.1b. The reason for this is because the much heavier and slower ions are controlling the scattering spectral width and not the lighter, faster electrons. Therefore, although the electrons are doing the scattering, it is the ions that are controlling the shape
Figure 4.1: Example power spectra, in arbitrary units, if (a) the electrons were independent from the ions, and (b) the electrons are dependent on the ions. [Modified from Kudeki and Milla, 2011, their Figure 3].

of the spectrum, and a wealth of information about ionospheric charged particles can be extracted from the measured power spectra of ISRs [e.g., Kudeki and Milla, 2011].

An ISR measures a time series of complex voltages as a function of range. An autocorrelation function (ACF) is constructed by taking products of complex voltages. An ACF is a temporal correlation, which compares two functions to see how similar they look at different times. The products of the complex voltages are known as lag products. Doing a Fourier transform on a lag product produces the ISR power spectrum, which can be fit for common incoherent scatter parameters such as electron density, electron temperature, ion temperature, and the line-of-sight (LOS) ion velocities [e.g., Evans, 1969].

4.4 Poker Flat Incoherent Scatter Radar

In Chapter 7, we use data obtained from the Poker Flat Incoherent Scatter Radar (PFISR), shown in Figure 4.2, to make observations of GWs in the thermosphere using the ionosphere as a passive tracer. PFISR is operated at the Poker Flat Research Range (PFRR) (65.13° N, 147.47° W, MLAT=65.4° N) near Fairbanks, Alaska. PFISR is a part of the Advanced Modular Incoherent Scatter Radar (AMISR) class of ISRs and has the capability to rapidly observe different regions of the ionosphere using pulse-to-pulse beam
steering. This makes PFISR an ideal instrument for investigating the three-dimensional properties of TIDs at high latitudes [e.g., Nicolls and Heinselman, 2007; Vadas and Nicolls, 2008; Waldock and Jones, 1986].

Figure 4.3 shows the location of PFISR in central Alaska. PFISR is capable of measuring 473 preprogrammed look directions within the grating lobe limits [e.g., Heinselman and Nicolls, 2008], as indicated in Figure 4.3. The individual beam positions are selectable as determined by the type of ionospheric observations to be investigated. The insert shows the latitudinal and longitudinal ranges (in km), with the vertical axis pointing northwards and the horizontal axis pointing eastwards, of the grating lobe limit plotted at 300 km altitude. The four-beam configuration used primarily for this study is indicated by the black dots.

The four-beam configuration we use here was introduced for the International Polar Year (IPY) that began on 1 March 2007. The primary goal of IPY was to obtain high-quality ionospheric data with good spatial coverage on a near continuous basis, and was run when no other special programs were in operation [e.g., Sojka et al., 2009]. PFISR IPY data for this study can be obtained through Open Madrigal. This mode was also operated at lower power, which resulted in a maximum working altitude range of ∼300–500 km, depending on the ionospheric conditions. The four-beam configuration was comprised of

Figure 4.2: The Poker Flat Incoherent Scatter Radar (PFISR) located at PFRR. [From http://amisr.com.]
Figure 4.3: Map showing the location of PFISR in central Alaska (labeled PFRR) and the latitudinal and longitudinal extent of the PFISR grating lobe limit and IPY beam configuration mapped at 300 km altitude.

A zenith-pointing beam (elevation (el) = 90°, azimuth (az) = 14°), a beam pointing up the magnetic field line (el = 78°, az = −154°), and two off-zenith beams to the north-northwest (NNW) (el = 66°, az = −35°) and to the east-northeast (ENE) (el = 66°, az = 75°), shown as points 0, 1, 2, and 3 in Figure 4.3 insert. While this four-beam configuration is less than the number of beams employed by Nicolls and Heinselman [2007] in their earlier MSTID study using PFISR, this configuration provides the necessary zenith and spatial sampling needed to clearly resolve the three-dimensional horizontal wave vectors (as described in Section 4.5.2) [e.g., Nicolls and Heinselman, 2007; Vadas and Nicolls, 2009; Waldock and Jones, 1986].

PFISR was operated near continuously over a ~32 month period from August 2010 to April 2013, thereby obtaining 428 days of observations. From August 2010 to Decem-
ber 2011, most of the data were obtained using the IPY mode; over ∼75% of the MSTIDs reported in this investigation were measured using this four-beam configuration. From January 2012 to April 2013, IPY measurements were also made, but less frequently because of other program operations not suitable for our study.

4.5 PFISR Data Analysis

4.5.1 MSTID Identification and Localization

Our analysis is based on the case study method employed by Nicolls and Heinselman [2007] who used PFISR in a 10-beam mode, but adapted to the four-beam IPY mode. It also includes additional checks to ensure clear MSTID identification within this large data set. We illustrate our analysis method with a well-defined MSTID event observed on 18 May 2011. Electron densities were measured by each beam as a function of time and range. Figure 4.4a shows the electron density profile for the vertically pointing beam for this event. Close visual inspection of these data reveal clear evidence of a vertically extensive, periodic ionospheric perturbation. This high frequency event was observed from ∼11–18 UT and was coherent over the altitude range of ∼100–350 km, as indicated by the dashed area in Figure 4.4a. Also note that at ∼100–200 km around 10–12 UT, there is a depleted electron density region associated with auroral upwelling, which can limit the ability to make accurate nighttime MSTID measurements.

To investigate this event, we employ a low-pass $6^{th}$ order Butterworth filter to remove the electron density perturbations with periods <120 minutes. This result is shown in Figure 4.4b. The relative electron density perturbations are then calculated via

$$\frac{\delta N_e}{N_{e_0}} = \frac{N_e - N_{e_0}}{N_{e_0}},$$

(4.4)

where $N_e$ is the measured electron density and $N_{e_0}$ is the filtered electron density. Figure 4.4c shows the relative electron density perturbations for this event. High frequency waves are clearly seen up to ∼300–350 km.
Figure 4.4: Illustration of the PFISR data preparation procedure using electron densities obtained on 18 May 2011. (a) Measured electron densities from the vertically pointing beam, (b) background electron densities estimated using a low-pass filter, and (c) the derived relative electron density perturbations.

Due to the large amounts of data, we have developed a method to detect and isolate the occurrence and periodicities of a broad-range of MSTIDs in the data set. This involves constructing spectrograms of the data for every PFISR run, and utilizing an uncorrected long-pulse measurement (∼5 km altitude resolution). This procedure was performed for each beam in each individual experiment. Figure 4.5 shows a spectrogram for the zenith-pointing
beam using our example event (from Figure 4.4). Each spectrogram is constructed by taking a subset (i.e., window) in altitude (in 25 km steps) and in time (120 min intervals) of the derived relative electron density perturbations. This procedure is typically performed over the altitude range 100–500 km. For each resolved altitude in the window, a Lomb Scargle (LS) [Press et al., 2007] spectral analysis is performed. The resultant LS powers are then averaged to yield a single power spectrum for that window, and are assigned a time stamp corresponding to the center of the time interval. The window is then incremented by 20 min and the process is repeated until the time series ends. This analysis is then repeated for the subsequent altitude bins, resulting in the spectrogram shown in Figure 4.5. Vertically-extensive, quasi-coherent wave events are then identified visually from the spectrogram via choosing events with discrete frequency signatures that exist in multiple consecutive altitude bins having similar durations. Our example MSTID event (from Figure 4.4) is clearly seen extending in altitude from 175–350 km and from $\sim$10–20 UT in Figure 4.5. In addition to identifying MSTID events, the spectrograms are also used to estimate the dominant periodicity of the event (in this case, $\sim$55 min).

Once an MSTID event is identified in a specific altitude range and duration, the relative electron density perturbations for all of the beams are limited to this same altitude range and duration. Figure 4.6a shows the derived relative electron density perturbations for the MSTID event described above for the zenith-pointing beam. The constant phase lines of the MSTID slope downwards in time, indicative of upward energy propagation for GWs; this is consistent with a lower atmospheric GW source. Finally, the relative electron densities are band-passed filtered using a 6th order Butterworth filter to further enhance the phase structures (Figure 4.6b), as done by Nicolls and Heinselman [2007]. For each event, the filter is centered on the dominant period of the MSTID. The filter bandwidth is chosen to contain the dominant wave periods present in the spectrogram. The cutoff periods of the bandpass filter are the same for all beams and at all altitudes for a particular event. This technique is successful in identifying MSTIDs in the thermosphere/ionosphere, and closely follows previously-published methods [e.g., Vadas and Nicolls, 2008]. Note that the
Figure 4.5: Example spectrogram for the MSTID event observed on 18 May 2011 for the PFISR zenith-pointing beam. Altitude bins of 25 km from 125–500 km altitude from 10–24 UT are shown. The box identifies the MSTID event.

The spectrogram method was first referenced in Georges [1968] and utilized by Nicolls et al. [2014] to identify GWs at the Arecibo Observatory.

4.5.2 Determination of TID Parameters

The wave analysis we use here is well developed, and has been employed in several prior studies [e.g., Nicolls and Heinselman, 2007; Reid, 1986; Waldock and Jones, 1986]. For each beam the electron density perturbations at each altitude are used to calculate the LS power spectrum and data with peak powers exceeding a >95% confidence level are subsequently
used. For the IPY data, we use the following beam pairs for analysis: (0,1), (0,2), (0,3), (1,2), (1,3), and (2,3). If the dominant periods from each beam in a beam pair differed by more than 10 minutes, then data from this beam pair at this altitude is also not used. These selection criteria have proved necessary to effectively identify dominant periodicities as a function of altitude. We perform this procedure automatically for all altitudes and for each beam pair.

The selected data are then used to calculate the complex cross-spectra using relative electron density perturbations from each beam pair. Briefly, the complex cross-spectra is obtained by taking the Fourier transform of one signal from one beam and multiplying it with the Fourier transform of the second signal from the other beam [e.g., Press et al., 2007]. We then compare the peak period of the spectrum with the dominant periods of each beam separately. If these two periods differed by more than 10 minutes, then this beam pair was also not used. This procedure was repeated for each beam pair to create a final data set.

In the process of obtaining the dominant period ($\tau$) from the cross-spectrum, the phase delay ($\gamma$) is also obtained. We then utilize the phase delays as a function of altitude

![Figure 4.6](image)

**Figure 4.6**: For the example MSTID event observed on 18 May 2011 using PFISR, the limited (a) derived relative electron density perturbations, and (b) band-passed perturbations are shown for the PFISR zenith-pointing beam.
to perform a selection test; in particular, we require the phase delay between two beam measurements to be consistent (see Figure 4.7). This is because the phase delay relates to the time ($\delta t$) that the MSTID takes to propagate from one beam to the other, and to the dominant period of the wave [e.g., Nicolls and Heinselman, 2007]:

$$y = \frac{2\pi \delta t}{\tau}.$$  \hspace{1cm} (4.5)

Errors in the phase delays are obtained by taking the difference between phase delays at the FWHM of the complex cross-spectrum (centered at the dominant period of the wave). A successful measurement requires at least two beam pairs at each altitude. This resultant data set is then used to compute the wave parameters as a function of time and altitude, as described below [e.g., Nicolls and Heinselman, 2007].

We calculate the vector $\vec{A} = (A_x, A_y, A_z)$, which points from one beam measurement to the other, using the measurement azimuth, elevation, and range (Equation 4.1). If we had only one beam pair and only compared two measurements, the wave vector can be

![Figure 4.7: Illustration of the phase delays as a function of altitude (175–350 km) using correlations with PFISR beams 1 and 0 for the case on 18 May 2011. The solid line shows a second-order polynomial least squares fit to the data.](image)
obtained using the phase delay,

$$\vec{A} \cdot \vec{k} = y, \quad (4.6)$$

where $\vec{k}$ is the wave vector. For $n$ measurements, the matrix form of Equation 4.6 can be written as:

$$
\begin{bmatrix}
A_{1x} & A_{1y} & A_{1z} \\
A_{2x} & A_{2y} & A_{2z} \\
\vdots & \vdots & \vdots \\
A_{nx} & A_{ny} & A_{nz}
\end{bmatrix}
\begin{bmatrix}
k_x \\
k_y \\
k_z
\end{bmatrix}
= 
\begin{bmatrix}
y_1 \\
y_2 \\
\vdots \\
y_n
\end{bmatrix}. \quad (4.7)
$$

Using a minimum mean square estimator via casting Equation 4.7 into a Bayesian linear model (which uses errors in the phase delays along the diagonal in the error covariance matrix), we solve for the wave vector, $\vec{k}$, via

$$
\vec{k} = (A^T C^{-1} A)^{-1} A^T C^{-1} y, \quad (4.8)
$$

where $C$ is the error covariance matrix composed of errors in the phase delay along the diagonal [e.g., Heinselman and Nicolls, 2008; Nicolls and Heinselman, 2007]. This method allows us to estimate the errors in the wavenumbers. Figure 4.7 shows the phase delays from correlations between the zenith-pointing beam and the off-zenith beam (beams 0 and 1) over the altitude range 175–350 km, where the signal was strongest. Again, we use the example event on 18 May 2011 to illustrate the analysis method. The solid line shows the least squares fit to the data. These results are typical of our analysis, and show phase delays that change with altitude in a consistent manner, within the limits of the measurements.

The results for the analysis of the 18 May 2011 event yield the observed periods (Figure 4.8a) and the horizontal wave vector (not shown) as functions of altitude. The horizontal wavelength (Figure 4.8b) is then calculated from the horizontal wavenumbers ($\lambda_H = 2\pi/\sqrt{k_x^2 + k_y^2}$). The phase speeds (Figure 4.8c) are derived using the observed period and the horizontal wavelength ($c_H = \lambda_H/\tau$). The direction of propagation is determined from the horizontal wave vector ($\phi = \tan^{-1} (k_y/k_x)$), shown in Figure 4.8d in
Figure 4.8: Derived horizontal wave parameters for the 18 May 2011 MSTID event observed with PFISR. (a) Periods, (b) horizontal wavelengths, (c) horizontal phase speeds, and (d) azimuths. The error bars represent the derived uncertainties for the individual wave measurements using the method of Nicolls and Heinselman [2007].

degrees clockwise from North (i.e., azimuth). Note that the full wave vector (i.e., including the vertical wavenumber) is not determined at this stage because correlations between adjacent beams are performed at similar altitudes. Here, we use similar altitudes in order to avoid possible aliasing and other geometrical effects [e.g., Nicolls and Heinselman, 2007].

For this event, the wave period increases from ∼55 min at 175 km to ∼60 min at 300 km altitude, with an altitude averaged value of 57 ± 1 min, where the error corresponds to the standard deviation of the mean. The horizontal wavelength also increases from ∼260 km to ∼400 km over the measured altitude range, with an altitude average of 372 ± 18 km. The derived phase speeds increase from ∼80 m/s to ∼110 m/s, with a mean of 108 ± 5 m/s. The direction of propagation changed with altitude from ∼10° to ∼50° at 300 km altitude, with a mean of 29°. These values clearly identify this event as a part of a MSTID wave
packet. Here we say wave packet because the constituent waves can propagate horizontally and vertically at the same time. However, as the observed parameters (such as period and phase speed) vary smoothly with altitude, we can conclude that the GWs we sample with altitude are from the same coherent wave packet excited by the same source. Therefore, the average values we calculate are the average values for this wave packet at this altitude and time.

4.5.3 Vertical Wavelengths

So far, we have investigated the horizontal parameters of the waves as a function of altitude. As noted in the previous section, the vertical wave number spectrum has not been determined yet since the analysis method would potentially introduce spatial aliasing. There are several different methods in the literature previously used to determine the vertical wavelength such as spatial spectral analysis using the vertically pointing beam. However, we have continued with our adapted method of Nicolls and Heinselman [2007] to estimate the vertical wavelength as a function of altitude using a correlation analysis (see Section 4.5.2) applied to the vertical-pointing beam data separated by 30 km altitude.

In the Nicolls and Heinselman [2007] study, they compared the measured vertical wavelengths with a (then) recently developed dispersion relation that included the effects of kinematic viscosity and thermal diffusion [e.g., Vadas and Fritts, 2005] (Equation 2.35). The measured vertical wavelengths as a function of altitude for a single MSTID event (average $\lambda_z = 230$ km) were found to be larger than those predicted by the dispersion relation, their Figure 4. For our study, we have continued to use their method to enable a statistical comparison of the vertical wavelengths with those obtained from the dispersion relation using our large data set. We have improved on the Nicolls and Heinselman [2007] method by incorporating daily averaged HWM14 horizontal winds [e.g., Drob et al., 2015] in the dispersion relation. We have also included daily averaged neutral temperatures and densities obtained from the NRLMSISE-00 [e.g., Picone et al., 2002] model, which included F10.7 data from NASA’s OMNIWeb [e.g., Mathews and Towheed, 1995].

We now describe the process to determine the vertical wavelengths using a dispersion
relation appropriate for the thermosphere. We begin by rewriting the dispersion relation for the thermosphere, Equation 2.35, in terms of the vertical wave number squared:

\[ m^2 = \frac{N^2}{\omega_{fr}^2 (1 + \delta_+ + \frac{k^2}{H})} \left[ 1 + \frac{\nu^2}{4\omega_{fr}^2} \left( K^2 - \frac{1}{4H_s^2} \right)^2 \left( 1 - Pr^{-1} \right)^2 \right]^{-1} - \frac{k^2}{H} - \frac{1}{4H_s^2}, \quad (4.9) \]

where the variables are the same as those defined in Chapter 2 Section 2.4. To calculate the dispersion vertical wavelengths for the example event on 18 May 2011, HWM14 neutral winds were averaged between the starting and ending times of the MSTID event. Figure 4.9a illustrates the HWM14 zonal (solid) and meridional winds (dashed) used in the dispersion relation. Neutral temperatures and densities were obtained from the NRLMSISE-00 model, where the values were averaged from the starting to ending times of the MSTID event. Figure 4.9b and c shows the temperatures and densities, respectively, used in the dispersion relation. Using the temperature profile in Figure 4.9b, the scale height \( H_s \) was derived using Equation 2.16, and is shown in Figure 4.9d. The Brunt-Väisälä period was derived using Equation 2.8 and setting \( \Gamma_p \approx -10 \text{ K}/\text{km} \), the adiabatic lapse rate for dry air, as shown in Figure 4.9e for the Brunt-Väisälä period. The kinematic viscosity, \( \nu = \frac{\mu}{\rho} \) (Figure 4.9f), was derived using the temperature profile (Figure 4.9b) and density profile (Figure 4.9c), and by using the Vadas and Nicolls [2009] definition of the coefficient of molecular viscosity \( \mu = (0.000334)T^{0.71} \) (Figure 4.9g) and setting \( Pr=0.7 \).

The vertical wavenumber squared \( (m^2) \) in Equation 4.9 cannot be isolated. In order to calculate the vertical wavenumber, a series of equally spaced (0.01 km) vertical wavelength values ranging from \(-1000\) to \(1000\) km were input into the dispersion relation along with the measured horizontal wave parameters individually at each altitude. If

\[ \frac{|LHS - RHS|}{RHS} < 0.01, \quad (4.10) \]

where \( LHS \) and \( RHS \) refer to the left hand side and right hand side of Equation 4.9, respectively, then the \( m \) wavenumber that produced this is returned and the process repeated.
Figure 4.9: Example of the neutral atmospheric parameters utilizing data from HWM14 and NRLMSISE-00 models that are input into the thermospheric dispersion relation. (a) HWM14 zonal (solid) and meridional (dashed) winds, (b) NRLMSISE-00 temperatures, (c) NRLMSISE-00 total mass densities averaged over the time of the MSTID event, (d) the derived scale height, (e) the derived Brunt-Väisälä period, (f) the derived kinematic viscosity, and (g) the derived coefficient of molecular viscosity.

until the end altitude is reached.

The measured vertical wavelengths for the MSTID event observed on 18 May 2011 are shown in Figure 4.10, illustrated as black dots. The derived vertical wavelengths are illustrated as the solid gray line. They have been smoothed using a 3 km running average. For this particular event, the altitude-averaged measured vertical wavelengths were found to be $|\lambda_z| \approx 253$ km and the dispersion vertical wavelengths $\sim 200$ km, illustrating that the dispersion vertical wavelengths are much smaller than the measured. However, they do exhibit the same features, such as increasing in magnitude with altitude.
Figure 4.10: Vertical wavelengths determined using correlation analysis of the PFISR vertically-pointing beam data (black dots) separated by 30 km and the calculated dispersion relation vertical wavelengths (gray line) using the thermospheric dispersion relation for the example MSTID event observed on 18 May 2011.
CHAPTER 5
INVESTIGATING THE CHARACTERISTICS AND PROPAGATION OF MLT SHORT-PERIOD GRAVITY WAVES OVER CENTRAL ALASKA (65° N)

5.1 Introduction

This chapter focuses on investigating the properties of short-period GWs (<1 hr) at high latitudes. As discussed in Chapter 1, GWs are known to play important roles in the dynamics and thermal structure of the neutral upper atmosphere on a local, regional, and global scale [e.g., Fritts and Alexander, 2003, and references therein]. In particular, GW studies in the mesosphere and lower thermosphere (MLT) region (∼80–110 km) have shown they generate strong responses in the atmosphere as they dissipate and deposit significant amounts of energy and momentum [e.g., Bacmeister and Schoeberl, 1989; Fritts, 1984; Schoeberl et al., 1983].

Knowledge of these short-period GW spatial and temporal characteristics, propagation directions, and global distributions are important in order to better understand their fluxes into the MLT region, their season variability, and their dominant sources. While various methods have been used to investigate GW properties, all-sky imaging techniques have been the most successful in investigating the climatology of short-period GWs at mid and low latitudes [e.g., Ejiri et al., 2003; Hecht et al., 2001, 2004; Medeiros et al., 2003; Nakamura et al., 2003; Pautet et al., 2005; Stockwell and Lowe, 2001a; Suzuki et al., 2004, 2011; Taylor et al., 1993, 1997; Walterscheid et al., 1999; Wu and Killeen, 1996]. A now established result is that the waves exhibit similar ranges for their horizontal wavelengths, phase speeds, and periods (despite their potential sources, for example, convection, wind shear, and jet streams) suggesting similar global properties. However, the reported wave propagation headings show considerable site-to-site, latitudinal, and seasonal variability and constitute a major unknown property at this time. Importantly, reports of short-period GW activity at polar latitudes have been rare due to the difficulties of obtaining data under adverse weather conditions and the frequent presence of auroral emissions at high latitudes, which
frequently overpower the faint airglow GW signatures.

The Mesospheric Airglow Imaging and Dynamics (MAID) project was initiated in January 2011 to help quantify to the GW occurrence and properties at high northern latitudes. MAID was a collaborative research program funded by the National Science Foundation led by Dr. K. Nielsen of Utah Valley University (UVU), and involving Utah State University (USU) and the University of Alaska, Fairbanks (UAF). In this chapter, we present novel results of short-period GW measurements from central Alaska obtained over three consecutive winter seasons (2010–2013). Observed yearly and monthly variability in the wave propagations are investigated using MERRA model winds and critical level wind filtering effects. The properties of these waves will be compared with limited published results on GWs at high northern latitudes in Chapter 8.

5.2 Observations

The main goal of the MAID project was to establish a long-term climatology of high latitude, short-period GWs. This was accomplished using an all-sky airglow imager (ASI) located at the Davis Optical Observatory at Poker Flat Research Range (PFRR) (65° N, 147° W), Alaska, see Figure 3.6. The ASI operated in an automatic mode from January 2011 to April 2013 (19 months) and successfully obtained high-quality observations of the four mesospheric airglow emissions yielding a fast \( \sim 2 \) min cadence for the OH data (and \( \sim 6 \) min cadence for the remaining three emissions) as described in Chapter 3. The ASI ran continuously during the long winter nights from September through April when the solar depression angle was \( >12^\circ \) yielding peak nightly observation time of 16 hours duration around winter solstice. During the first two years for this study (January 2011 to April 2012), a total of \( \sim 2100 \) hrs of observations with \( \sim 810 \) hrs of clear sky data were collected, which contain \( \sim 350 \) hrs of discrete GW events. The GW analysis for the third winter season (November 2012 to April 2013) was performed by K. Johnson at UVU. Figure 5.1 shows a summary of the total number of hours of observations for each month, excluding data from November 2012 to April 2013.

Individual GW events were easily identified by their characteristic coherent wave fronts
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Figure 5.1: Summary of the all-sky measurements during the first two years of observations (January 2011 to April 2012) recorded by the PFRR ASI. A total of $\sim2100$ hrs of observations (black) with $\sim810$ hrs of clear sky (gray) and $\sim350$ hrs of discrete GW events (white).

in the OH airglow emission as reported by many prior observers [e.g., Medeiros et al., 2004; Nielsen et al., 2006; Suzuki et al., 2009; Wrasse et al., 2006]. A wave event begins when a clear wave front is observed to enter in the field of view (FOV) or as a wave growing in time within the FOV of the ASI, and ends when the wave pattern is no longer visible. At high latitudes, tropospheric clouds and frequent aurora often limited the duration of the observed wave events. Figure 5.2a shows a distribution of the average time of each event illustrating that a majority of the wave events occurred between 2–16 UT (LT = UT – 9hr). Figure 5.2b shows a distribution of the duration of each wave event, illustrating that a majority of wave events occurred with a duration <3 hours.

5.3 Results

Following the analysis procedure described in Chapter 3, the average observed periods ($\tau$), horizontal wavelengths ($\lambda_H$), phase speeds ($C_H$), and wave azimuths ($\phi$) were obtained from 289 short-period GW events. Figure 5.3a–d shows the individual wave events as gray dots with monthly averages as a solid black line (except for wave azimuths, which are
investigated in a later section). Large gaps in the data set correspond to summer times when the ASI was not in operation. Within each winter season, clusters of wave events can be seen, and the gaps correspond to full-moon periods when no image measurements were made. While there is significant variability in the individual wave periods, horizontal wavelengths, and horizontal phase speeds, they all lie within the same ranges and are consistent throughout the three-year observation period.

Figure 5.4 plots distributions of the observed wave parameters for the combined 2011–2013 winter seasons. The horizontal wavelengths (plot b) exhibit a peak between 20–30 km with an average wavelength of $\sim 29 \pm 1$ km, where the error corresponds to the standard deviation of the mean. The majority of events exhibited wavelengths between 10–50 km. The observed phase speeds (plot c) exhibited a broader peak and a mean value of $44 \pm 2$ m/s with a majority of phase speeds between 30–60 m/s. The derived observed periods (plot a) exhibit typical values between 5–20 minutes with an average period of $13 \pm 1$ min. These distributions and representative values are in good accord to those reported by similar studies at low, mid, and high latitudes, discussed in Chapter 8. Table 5.1 summarizes the horizontal wave parameter averages, medians, and interquartile ranges ($25^{th}$ to $75^{th}$

**Figure 5.2**: Histograms showing the number of GW events observed by the PFRR ASI versus (a) the UT time when the GW event occurred, and (b) the apparent duration of the events, often limited by observing conditions.
Figure 5.3: Horizontal GW parameters as a function of time obtained from PFRR ASI during the three winter seasons. Gray dots represent individual GW events and the solid lines represent monthly averaged wave parameters.

percentile) for the combined (three year) data set.

Figures 5.4d, e, and f all show different plots of the observed wave azimuth distributions. Each of these formats have been used by previous researchers to study anisotropy in wave propagation [e.g., Ejiri et al., 2003; Hecht et al., 2001; Nakamura et al., 2003; Taylor et al., 1997]. They show different important aspects of the wave propagation data. For

Table 5.1: Summary of the MLT GW Parameters for Three Years of Observations at PFRR, a Total of 289 Events

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Mean</th>
<th>Median</th>
<th>Range</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\tau$ [min]</td>
<td>$12 \pm 1$</td>
<td>12</td>
<td>9–15</td>
</tr>
<tr>
<td>$\lambda_H$ [km]</td>
<td>$29 \pm 1$</td>
<td>27</td>
<td>22–34</td>
</tr>
<tr>
<td>$c_H$ [m/s]</td>
<td>$44 \pm 1$</td>
<td>41</td>
<td>30–54</td>
</tr>
</tbody>
</table>
Figure 5.4: Distributions of the observed horizontal GW parameters by the PFRR ASI for the combined 2011–2013 data set. (a) Horizontal wavelengths, (b) phase speeds, (c) periods, (d) wave azimuths, and (e) and (f), which show the same phase velocities (in m/s) for individual GW events represented as lines (e) and dots (f).

example, the pie chart (plot d) identifies three regions of enhanced wave activity with a dominant direction for waves propagating towards the northwest, and two additional preferred directions towards the northeast and southeast are also identified. Plot e shows the wave vectors for the individual GW events. Examination shows that the main peak towards the northwest is comprised mainly of low phase velocity waves (<50 m/s). In contrast, the secondary, northeast peak, comprises much higher velocity waves (>50 and up to 120 m/s). The southeast peak also shows enhanced wave velocities up to 80 m/s. Plot f clearly shows that most of the low-velocity waves have a westward component of propagation, while the higher velocity waves exhibit strong eastward motion. To summarize, the short-period GWs observed during this three-year period indicate clear preferences for both the magnitude and direction of the waves with most events exhibiting westward motion with phase speeds <50 m/s, while the significantly higher phase speed waves exhibited eastward propagation.
5.4 Intra-Seasonal and Inter-Annual Variation

Recognizing these conclusions were drawn from data obtained over three winter seasons, we now investigate their individual winter season distributions. During the first year, data were obtained over four months (January to April). The second year contained data over the full nine months (August to April), while the third winter comprised data over six months (November to April). Table 5.2 summarizes the averaged wave parameters for these three consecutive winter seasons, again illustrating the wave parameters remaining consistent.

Variability in the preferred propagation directions between years is explored in Figure 5.5. Of the total number of events (289) 67 were obtained during the first year (2010/2011), 149 events were obtained in the extended observations in year 2 (2011/2012), and 78 during the final year (2012/2013) of the measurements. While there are significant differences between the three individual years, each of the three winter seasons clearly identifies the three preferred propagation azimuths to the northwest, northeast, and southeast. The wave vector plots (d, e, and f) for the separate winter seasons illustrate the same general behavior with lowest phase speeds to the northwest, highest phase speeds to the northeast, and intermediate phase speeds to the southwest, consistent with our conclusion drawn from the combined data set. This year-to-year repeatability in the phase velocity distribution establishes their climatology at this high latitude location.

Taking this a step further, we now examine the intra-seasonal variability (monthly variations) of the horizontal azimuthal propagation directions by combining data for each month into a single monthly plot as shown in Figure 5.6. For example, data from January 2011, 2012, and 2013 are combined into the January plot in Figure 5.6. Data from the months of August and September and the months of March and April are combined.

Table 5.2: Yearly Averages of the Horizontal Wave Parameters Obtained Using the PFRR ASI

<table>
<thead>
<tr>
<th>Year</th>
<th>N</th>
<th>τ [min]</th>
<th>λ_H [km]</th>
<th>c_H [m/s]</th>
</tr>
</thead>
<tbody>
<tr>
<td>2010/2011</td>
<td>67</td>
<td>11 ± 1</td>
<td>28 ± 1</td>
<td>48 ± 2</td>
</tr>
<tr>
<td>2011/2012</td>
<td>149</td>
<td>13 ± 1</td>
<td>29 ± 1</td>
<td>42 ± 2</td>
</tr>
<tr>
<td>2012/2013</td>
<td>73</td>
<td>13 ± 1</td>
<td>28 ± 2</td>
<td>44 ± 3</td>
</tr>
</tbody>
</table>
Figure 5.5: Wave azimuth distributions obtained by the PFRR ASI for the three consecutive winter seasons (a) 2010/2011, (b) 2011/2012, and (c) 2012/2013. Note, distributions have been normalized by the number of events within each year to facilitate visual comparison. Plots (d)–(f) show the wave phase velocities for each winter season.

because of the low number of events within each month. The resultant seven plots depict changes in the preferential wave directions during the three year average winter season. The top row shows the number of events versus direction, while the bottom row identifies their phase velocities. Most of the data were obtained during the four main winter months (November to February). During November and December dominant wave motions are westward with a preference for northwest propagation. These waves exhibited phase speeds $<50$ m/s. In contrast, January and February show a marked changeover to primarily eastward motion in the preferred northeast and southeast direction and generally higher wave phase speeds $>50$ m/s. During the spring, the wave motions remained strongly eastward and were dominated by $>50$ m/s events to the southeast. In summary, in the fall period (August to October), the main winter period (November to February), and spring (March/April) all show signatures of the three dominant directions of motion (northwest, northeast, and southeast) that we have previously identified. Furthermore, during the early
winter season, the waves exhibited a strong westward component of motion, while eastward motion dominated only during the rest of the winter season. This result is further illustrated in Figure 5.7, which plots the ratio of the number of waves having an eastward to westward component of propagation. The upper plot indicates a systematic change from westward during the early winter to eastward in late winter and early fall months. Also note the consistent and limited preference for northward wave propagation during the winter in the lower plot.

Table 5.3 shows the monthly averages of the horizontal wave parameters for the three years combined. Note the general consistency of the wave parameters from month to month.

5.5 Effects of Critical Level Wind Filtering

To investigate the observed characteristics of the wave propagation directions over central Alaska, we have examined the effects of critical level wind filtering, which was introduced in Chapter 2. Briefly, critical level wind filtering occurs when the observed phase speed of the wave equals the background winds in the direction of wave propagation [e.g., Hines and Reddy, 1967; Taylor et al., 1993]. Under this condition, the wave will be absorbed into the background mean flow.

Figure 5.8a shows the location of PFRR along with a $10^\circ \times 10^\circ$ latitude by longitude shaded region centered on PFRR illustrating the latitude and longitude boundaries used for

<table>
<thead>
<tr>
<th>Month</th>
<th>N</th>
<th>$\tau$ [min]</th>
<th>$\lambda_H$ [km]</th>
<th>$c_H$ [m/s]</th>
</tr>
</thead>
<tbody>
<tr>
<td>August/September</td>
<td>7</td>
<td>13 $\pm$ 2</td>
<td>25 $\pm$ 2</td>
<td>35 $\pm$ 3</td>
</tr>
<tr>
<td>October</td>
<td>26</td>
<td>13 $\pm$ 1</td>
<td>35 $\pm$ 2</td>
<td>56 $\pm$ 6</td>
</tr>
<tr>
<td>November</td>
<td>35</td>
<td>14 $\pm$ 1</td>
<td>29 $\pm$ 1</td>
<td>43 $\pm$ 3</td>
</tr>
<tr>
<td>December</td>
<td>61</td>
<td>12 $\pm$ 1</td>
<td>27 $\pm$ 2</td>
<td>44 $\pm$ 3</td>
</tr>
<tr>
<td>January</td>
<td>56</td>
<td>13 $\pm$ 1</td>
<td>29 $\pm$ 1</td>
<td>43 $\pm$ 3</td>
</tr>
<tr>
<td>February</td>
<td>39</td>
<td>13 $\pm$ 1</td>
<td>30 $\pm$ 2</td>
<td>42 $\pm$ 3</td>
</tr>
<tr>
<td>March/April</td>
<td>65</td>
<td>11 $\pm$ 2</td>
<td>28 $\pm$ 1</td>
<td>45 $\pm$ 2</td>
</tr>
</tbody>
</table>
Figure 5.6: Monthly wave azimuth distributions obtained by the PFRR ASI for the three combined winter seasons.
the MERRA wind data (discussed in Chapter 2). Figure 5.8b, c, and d plot the MERRA vector winds over Alaska at the surface, near the peak of the stratospheric winds ~40 km, and an altitude of ~77 km, respectively. A mean wind profile was created by averaging the six-hour MERRA vector winds within the sample volume as a function of altitude. These were then averaged to create a daily (24 hr) profile that was used to investigate effects of critical layer filtering on individual wave events, while the monthly averaged winds derived from the MERRA data were used to generate blocking diagrams to investigate the characteristics of the seasonal wave filtering.

Figure 5.9 shows an example of a daily averaged wind profile and the derived blocking diagram for this day using the MERRA winds from 17 March 2011. On this occasion, a wave event was observed ~11–14 UT exhibiting an average horizontal wavelength of ~30 km, an observed phase speed of ~55 m/s, and a direction of propagation of ~59°. Plotted in Figure 5.9a, are the zonal winds (gray), meridional winds (black), derived wind in the direction of the wave propagation (blue), and the derived intrinsic phase speed of the wave (red) as a function of altitude. A blocking diagram (see Chapter 2) was constructed using the daily average MERRA winds up to an altitude of 77 km. This is plotted in Figure 5.9b
Figure 5.8: Illustrations of the MERRA winds encompassing PFRR used for the blocking analysis. (a) Map showing location of PFRR (black dot). The red shaded box encompassing PFRR denotes the spatial limits used for MERRA horizontal wind data. (b) Arrows represent the direction of the winds at the surface (0 km) and are normalized so that all arrows are the same length. The underlying contour shows the wind speed. (c) The MERRA winds at an altitude of $\sim 40$ km where the peak stratospheric winds occur, showing the winds are towards the east. (d) MERRA horizontal winds at an altitude of $\sim 77$ km, showing the winds are southeast.

As the shaded area along with the observed wave phase velocity. On this occasion, the wave phase velocity is greater than the 24-hr averaged blocking diagram in both magnitude and direction, suggesting that a wave of tropospheric origin could have propagated into the MLT region. Repeating this analysis for each of the 289 observed wave events has yielded a large fraction ($\sim 72\%$) of the waves exceeded their blocking region and were able to attain MLT heights. To test the sensitivity of this result, a change in the background winds by up to $\pm 10\%$ resulted in an increase/decrease in the waves able to penetrate into the MLT by $\sim 3$-$5\%$. 
Figure 5.9: Example of a single GW event observed on 17 March 2011 by the PFRR ASI exceeding the blocking MERRA winds. (a) Daily-averaged MERRA zonal (gray) and meridional (black), wind in the direction of the wave (blue) and intrinsic phase speed (red). (b) Blocking diagram with wave phase vector. Data from the GW event with a horizontal wavelength $\sim 30.9$ km, phase speed $\sim 55.89$ m/s, direction of propagation $\sim 58.89^\circ$, and period of $\sim 9$ min were utilized.

Using the daily average MERRA winds, blocking diagrams for each month were constructed, as shown in Figure 5.10. Due to the predominantly eastward winds, the largest blocking regions occurred to the east at MLT heights and were dominant during the winter months. Under these conditions low phase speed westward propagating waves can readily achieve MLT heights, while eastward propagating waves must have significantly higher phase speeds to overcome the critical level blocking region. From Figure 5.10, it is evident that the magnitude of the blocking regions varied significantly from month to month. For each month, the majority of the events (red dots) occurred well outside these blocking regions consistent with critical level filtering theory. (Note, the blocking diagrams are monthly climatologies and, as such, not all the waves occurred outside the blocking region.)
Figure 5.10: Monthly blocking diagrams (solid gray) using MERRA winds along with the PFRR ASI observed phase velocities for individual GW events (red dots).
Furthermore, the individual plot in Figure 5.10 also explains our results showing the wave propagation changing over from mainly westward in the early winter months, where the eastward wind blocking was largest, to both eastward and westward in the later winter months where the eastward blocking was significantly reduced. In summary, westward propagating waves were minimally inhibited, but the strong eastward winds effectively blocked most of the low phase speed eastward propagating waves during the main winter season. These results provide new convincing evidence that wind filtering was most probably responsible for both the observed range of phase speeds for the waves and their apparent switch over from eastward to westward propagation for waves reaching MLT heights. The critical level filtering does not explain the observed three preferential directions during the winter season over Alaska, which may be associated with differing wave sources.

5.6 Summary

In this chapter the properties of high northern latitude (65° N) short-period GWs observed in all-sky imagery from Poker Flat Research Range, Alaska, have been investigated. Observations over three consecutive (2011–2013) winter seasons provide a detailed climatology for the observed wave parameters (horizontal wavelength, phase speeds, and observed periods) for GWs in the MLT region. The basic properties of the waves were found to be consistent from year to year and will be compared and contrasted with other more limited reported studies in Chapter 8.

The wave azimuth distribution for the combined three winter seasons identified three directions of enhanced wave activity with apparent dominant direction for waves propagating toward the northwest and two additional preferred directions toward the northeast and southeast. Clear preferences in both the magnitude and direction of the wave phase velocities were evident, characterized by westward progressing waves with phase speeds <50 m/s, while significantly higher phase speed waves exhibited eastward propagation. The separate winter seasons each illustrated the same general behavior with lowest phase speeds to the northwest, highest phase speeds to the northeast, and intermediate phase speeds to the southwest. This year to year repeatability in the phase velocity distribution further es-
ablishes their climatology at this high northern latitude location. A consistent result was obtained when investigating the individual monthly variability in the distributions of the wave azimuths during the fall, winter, and spring.

Using MERRA winds, it was shown that while the westward propagating waves were minimally inhibited, the strong eastward winds effectively blocked most of the low-phase-speed eastward propagating waves during the main winter season. Together, these results provide new observational evidence that wind filtering was most probably the cause for both the observed range of phase speeds for the waves and their apparent switch-over during midwinter from eastward to strongly westward propagation for waves reaching MLT heights.
CHAPTER 6
HIGH-LATITUDE ARCTIC INVESTIGATION OF MLT SHORT-PERIOD GRAVITY WAVES OVER NORTHERN NORWAY (69° N)

6.1 Introduction
In this chapter, results obtained from the Advanced Mesospheric Temperature Mapper (AMTM), introduced in Chapter 3, are presented. Data for this study were obtained over one winter season from October 2011 to March 2012. A total of 310 short-period gravity wave (GW) events were identified from observations of the OH airglow emissions. Following the analysis method described in Chapter 3, the ground-based wave parameters were derived and the monthly variations in the wave azimuths were investigated.

During this winter, radar wind data were available for almost every night of observation and enabled an investigation of the intrinsic wave parameters from which the vertical wavelength spectrum was also determined using the dispersion relation appropriate for the mesosphere lower thermosphere (MLT) region (∼80–110 km). These results complement and expand significantly on the measurements obtained at the Poker Flat Research Range (PFRR), discussed in Chapter 5.

6.2 Observations
For this study, we used data obtained by the AMTM to quantify short-period GWs in the MLT region above Northern Norway from the Arctic Lidar Observatory for Middle Atmospheric Research (ALOMAR) (69° N, 16° E). The AMTM results presented here were gained during a single winter season from October 2011 to March 2012 with ∼2080 hrs of total observation time (Figure 6.1 black bars). Of the total observation time, ∼660 hrs were clear sky (Figure 6.1 gray bars) with ∼470 hrs of wave events (Figure 6.1 white bars). In contrast to the all-sky imager (ASI) (Chapter 5), the AMTM is capable of making observations during aurora and full moon periods providing significantly more data. PFRR and ALOMAR are at similar latitudes, but separated by over 160° longitude. We will
Figure 6.1: Summary of the ALOMAR AMTM observing conditions. Black bars correspond to the total number of observation hours, gray bars to the total clear sky hours, and white bars to the total hours of GW observations.

follow a similar analysis procedure to that used in Chapter 5 to investigate these additional high-latitude data.

6.3 Results

Chapter 3 described how individual GW events were identified and analyzed using the ASI. In this chapter, we use the same method to investigate GW events within the AMTM image data. As noted above, a total of 310 short-period GW events were detected. Using the same general format as Figure 5.2, the average wave parameters for each individual wave event (gray dots) along with their monthly averages (solid black line) are shown in Figure 6.2. The monthly averaged wave periods (Figure 6.2a) varied little with a slight decrease from ~15 min at the beginning of the observation period (October) to ~10 min at the end of the season (March) with some additional higher wave periods (>40 min) from ~January to March. The mean horizontal wavelengths (Figure 6.2b) remained essentially the same throughout the six month observation period at ~20 km. The phase speeds (Figure 6.2c) are also consistent throughout the winter and show a significant increase from
Figure 6.2: Horizontal GW parameters as a function of time obtained from ALOMAR during the winter season 2011/2012. Gray dots represent individual GW events and the solid lines represent monthly averaged wave parameters.

∼27 m/s to ∼40 m/s during the winter period. The wave azimuths (Figure 6.2d) were well scattered throughout the observation period; however, they also reveal preferred azimuths between 200–360° (northwest) and 0°–100° (northeast).

Figure 6.3 is a histogram summary of the observed horizontal wave parameters for the six months of observations at ALOMAR. The distribution of the observed wave periods exhibit a mean period of 13 min with an interquartile range from ∼8–15 min. Observed horizontal wavelengths exhibits a mean of ∼21 km ranging from 17–25 km and horizontal phase speeds with a mean of 35 m/s ranging from 23–46 m/s. Table 6.1 summarizes the
Table 6.1: Mean, Median, and Range Values for all GW Horizontal Wave Parameters Obtained Using the ALOMAR AMTM.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Mean</th>
<th>Median</th>
<th>Range</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\tau$ [min]</td>
<td>$13 \pm 1$</td>
<td>10</td>
<td>8–25</td>
</tr>
<tr>
<td>$\lambda_H$ [km]</td>
<td>$21 \pm 1$</td>
<td>21</td>
<td>17–25</td>
</tr>
<tr>
<td>$c_H$ [m/s]</td>
<td>$35 \pm 1$</td>
<td>32</td>
<td>23–46</td>
</tr>
</tbody>
</table>

average, median, and interquartile ranges of these data. Figure 6.3d shows the distribution of wave azimuths summed in $30^\circ$ bins. Two peaks in the distribution are evident; a primary peak toward the northwest and a secondary peak toward the northeast. (Note, these two directions were also prominent in the PFRR data of Figure 5.4e, and will be discussed further in Chapter 8).

Figure 6.3e shows the wave phase velocities for each GW event. The strong activity toward the northwest mainly comprised waves with phase speeds ranging from 30–50 m/s. In comparison, the northeastward waves frequently exhibited phase speeds $>50$ m/s. Although not evident in the pie chart (plot d), wave activity toward the southeast in plot e comprised a range of waves between 30–50 m/s, similar to the intermediate phase speed peak in the PFRR ASI data. In comparison, the southwest quadrant was fairly uniform and exhibited generally lower phase speeds $<40$ m/s. Lastly, an examination of Figure 6.3f, reveals that waves with westward propagation components were strongly clustered and comprised comparatively low phase speed waves ($<40$ m/s), while the waves with an eastward propagation component were equally distributed, but comprised higher phase speed events.

6.4 Monthly Variation

Monthly averaged wave parameters, summarized in Table 6.2, show little variation throughout the observation period. Examination of the main winter period (December to February), Figure 6.4, shows substantial northwest propagation in each month that comprised waves with phase speeds of typically 40 m/s. In comparison, the northeast peak is generally not as prominent, but is also evident throughout the winter and was characterized by waves with phase speeds $>40$ m/s. Wave propagation toward the southeast was more
Figure 6.3: Distributions of the observed horizontal GW parameters observed with the ALOMAR AMTM. (a) Horizontal wavelengths, (b) phase speeds, (c) periods, and (d) wave azimuths for the 2011–2012 data set. Phase velocities (in m/s) for individual GW events represented as lines (e) and dots (f).

Table 6.2: Monthly Averaged Horizontal Wave Parameters Obtained Using the ALOMAR AMTM

<table>
<thead>
<tr>
<th>Month</th>
<th>N</th>
<th>$\lambda_H$ [km]</th>
<th>$c_H$ [m/s]</th>
<th>$\tau$ [min]</th>
</tr>
</thead>
<tbody>
<tr>
<td>October 2011</td>
<td>23</td>
<td>19 ± 1</td>
<td>27 ± 2</td>
<td>15 ± 2</td>
</tr>
<tr>
<td>November 2011</td>
<td>36</td>
<td>19 ± 1</td>
<td>27 ± 2</td>
<td>13 ± 1</td>
</tr>
<tr>
<td>December 2011</td>
<td>63</td>
<td>22 ± 1</td>
<td>39 ± 2</td>
<td>11 ± 1</td>
</tr>
<tr>
<td>January 2012</td>
<td>86</td>
<td>22 ± 1</td>
<td>32 ± 2</td>
<td>15 ± 1</td>
</tr>
<tr>
<td>February 2012</td>
<td>56</td>
<td>20 ± 1</td>
<td>39 ± 3</td>
<td>11 ± 1</td>
</tr>
<tr>
<td>March 2012</td>
<td>46</td>
<td>22 ± 1</td>
<td>40 ± 2</td>
<td>10 ± 1</td>
</tr>
</tbody>
</table>

evident in the early winter months and almost absent later in the winter. Overall, the preferred directions of motion, as illustrated in Figure 6.5, identifies a dominant westward motion during the fall and winter switching to eastward motion in the spring. During the winter, the majority of the waves also exhibited a northward component of motion; however, during the late fall/early winter the waves exhibited nearly equal numbers of southward and northward propagation with only a slight preference for southward propagation.
Figure 6.4: Distributions of the monthly propagation directions in the form of histograms (top row) and individual phase velocities (m/s) (bottom row) obtained from the ALOMAR AMTM.
Figure 6.5: Ratio of the (a) number of eastward/westward propagating waves, and (b) the number of northward/southward propagating waves obtained from the ALOMAR AMTM, identifying systematic changes in direction during the winter season.

6.5 Intrinsic Wave Characteristics

Using the analysis method described in Chapter 3, the intrinsic periods for 224 wave events have been calculated from the initial 310 wave events using available background wind data. Figure 6.6a plots the intrinsic period distribution, which is characterized by a strong narrow peak between 5–15 min. For direct comparison, plot b presents the observed period distribution re-calculated from Figure 6.3 using the same 224 events. Note, the scales of both figures are the same to facilitate visual comparison. While the observed distribution is similar in shape, it is not as prominent as the intrinsic period distribution and comprises more longer-period waves up to ~50 min. This narrow width of the intrinsic period distribution is supported by examination of Table 6.3, which shows the median, interquartile range limits, and interquartile range.

Figure 6.7 summarizes the derived intrinsic phase speeds together with the observed phase speed distribution for the same 224 events. While both distributions peak between 20–40 m/s, there are more events in the observed phase speed distribution within this
Table 6.3: Mean, Median, Interquartile Range Limits, and IQR’s for the Observed and Intrinsic Wave Parameters Obtained Using the ALOMAR AMTM

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Mean</th>
<th>Median</th>
<th>Range</th>
<th>IQR</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\tau_{\text{obs}}$ [min]</td>
<td>$13 \pm 1$</td>
<td>11</td>
<td>8–15</td>
<td>7</td>
</tr>
<tr>
<td>$\tau_{\text{int}}$ [min]</td>
<td>$12 \pm 1$</td>
<td>9</td>
<td>7–13</td>
<td>6</td>
</tr>
<tr>
<td>$c_{\text{obs}}$ [m/s]</td>
<td>$35 \pm 1$</td>
<td>32</td>
<td>23–46</td>
<td>23</td>
</tr>
<tr>
<td>$c_{\text{int}}$ [m/s]</td>
<td>$42 \pm 1$</td>
<td>39</td>
<td>28–55</td>
<td>27</td>
</tr>
<tr>
<td>$\lambda_H$ [km]</td>
<td>$22 \pm 1$</td>
<td>21</td>
<td>18–25</td>
<td>7</td>
</tr>
<tr>
<td>$\lambda_z$ [km]</td>
<td>$20 \pm 1$</td>
<td>16</td>
<td>10–26</td>
<td>16</td>
</tr>
</tbody>
</table>

Figure 6.6: Distributions of the (a) intrinsic wave periods (obtained using the ALOMAR Saura radar winds from 85–95 km), and (b) observed periods (replotted from Figure 6.3c for comparison). Note the scales of both figures are the same to facilitate visual comparison.

range and the intrinsic phase speed distribution is significantly broader (Table 6.3) with more waves exhibiting higher intrinsic phase speeds.

To investigate this broadening, Figure 6.8a-b plots the observed phase velocity distribution (as introduced in Section 6.3) and (c-d) the intrinsic phase velocities for the individual 224 wave events (Table 6.3). As noted, the observed westward wave distribution was strongly clustered and comprised comparatively low phase speed waves, while the eastward
waves were significantly distributed and comprised higher velocity events. Examination of the Saura wind data shows that the westward waves were propagating against the background wind field at mesospheric altitudes causing their intrinsic phase speeds to increase (Chapter 2 Equation 2.37). The net effect of the background winds on the observed wave events was, therefore, to create a more uniform and broader distribution of the intrinsic phase velocities, as clearly illustrated in Figure 6.8.

6.6 Vertical Wavelength Distribution

This section presents the results of the derived vertical wavelengths for 224 wave events. Following the analysis method described in Chapter 3, the vertical wavelength distribution was derived for the 224 events. Figure 6.9a compares the distributions of the derived vertical wavelength, plot a, with the measured horizontal wavelength distribution, plot b. The distribution of the horizontal wavelengths is relatively narrow peaking around 20 km, while the derived distribution of the vertical wavelengths was significantly broader peaking around
Figure 6.8: Comparison of the observed phase velocities (in m/s) (top row) and derived intrinsic phase velocities (in m/s) (bottom row) for 224 GW events obtained from the ALOMAR AMTM.

15 km and extended to >60 km (Table 6.3). This broad distribution results from the effects of the prevailing eastward background winds on the westward component of propagating waves, causing their vertical wavelength to increase, while the eastward propagating waves exhibit shorter vertical wavelengths, further broadening the distribution.

6.7 Effects of Critical Level Wind Filtering

In the previous section, we utilized the Saura radar neutral winds to establish their effects on the GW intrinsic period, phase speed and vertical wavelength spectrum. In this section we further investigate the effects of the neutral winds on the GW propagation.

Radar winds were available for most nights of observations, but they only cover the MLT
Figure 6.9: Distributions of the (a) derived vertical wavelengths, and (b) observed horizontal wavelengths obtained from the ALOMAR AMTM. Vertical wavelengths were calculated using a dispersion relation for the MLT region with constant values for the Brunt-Väisälä frequency \( N = 0.02 \text{ s}^{-1} \) and scale height \( H_s = 7 \text{ km} \) along with Saura radar winds from 85–95 km.

To investigate the winds below 80 km, we use MERRA horizontal winds as introduced in Chapter 2. Using a combination of the monthly mean MERRA and Saura data, a similar blocking analysis to that applied to the PFRR ASI data from Chapter 5 has been performed. An example of the combined MERRA and Saura is plotted in Figure 6.10, which shows the daily averaged (red solid) zonal (a) and meridional (b) winds every six hours (dashed black lines) for the MERRA (centered on ALOMAR) and every 30 min for Saura (gray dashed lines) with daily averaged winds (solid blue line) for 4 January 2012.

Blocking diagrams were constructed for six winter months using combined MERRA and Saura monthly mean wind data. Figure 6.11 plots them for each month as the gray shaded regions. Red dots represent the phase velocities of individual GW events. As expected from our PFRR analysis, the blocking for each month was primarily eastward and largest during the winter, except for January where the blocking was almost nonexistent due to the
occurrence of a Sudden Stratospheric Warming (SSW) event. This special situation will be discussed later in Chapter 8.

During the fall/early winter months (October and November), wave events occurred primarily to the west with phase speeds greater than the blocking region. Most events (>140) were observed during midwinter (December to February). These waves were more omnidirectional with phase speeds mostly exceeding their climatological blocking limits. In January and March, the blocking was the least and many events exhibited low phase speeds (<40 m/s). In summary, the observed wave distributions throughout the winter at ALOMAR also appear consistent with critical level filtering.

6.8 Summary

This chapter presents novel results obtained by the Advanced Mesospheric Temperature Mapper (AMTM) on the propagation characteristics of 310 GW events imaged over northern Norway during the winter 2011–2012. The data were analyzed in a similar way to the
Figure 6.11: Blocking diagrams for each month (gray) using combined monthly averaged MERRA and the ALOMAR Saura winds along with individual GW phase velocities (red) obtained from the AMTM.

PFRR data from Alaska (Chapter 5) facilitating a comparison of the wave characteristics, as presented in Chapter 8. Available Saura MLT radar wind data enabled a more in-depth investigation of the waves including their intrinsic parameters (224 events) and their vertical wavelength spectrum using the dispersion relation for the MLT region. Monthly variations in the wave azimuths were investigated and provide additional strong evidence for the effects of critical level wind filtering on the observed phase speed spectrum using a combination of MERRA and Saura radar winds.
CHAPTER 7
SEASONAL PROPAGATION CHARACTERISTICS OF MSTIDS OBSERVED AT HIGH LATITUDES OVER CENTRAL ALASKA USING THE POKER FLAT INCOHERENT SCATTER RADAR

This chapter includes results of a research paper submitted in October 2017 to the Journal of Geophysical Research (JGR) Space Physics, with M.R. Negale as lead author, and is currently under review.

7.1 Introduction

This study focuses on thermospheric gravity waves, termed MSTIDs, as introduced in Chapter 1. The measurements were made using the Poker Flat ISR (PFISR) operated at the Poker Flat Research Range (PFRR) (65.13° N, 147.47° W, MLAT=65.4° N) near Fairbanks, Alaska. PFISR is a part of the Advanced Modular Incoherent Scatter Radar (AMISR) class of ISRs, and has the capability to rapidly observe different regions of the ionosphere using pulse-to-pulse beam steering. This makes PFISR an ideal instrument for investigating the three-dimensional properties of TIDs at high latitudes [e.g., Nicolls and Heinseelman, 2007; Vadas and Nicolls, 2008, 2009; Waldock and Jones, 1986].

While MSTIDs have been studied extensively in the high and mid latitude F region, their full seasonal propagation characteristics are still not well known [e.g., Bristow and Greenwald, 1996; Bristow et al., 1994; Evans et al., 1983; Frissell et al., 2014; Grocott et al., 2013; Hernández-Pajares et al., 2012; Ishida et al., 2008; Kotake et al., 2007; Ogawa et al., 1987; Samson et al., 1989, 1990; Waldock and Jones, 1986]. Table 7.1 lists results from several studies of MSTIDs at high latitudes where a large number of events have been measured, providing statistics on their propagation characteristics. Most of these studies determined the horizontal wavelength ($\lambda_H$), phase speed ($c_H$), observed period ($\tau$), and direction of propagation ($\phi$) of the wave, but often over a restricted altitude range in the thermosphere. To date, most MSTID studies have utilized multi-site radar systems, such
Table 7.1: MSTID Wave Parameter Results From Selected Studies Performed at High Latitudes Illustrating the Broad Range of Wave Measurements

<table>
<thead>
<tr>
<th>Loc.</th>
<th>Meth.</th>
<th>N</th>
<th>Dates</th>
<th>Times</th>
<th>$\lambda_H$ [km]</th>
<th>$\tau$ [min]</th>
<th>$c_H$ [m/s]</th>
<th>$\phi$ [deg]</th>
<th>References</th>
</tr>
</thead>
<tbody>
<tr>
<td>53.32°N, 49.39°N, 52.16°N, 53.98°N</td>
<td>Super-DARN</td>
<td>304</td>
<td>Nov. 2012-Apr. 2015</td>
<td>Days</td>
<td>100-450</td>
<td>30-40</td>
<td>75-325</td>
<td>125-325° Southward</td>
<td>Frissell et al. [2016]</td>
</tr>
<tr>
<td>65.1°N</td>
<td>Airglow</td>
<td>44</td>
<td>Nov. 2001-Apr. 2002</td>
<td>Days &amp; nights</td>
<td>100-400</td>
<td>20-60</td>
<td>80-200</td>
<td>210-270° Southward</td>
<td>Kubota et al. [2011]</td>
</tr>
<tr>
<td>78°N</td>
<td>EISCAT</td>
<td>244</td>
<td>Mar. 2007-Feb. 2008</td>
<td>Nights</td>
<td>30-120</td>
<td>~250</td>
<td>–</td>
<td>–</td>
<td>Vlasov et al. [2011]</td>
</tr>
</tbody>
</table>
as the SuperDARN network and recently, satellite observations of the TEC [e.g., Ding et al., 2011; Frissell et al., 2016; Ishida et al., 2008]. The results illustrate a broad range of wave parameters, although all lie within the MSTID category. The results consistently show that the observed waves almost always propagated southward. However, individual events exhibited propagation directions that varied during the course of the day and from day-to-day.

In this chapter, we use extensive electron density profile data obtained by PFISR. Over 650 MSTID events were detected and measured over an extended period from August 2010 to April 2013, which encompassed the mesospheric optical measurement program. To our knowledge, this study provides the most extensive seasonal height-resolved investigation of MSTIDs to date, as it includes year-round measurements including novel summertime data. MSTID detection and selection was described in Chapter 4 using a single MSTID event observed on 18 May 2011. Using previously developed, well-proven methods of spatially separated ionospheric measurements [e.g., Afraimovich et al., 1999; Bristow and Greenwald, 1996; Nicolls and Heinselman, 2007; Reid, 1986], we determined the period and horizontal wave vector as a function of altitude for this MSTID event, and hence, derived their vertical wavelengths and altitudinal structure. This is followed by a discussion of our results, which includes an investigation into the seasonal characteristics of the propagation directions, comparisons with published SuperDARN measurements, and new observations of the vertical wavelength spectra. This is followed by a discussion on the effects of critical level and dissipative filtering on the MSTIDs.

7.2 Previous PFISR Observations

Nicolls and Heinselman [2007] made observations of a single MSTID event that occurred on 13 December 2006. In their study, they used the (then) recently developed phased array PFISR to make novel multi-beam (10 beams used) measurements of an MSTID. Their observations provided measurements of the MSTID over an extended altitude range \( z \sim 160–220 \) km. The wave event lasted for \( \sim 1.5 \) hrs, and exhibited height integrated horizontal wavelength of \( \sim 187 \) km, horizontal phase speed of \( \sim 140 \) m/s, observed period of \( \sim 22 \) min,
southeastward ($\sim 150^\circ$) propagation direction, and vertical wavelength of $\sim 231$ km. The following results build extensively on this original work.

### 7.3 MSTID Statistical Characteristics

The analysis methods described in Chapter 4 have been used to investigate a total of 652 MSTID events that were identified and measured from August 2010 to April 2013 PFISR data. The observed MSTID events ranged from 100–450 km in altitude. Due to the reduced signal strength at higher altitudes, we have limited the data products presented here to an altitude range of 100–300 km. This also facilitates a better comparison of our statistical results with published measurements. Figure 7.1d shows a summary of the mean altitude where each MSTID event occurred. For example, if an event occurred over the altitude range of 100–300 km, then then mean altitude for that event was $\bar{z} = 200$ km. Figure 7.1a–c

![Figure 7.1](image.png)

**Figure 7.1**: Frequency distributions of the observed horizontal wave parameters obtained from PFISR. (a) Period, (b) horizontal wavelength, (c) horizontal phase speed, (d) mean altitude of each observation ($\sim 68\%$ of events occurred with mean altitude $\bar{z} \leq 250$ km), and (e) individual horizontal phase velocities.
shows the distributions for the altitude averaged horizontal wave parameters. The periods ranged from $\sim 30$–60 min, horizontal wavelengths from $\sim 200$–800 km, and phase speeds from $\sim 100$–350 m/s. Mean values, medians, and interquartile ranges (25–75th percentile) are summarized in Table 7.2. These results compare exceptionally well with other MSTID studies at high, mid, and low latitudes [e.g., Bristow and Greenwald, 1996; Bristow et al., 1994; Evans et al., 1983; Frissell et al., 2014, 2016; Grocott et al., 2013; Hernández-Pajares et al., 2012; Ishida et al., 2008; Kotake et al., 2007; Ogawa et al., 1987; Samson et al., 1989, 1990; Waldock and Jones, 1986]. However, our data also enable a novel investigation into the variability with altitude, and seasonal changes of the wave parameters.

Figure 7.1e shows the height-averaged phase velocity distribution of 651 MSTID events with the maximum phase speed limited to 400 m/s. The predominant propagation direction of MSTIDs was southeast, but extends from north through the southwest directions. Remarkably, very few (<5%) of the waves were detected propagating westward over a broad $120^\circ$ azimuth range extending from $240$–$360^\circ$. Note also that one very-high phase speed event (not plotted here) exhibited a phase speed of $\sim 500$ m/s (azimuth $112^\circ$), which is close to the speed of sound in the thermosphere. In comparison, the smallest observed phase speed was $\sim 86$ m/s. In total, 97% of the observed MSTIDs exhibit phase speeds between 100–300 m/s (mean value 187 m/s, see Table 7.2).

### 7.4 Altitudinal Variability

We now investigate the variability of the wave parameters with altitude. Figure 7.2 shows the averaged wave parameters in 50 km altitude bins from 100–300 km, where the error bars represent the standard deviation of the mean. Of key importance is that the

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Mean</th>
<th>Median</th>
<th>Range</th>
</tr>
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<tbody>
<tr>
<td>$\lambda_H$ [km]</td>
<td>446</td>
<td>431</td>
<td>369–514</td>
</tr>
<tr>
<td>$\tau$ [min]</td>
<td>41</td>
<td>41</td>
<td>37–44</td>
</tr>
<tr>
<td>$C_H$ [m/s]</td>
<td>187</td>
<td>181</td>
<td>155–212</td>
</tr>
<tr>
<td>$\phi$ [deg]</td>
<td>120</td>
<td>120</td>
<td>84–160</td>
</tr>
</tbody>
</table>
wave periods and azimuths are nearly constant with altitude, while both the horizontal wavelengths and the phase speeds increase systematically and significantly with altitude (i.e., \( \sim 20\% \) increase for both the wavelengths and the phase speeds).

To date there have been very few reports showing the variability of MSTID parameters with altitude at high latitudes. These investigations mainly used ISR and ionosonde techniques, but with limited data and only one study at high latitudes using the EISCAT radar [e.g., Djuth et al., 1997; Ma et al., 1998; Nicolls et al., 2014; Oliver et al., 1997; Ratovsky et al., 2008; Tedd et al., 1984; Vadas, 2007]. In particular Nicolls et al. [2014] reported TID height structure from Arecibo Observatory (18° N) obtained on two consecutive days (23–25 July 2009). Their Figure 13, plotted over the same altitude range as our results (\( z=100–300 \) km), shows that both the horizontal wavelength and phase speed increase with altitude, while the wave periods remained essentially constant with altitude. These limited results agree very well with our much larger ensemble of high latitude MSTIDs. Along with our new measurements, these studies confirm that the wave parameters change with altitude in the thermosphere.
7.5 Discussion

In this Chapter, we have used data obtained from PFISR from August 2010 to April 2013 to determine new results on the statistical characteristics of high latitude MSTIDs and their vertical variability. This is followed by an investigation of critical level and dissipative filtering effects on the propagation spectrum and a novel investigation of the vertical wavelength spectrum.

7.5.1 MSTID Full Seasonal Variability

Figure 7.3a–d plots the wave parameters ($\tau$, $\lambda_H$, $c_H$, and $\phi$) for each MSTID event (gray dots) as a function of time for the duration of this 32-month study, as well as the mean altitude (Figure 7.3e) for each event. Monthly averages are shown as solid black lines. All plots utilize the same time scale to facilitate comparison. Note the reduction in number of measurements after January 2012 when fewer IPY measurements were conducted, as noted in Chapter 4, Section 4.4. The observed wave period (Figure 7.3a) varied little over the course of this 32-month study ranging from 30–60 minutes for nearly all events. The horizontal wavelengths (Figure 7.3b) and phase speeds (Figure 7.3c) also exhibited no significant variation in their mean values during this extended period. However, the directions of propagations clearly exhibited significant variability with time, as indicated in Figure 7.3d.

During the nine months of continuous observations, the MSTIDs are seen to propagate more southward (angles increase to $\sim 150^\circ$) and then return to eastward propagation ($\sim 100^\circ$) during the months March 2011 to September 2011. The directions of propagation increase again during the months September 2011 to April 2012, suggesting a cyclic behavior.

To further investigate the seasonal variability of the MSTID propagation directions during this period, we summed the data into three seasons: summer, winter, and equinox (spring + fall), each comprising four months of the year. Figure 7.4 shows polar plots with the results for the three winter seasons (November to February), three summer seasons (May to August), and three equinox periods (March to April and September to October). The results reveal striking seasonal changes in the dominant wave propagation direction.
Figure 7.3: Measured wave parameters from PFISR as a function of time over the 32-month duration of this study. (a) Periods, (b) horizontal wavelengths, (c) horizontal phase speeds, (d) wave azimuths, and (e) mean altitude of PFISR experiment. Each dot represents a single MSTID event with solid lines denoting monthly averages.

The MSTIDs observed during the winter months (Figure 7.4a) propagated predominantly southward, with 45% of the waves propagating within ±30° of 165° azimuth. In contrast, during the summer months (Figure 7.4c) the waves exhibit a strong preference for eastward propagation, with ~50% of the waves propagating within ±30° of due east. The spring and fall equinox periods (Figure 7.4b) yield similar results to each other, with the waves mainly propagating southeastward, with 45% propagating within ±30° of 120° azimuth. This figure further demonstrates the sparseness of MSTIDs propagating in the southwest through northwest quadrants (also see Figure 7.1e). Importantly, our full seasonal measurements suggest a consistent cyclic picture with eastward propagation during the summer months,
transitioning to southeastward during the fall equinox periods and then to predominantly southward during the winter seasons. Thereafter, the cycle continues by transitioning back to southeastward during the spring and returning to eastward during the summer, as observed over nine consecutive cycles.

7.6 PFISR Comparisons with Published Results

Our novel seasonal findings build and extend on the results of Ishida et al. [2008] and Frissell et al. [2016] who used SuperDARN daytime HF radar data to determine strong southward MSTID propagation during only the winter/fall periods.

Our findings obtained over several seasons, indicate strong seasonal variability in the MSTID propagation directions at this high-latitude location. Although there are several previous climatological studies of MSTIDs at mid and low latitudes [e.g., Crowley et al., 1987; Ding et al., 2011; Frissell et al., 2014; Ishida et al., 2008; Kubota et al., 2011], seasonal measurements at high latitudes are sparse and their properties are not well known.

Hernández-Pajares et al. [2012] reported a long-term climatology study of MSTIDs observed using TEC measurements at high, mid and low latitudes. For their high northern latitude study, they used data from 13 GPS receivers located in southern Alaska recorded from 2004 to 2011 (see Table 7.1). In their seasonal study of the phase velocities, they determined that the fall and winter propagation directions are mainly southeastward. Their
data are presented in scatter plots and indicate many events during the winter season that agree fully with our results (Figure 7.4a, b). In contrast, their summertime scatter plots are sparse and indicate MSTIDs propagating towards the northwest and the southeast, in contrast to our measurements of strong eastward propagation. This may be because the primary contributions to the TEC measurements are at the F peak, which is at much higher altitudes than our measurements here.

The SuperDARN network of HF radars has provided extensive high latitude studies, enabling a quantitative comparison with our PFISR results. SuperDARN radars use the ground scatter technique to image MSTIDs when ionospheric conditions enable observations of the F region [e.g., Bristow et al., 1994]. This condition occurs primarily during daylight hours in the fall and winter months. We will now compare the propagation directions of the observed MSTIDs with those reported by Ishida et al. [2008] and Frissell et al. [2016]. Figure 7.5 shows the sites of PFRR (labeled PF), the two radars used by Ishida et al. [2008] (labeled IS), and the four radars used by Frissell et al. [2016] (labeled Fr).

The study by Ishida et al. [2008] reported 15 months of data during the fall and winter (December 2003 to February 2007) from the Kodiak and King Salmon radars in southern Alaska (see Table 7.1 for details). This allows for a direct comparison with our TID results for the winter/fall period. Their Figure 3a summarizes the results over three consecutive

Figure 7.5: Map of the Northern Hemisphere showing locations of PFRR PFISR (red dot), locations of SuperDARN radars used by Ishida et al. [2008] (two black dots), and locations of SuperDARN radars used by Frissell et al. [2016] (four blue dots).
winter seasons comprising 134 MSTID events. Figure 7.6 shows the distributions from each study. We see that our observed wave parameters are very similar to their results (see Table 7.1 for a comparison of the typical MSTID wave parameters). Importantly, Ishida et al. [2008] also found that the wintertime propagation directions were strongly southward, with a noticeable southwestward component.

To illustrate the high correlation with our results, Figure 7.7a overlays the azimuthal propagation distributions from Ishida et al. [2008] (red) with this PFISR study (gray) for the winter period (November to February). The data compare very favorably. It is interesting to note that Ishida et al. [2008] also show very few northward waves. Since the Ishida et al. [2008] results span a period of 10 years, Figure 7.7a strongly suggests that the climatology for high latitude MSTIDs is reasonably stable during the winter at this geographical location. Figure 7.7c shows the monthly averaged Kp indices (obtained from NASA’s OMNIWeb)

![Figure 7.6](image)

**Figure 7.6:** Comparisons of the distributions of wave parameters from PFISR (black), Ishida et al. [2008] (red), and Frissell et al. [2016] (blue). (a) Periods, (b) horizontal phase speeds, (c) horizontal wavelengths, and (d) vertical wavelengths.
Figure 7.7: Comparisons of the wave azimuth distributions of PFISR and SuperDARN radars along with the amount of geomagnetic activity during those studies. (a) Comparison of propagation directions with Ishida et al. [2008] (red bars) (auroral region measurements) (2003–2007) and PFISR (gray bars). (b) Comparison of the propagation directions with Frissell et al. [2016] (blue bars) (high latitude measurements) (May 2012 to May 2015) and PFISR (gray bars). (c) Kp (gray line). The overlaid red line is the date range for Ishida et al. [2008], the overlaid blue line is the date range for Frissell et al. [2016] and the overlaid black line is the date range for this PFISR study.

during this 10-year period. The Ishida et al. [2008] measurements were obtained during slightly more active geophysical conditions than in our study (which ranged from ~1–3).

In Table 7.1, we also include results from Kubota et al. [2011], who observed MSTIDs on 44 nights over central Alaska using OI (630 nm) emission (mean altitude $\bar{z}$~250 km) during one winter season. They reported strong southward propagation with a significant westward component. The Kp indices for that study were comparable to those by Ishida et al. [2008] (not shown). Using ray tracing methods, Kubota et al. [2011] suggested a source region around the lower latitude edge of the auroral oval.
Most recently, Frissell et al. [2016] made comprehensive observations of winter/fall daytime MSTIDs using a network of SuperDARN radars located at mid and high latitudes (Table 7.1). Their measurements, limited to winter/fall, also confirm strong southward propagations for both high and mid latitude MSTID events. Figure 7.7b compares their propagation directions (blue) with our results (gray) over the period from November to April. The southward propagation directions agree exceptionally well; however, our PFISR study indicates more events with an eastward component of propagation. For further comparison, Figure 7.7c (blue) shows the Kp index for this study in comparison with our study (gray). Although the Ishida et al. [2008] results were obtained under slightly stronger Kp conditions, all three data sets show strong southward wave propagation during the winter months. In addition our PFISR data show strong eastward propagation during the summer months with a clear transition to southeastward during the equinox periods.

7.6.1 Critical Level Wind and Dissipative Filtering

To investigate the strong seasonal dependence of the observed MSTID propagation directions at PFISR, we again considered the effects of filtering due to (a) critical level wind and additionally (b) dissipative filtering. The critical level wind filtering mechanism is well known, as discussed in previous chapters, and can significantly affect the spectrum and directionality of the GWs as they propagate upwards. To recap, when the observed phase speed of a GW is equals the wind speed in the direction of the wave propagation, the wave is absorbed into the mean flow [e.g., Booker and Bretherton, 1967; Crowley et al., 1987; Francis, 1973; Fritts, 1978; Hazel, 1967; Hines, 1960; Hines and Reddy, 1967; MacDougall et al., 2009; Vadas, 2007; Waldock and Jones, 1984]. Thus, GWs propagating against the wind are favored, as long as the GW intrinsic period is longer than the buoyancy period. The dissipative filtering mechanism occurs when viscosity is strong, and thus is very important in the more rarified thermosphere. In this mechanism, those GWs propagating against the wind have larger vertical wavelengths, and therefore, are affected less by molecular viscosity than those GWs propagating with the wind (that have smaller vertical wavelengths). Thus, GW propagation against the wind is also favored in the thermosphere, and these two
mechanisms work together to filter those GWs propagating in the same direction as the wind.

Because there are no direct background wind measurements, we have used monthly averaged HWM14 [e.g., Drob et al., 2015] meridional and zonal winds as introduced in Chapter 2. We construct critical level blocking diagrams (showing regions where wave propagation is restricted by the winds [e.g., Taylor et al., 1993]) as a function of altitude. HWM14 and critical level filtering are introduced in Chapter 2. Figure 7.8 shows the results. The shaded regions indicate where we would not expect to observe MSTIDs up to an altitude of ∼300 km, assuming they originated in the lower atmosphere. Also plotted in this figure are the observed phase velocities from each MSTID event (shown as black dots in Figure 7.8). The top row in Figure 7.8 shows the blocking for each of the summer months, the middle row depicts the monthly wintertime blocking, while the bottom row shows the

![Figure 7.8](image)

**Figure 7.8**: Seasonal blocking diagrams (shaded regions) for each month using HWM14 monthly winds with the 652 MSTID events (black dots) from PFISR.
blocking for the equinox months.

During the summer months, the neutral winds are large (~200 m/s) and are westward, creating a large blocked region in the westward direction. This agrees with our result that very few MSITDs are seen propagating in this direction, and instead, nearly all the observed MSTID events propagated with a large eastward component. During the winter months, the winds remain westward, but are much smaller (~100 m/s), while some MSTIDs propagated northeastward and the majority of the waves exhibited strong southward motion. During the equinox months, the neutral winds were ~150 m/s and westward. During this time the waves remained eastward, but with an overall southeastward preference. All of these data are consistent with the effects of critical level and dissipative filtering for the observed GW propagation directions, but this is clearly not the only process controlling the observed distribution [e.g., Crowley and Rodrigues, 2012; Crowley et al., 1987; Del Genio et al., 1979; Francis, 1975; Friedman, 1966; Fritts and Vadas, 2008; Heale et al., 2014; Hines and Hooke, 1970; Klostermeyer, 1972; Pitteway and Hines, 1965; Vadas and Fritts, 2005; Zhang and Yi, 2002]. For example, so few westward waves occurred during the winter when the blocking was reduced. To our best knowledge, this study provides the most conclusive evidence for the effects of the background winds on the MSTIDs.

Evidence for other systematic effects is also present in Figure 7.8. Closer examination of the individual MSTID events (black dots) on a month-by-month basis (throughout this three-year summary) suggests a systematic broadening and shrinking in the azimuthal distribution of the wave events during the course of the year. Starting in January, the events are nearly all clustered in the SE quadrant, as depicted in Figure 7.8. By February/March, the event cluster has broadened northward and westward in its azimuthal extent, April exhibits further broadening of the azimuthal spread in the events, and by May MSTIDs are seen to propagate over all eastward azimuths ranging from ~0–180°. This general situation prevails throughout the summer months, including September. However, by October/November the azimuthal distribution has reduced significantly with most events propagating in the SE quadrant. By December the distribution resides primarily in the SE quadrant, exhibiting
similar characteristics to the January distribution. This additional annual behavior, suggest-
ing broadest azimuthal variability during the previously little-measured summer period, may be associated with seasonal changes in the abundance and sources of the MSTIDs and is currently under further investigation.

7.6.2 Vertical Wavelengths

For our study, we have used the Nicolls and Heinselman [2007] method to calculate the vertical wavelengths as described in Chapter 4. This enables a statistical comparison of the vertical wavelengths with those obtained from the dispersion relation using our large data set. Chapter 4 describes how the vertical wavelengths are measured using PFISR and how the dispersion vertical wavelengths are derived. We have improved on the Nicolls and Heinselman [2007] method by incorporating daily-averaged HWM14 horizontal winds [e.g., Drob et al., 2015] in the dispersion relation. We have also included daily-averaged neutral temperatures and densities obtained from the NRLMSISE-00 [e.g., Picone et al., 2002] model, which included F10.7 data from NASA’s OMNIWeb [e.g., Mathews and Touheed, 1995].

Figure 7.9a shows a histogram of the vertical wavelengths derived from the PFISR data. The distribution shows a broad peak around 200–300 km, while the majority of the vertical wavelengths are <500 km. In comparison, Figure 7.9b shows the vertical wavelengths derived from inputting the GW horizontal parameters with winds, temperatures, and densities from the HWM14 and NRLMSISE-00 models, respectively, into the GW dissipative dispersion relationship [e.g., Vadas and Fritts, 2005]. The distributions are clearly very similar, but the dispersion relation peak occurs at somewhat shorter vertical wavelengths (~150–200 km). This difference is likely due to the fact that the neutral winds from HWM14 are climatology winds derived from limited data in the lower thermosphere (below z~220 km) [e.g., Drob et al., 2015; Hedin, 1991] due to lack of measurements. Thus, it is not expected that these neutral winds are accurate on an hour-by-hour or day-by-day basis.

Figure 7.10 builds on this analysis by comparing the MSTID vertical wavelengths (black line) to those derived from the dispersion relation vertical wavelength (gray line); both
Figure 7.9: Distributions of the (a) PFISR measured vertical wavelengths and (b) derived vertical wavelengths using a dispersion relation for the thermospheric region. The measured vertical wavelengths exhibited a mean of 273 km, a median of 216 km, and interquartile range 124–331 km. The dispersion vertical wavelengths exhibited a mean of 141 km, a median of 133 km, and interquartile range 93–178 km. Neutral temperatures and densities input into the dispersion relation were obtained from NRLMSISE-00 with F10.7 data from NASA’s ONMIWeb. Neutral winds input into the dispersion relation were obtained using daily averaged HWM14 winds.

are averaged in 50 km altitude bins. The dispersion vertical wavelengths continue to use HWM14 and NRLMSISE-00 data. The observed and theoretical vertical wavelengths are in reasonable agreement, and both show a systematic increase in vertical wavelength in altitude. This is consistent with previous observations, which reported the vertical wavelength increasing significantly with altitude in the thermosphere [e.g., Djuth et al., 1997; Oliver et al., 1997]. It is also consistent with theoretical results of Vadas [2007], which show that the increase of the vertical wavelength with altitude occurs because of dissipative filtering of GWs due to kinematic viscosity and thermal diffusivity.

7.6.3 Kp versus Propagation Direction

In order to estimate the level of geomagnetic activity, Figure 7.11a (gray dots) shows a 30-day running mean of daily averaged Kp indices. The Kp indices were obtained from NASA’s OMNIWeb [e.g., Mathews and Towheed, 1995]. These averaged Kp indices are below three, which implies that there was weak/low geomagnetic activity during the time
the MSTIDs occurred. To further focus this comparison, the Kp indices are averaged over the duration of each MSTID event; we find that \(~\)90\% of the observed MSTIDs occurred when the Kp index was \(<3\).

A visual comparison of the 30-day running mean of the Kp indices and wave azimuth, suggests a correlation between the magnitude of the Kp and the dominant wave propagation direction during this extensive 32-month study, which comprised nine seasonal changes. To help investigate this possible relationship, Figure 7.11a shows the results of a correlation analysis between these two data sets. The daily-averaged propagation directions were calculated with a 30-day running average. The propagation directions (black dots) along with Kp indices (gray dots) show remarkable agreement in both the broad (several month) and finer scale (month-to-month) structure and relative changes. Furthermore, a Pearson correlation analysis results in a coefficient of 0.72, which is significant. This is an intriguing result, because there have been many prior studies investigating the relationship between the occurrence of MSTIDs with geophysical activity [e.g., Chimonas and Hines, 1970; Crowley et al., 1987; Francis, 1974, 1975; Frissell et al., 2014, 2016; Grocott et al., 2013; He et al.,
Figure 7.11: Time series of the PFISR wave azimuths, geomagnetic activity, and wave occurrence frequency. (a) Daily-averaged propagation directions with a 30-day running average (black dots) and daily-averaged Kp indices from NASA’s OMNIWeb with a 30-day running average (gray dots). (b) Number of MSTID events within each month.

More recently, Frissell et al. [2016] investigated whether or not space weather had an effect on MSTID occurrence. They concluded the AE and SYM-H indices did not correlate with MSTID occurrence. The new availability of our year-round MSTID measurements, as presented herein has opened the door to new correlative investigations. This is further illustrated in Figure 7.11b, which shows a histogram of the number of MSTID events observed within each month.

7.7 Summary

Over 650 MSTID events were observed by PFISR over a ~three year period enabling a comprehensive study of their characteristics and seasonal variabilities. We have determined the observed period, horizontal wavelength, phase speed, and propagation directions as a function of altitude (ranging from z=100–300 km) for all these events. The altitude averaged
periods exhibited a mean of 41 min with typical values ranging from \( \sim 37 - 44 \) min, horizontal wavelengths with a mean of 446 km ranging from \( \sim 369 - 514 \) km, and phase speeds with a mean of 187 m/s ranging from 155–212 m/s. Note, these values are all significantly higher than the GW events observed in the mesosphere, as discussed in Chapters 5 and 6. The wave periods and azimuths were found to be nearly constant with altitude, while both the horizontal wavelengths and the phase speeds exhibited \( \sim 20\% \) increase with altitude.

The distribution of vertical wavelengths exhibits a broad peak around 200–300 km, with the majority of the vertical wavelengths \(< 500 \) km. These measured vertical wavelengths were compared with calculated vertical wavelengths using a thermospheric GW dissipative dispersion relation [e.g., Vadas and Fritts, 2005]. We also investigated the vertical wavelengths as a function of altitude for all MSTID events. The observed and theoretical vertical wavelengths were found to be in reasonable agreement, and show a systematic increase with altitude.

While the altitude averaged observed wave periods, horizontal wavelengths and phase speeds are consistent over the course of this 32-month study, the propagation directions reveals striking seasonal changes in the dominant wave propagation directions. Importantly, during the summer months, the waves exhibit a strong preference for eastward propagation, adding new results on MSTIDs to the literature. In contrast, the MSTIDs observed during the winter months propagated southward. Together, these data suggest a consistent cyclic picture with eastward propagation during the summer months, transitioning to southeastward during the equinox periods and then to predominantly southward during the winter seasons. Thereafter, the cycle reversed back to eastward during the summer.

The effects of critical level wind and dissipative filtering of the MSTID propagation directions were investigated using the HWM14 winds. The blocked regions are towards the west during all seasons and agrees with our results of eastward wave propagation. To our best knowledge, this study provides the most conclusive evidence for the effects of the background winds on the MSTIDs.

In conjunction with the strong summertime eastward propagation peak, there is also
a large broadening in the azimuthal spread of the observed MSTIDs. This spread systematically reduces during the equinox periods to a minimum in the winter season, where the MSTIDs are most highly focused on southward propagation. This additional finding, together with the observed near total lack of waves propagating westward, points toward (even under reduced wind-blocking conditions) additional causes of the observed propagation anisotropy, most likely associated with the wave sources, which are currently under investigation. Finally, there have been many studies investigating the relationship between the occurrence of MSTIDs with geophysical activity. In this study, we have identified an unexpected strong correlation between the MSTID propagation directions and the auroral Kp index, which is also under further investigation.
8.1 Introduction

In this chapter, we investigate gravity wave (GW) properties obtained from the PFRR all-sky airglow imager (ASI), the ALOMAR AMTM, and PFISR. Specifically, we have used coincident data obtained during the winter season 2011/2012 to make a novel comparative study.

8.2 Gravity Wave Comparisons

Figure 8.1 shows the locations of PFRR (red dot) and ALOMAR (blue dot), illustrating their similar high latitudes (65°–69° N), but large longitudinal separation, (>160°) greater than 5000 km. The ASI and AMTM imager results were obtained using the OH emission (at a nominal altitude of ∼86 km), while the PFISR thermospheric results were obtained over the altitude range ∼100–300 km.

Figure 8.2 shows the derived GW parameters in the same format as Figures 5.3, 6.2, and 7.3, but for comparison, now limited to the winter season 2011/2012, characteristic of individual wave events observed by PFRR ASI (black dots), AMTM (blue), and PFISR (red), are plotted as a function of time, together totaling over 500 events with the majority (450 events) from the mesosphere.

Figure 8.3 shows the distributions of the observed wave parameters as line plots using the same color identification scheme as Figure 8.2. As expected from the results in previous chapters, the distributions (λ_H, c_H, and τ) for both the ASI and AMTM data are similar in shape and both peak at almost the same values. The thermospheric wave (MSTID) parameters obtained from PFISR are also plotted in red and exhibit significantly larger values with peak observed periods (plot a) and phase speeds (plot b) about four to five times larger, while the horizontal and vertical wavelengths are about 20 times larger. (Note the different scales used for the mesospheric and thermospheric wavelengths in plots c and d.)
These results clearly establish that the waves detected during the same time interval at mesospheric and thermospheric altitudes are from different parts of the upward-propagating GW spectrum during this winter season.

Investigations of the mesospheric GWs at PFRR and ALOMAR (separated by 5,000 km) showed remarkably similar wave propagation distributions during the same time period. This unexpected result is illustrated in Figure 8.4 top row. At each site, westward wave motion dominated throughout the winter months, mainly comprising low phase speed waves, while eastward propagating waves generally exceeded their blocking background winds. Importantly, at both sites, wave propagation to the northwest was most common and comprised relatively low phase speed waves ($<40$ m/s). Both sites also show a second wave preference to the northeast comprised significantly higher phase speed ($>40$ m/s) waves. A third less prominent peak to the southeast contained intermediate phase speeds is more evident in the wave phase speed vectors (Figure 8.4 bottom row). While the preference for westward propagation is consistent with similar levels of critical level wind filtering experienced at both sites, it is surprising that they exhibit similar azimuthal preferences given the large site
Figure 8.2: GW parameters from the PFRR ASI (black dots), ALOMAR AMTM (blue dots), and altitude averaged PFISR (red dots) as a function of time. (a) Periods, (b) horizontal wavelengths, (c) horizontal phase speeds, and (d) azimuths.

separation. This could suggest that both sites had similar sources for the waves, especially the northwestward and northeastward propagating events. The dominant sources of these waves is currently unknown and has yet to be investigated.

In contrast, comparison of the mesospheric wave distributions with the larger-scale thermospheric MSTID events reveals an almost complete opposite preference for the observed wave propagation distribution, with strong southeastward motion at thermospheric heights during the extended winter season (October to March). Indeed, very few thermospheric waves propagated towards the northwest, which was the preferred direction for the wintertime mesospheric waves. This is most probably because the waves seen at both altitudes comprised different parts of the upward propagating wave spectrum, as noted above.
In particular, the low-velocity northwestward mesospheric waves were not able to penetrate deeply into the thermosphere due to strong westward wind filtering, while the higher-velocity mesospheric waves that propagated towards the southeast are free to propagate upwards into the lower thermosphere, consistent with our GW observations by PFISR.

### 8.3 Comparisons of Other Mesospheric Data

Similar type, but much more limited, GW studies have previously been performed at Resolute Bay (74° N) [e.g., Suzuki et al., 2009] and at Longyearbyen, Svalbard (78° N) (M. Dyrland private communication). A map showing the location of these two additional Arctic sites and their strategic locations between Alaska and Norway are plotted in Figure 8.1. A total of 143 small-scale (<100 km wavelengths) GW events were observed in the Na emission
Figure 8.4: Wave azimuth (top row) and phase velocity (m/s) (bottom row) distributions for PFRR ASI (gray), ALOMAR AMTM (blue), and PFISR (red).

(nominal altitude 90 km) from Resolute Bay during the winter seasons of 2005/2006. A total of 59 events were observed in the OH emission (in close proximity to the Na emission) during the period 2010–2012 from Longyearbyen. Figure 8.5 plots the distributions of the observed wave parameters from each site [PFRR (black), ALOMAR (gray), Resolute Bay (blue), and Svalbard (red)]. To aid this comparison the distributions have been normalized by the total number of events and show similar characteristics for the observed period and phase speed, but with somewhat larger horizontal wavelengths for Resolute Bay, and all with similar ranges.

Figure 8.6 top row compares the wave azimuth distributions for each of the four sites, while the bottom row plots their corresponding phase velocities. Note, the full distributions for the PFRR and AMTM imagers are used for this comparison with the Resolute and Svalbard data, which were taken over an extended period of time. The strong preference for westward propagation is again evident in these more limited data sets, as expected from
Figure 8.5: Comparisons of the distributions of the MLT wave parameters for PFRR (solid black), ALOMAR (gray), Resolute Bay (blue), and Svalbard (red). (a) Periods, (b) horizontal phase speeds, (c) horizontal wavelengths, and (d) vertical wavelengths. Each distribution has been normalized by the total number of events of each data set.

Critical level wind filtering. Furthermore it appears that northwestward motion predominates, as also observed in the ASI and AMTM data. Essentially, these distributions are similar in their propagation characteristics, further contributing to the northern hemisphere high-latitude mesospheric GW climatology.

In their paper, Suzuki et al. [2009] also investigated possible source regions. Potential auroral source regions were studied by comparing wave occurrence with the Kp index. They reported no association with auroral activity. They also investigated wave propagation from lower atmospheric sources by considering the effects of wind filtering. They found that the majority of the wave events propagated westward (∼65% estimated from their Figure 4d), but remained unidentified. They also concluded that the westward propagation was consistent with critical level filtering. No further analyses have been reported on the Svalbard data to date.
Figure 8.6: Comparisons of the distributions of wave azimuths (top row) and phase velocities (bottom row) from PFRR (gray), ALOMAR (blue), Svalbard (red), and Resolute Bay (gray). The distributions contain data from the total combined data sets (meaning they are not limited to the time period of the ALOMAR study).

8.4 Effects of Sudden Stratospheric Warmings on Wave Propagation

One of the most dominant large-scale perturbations in the Arctic winter atmosphere is the occurrence of Sudden Stratospheric Warming (SSW) events [e.g., Chandran et al., 2011; Hoffmann et al., 2007; Limpasuvan et al., 2011, 2012]. During the Arctic winter, the predominant stratospheric wind pattern is the polar vortex: a powerful cyclonic circulation (eastward winds), which strengthens during the fall due to the decrease of solar radiation, reaches its peak in the midwinter, and weakens in the spring. An SSW event is a temporary breakdown of the polar vortex resulting from interactions between planetary waves and the zonal mean background flow, and can significantly alter the dynamical state of the atmosphere. Furthermore, due to the wind strength and reversal, the SSW events can play an important role in the vertical propagation of the GWs at high latitudes via changes in critical layer filtering. During major SSWs, a strong reversal of the zonal mean zonal wind occurs at 60° latitude at or below 10 hPa. (Note, a minor warming event is characterized more by the temperature gradients at or below 10 hPa, and reversal in the zonal mean zonal...
wind is absent, but local reversals in the wind can occur, as discussed below.)

8.4.1 PFRR

To investigate the occurrence of SSWs above PFRR, the midwinter (December to February) local zonal flow (using the MERRA winds averaged over a $10^\circ \times 10^\circ$ latitude-by-longitude region) during the three consecutive winters from 2010–2013, are plotted in Figure 8.7. As previously noted in Chapter 5, the winter 2010–2011 experienced no SSWs and was dominated by eastward winds, while Figure 8.7 reveals two localized SSW events in January 2012 and 2013. During the period 8–23 January 2012 (duration of 15 days), the local zonal winds over Alaska exhibited a temporary, but strong, reversal to westward winds with a peak amplitude up to $\sim 90$ m/s. For the following winter, a major SSW occurred several days earlier (1–20 January 2013) and lasted about a week longer, again exhibiting high westward winds, typically $\sim 50$ m/s, during this time.

To further examine the effects of these two SSW events, we have analyzed MERRA

![Figure 8.7: MERRA zonal winds from December to February over PFRR capturing the temporary reversal in the wind field (blue contours) associated with the two SSW events observed in 2012 and 2013.](image-url)
winds, now focusing directly on the SSW events. The resultant blocking diagrams are shown in Figure 8.8, together with the wave events detected during the SSWs. Both plots distinctly show the change over to primarily westward blocking and the preponderance for low-velocity (<30–50 m/s) eastward waves temporarily enabled during each of the SSW events. This said, the distributions in Figure 8.8 also reveal a preference for southward wave propagation during the 2012 SSW and northward during the 2013 SSW, possibly due to different dominant sources during these two winter seasons. In addition to the changes in wave propagation direction, it is also important to note that the number of wave events observed during both of these SSWs (total 31) was found to be twice as many as compared with other observation nights in January 2012 and 2013. Summarizing the effects of the SSWs and their associated local wind reversals were evident in the monthly averaged blocking regions (Figure 5.10), which were significantly reduced compared with other winter months, apparently allowing for more isotropic wave propagation.

8.4.2 ALOMAR

Figure 8.9a–c shows the variation of the intrinsic wave parameters and vertical wavelengths as a function of time at ALOMAR. The intrinsic phase speeds, plot b, and vertical

![Figure 8.8: Blocking diagrams (shaded region) for the (a) 2012 minor, and (b) 2013 major SSW events along with individual GW events (red) obtained with the PFRR ASI. Note the local reversal in winds during both of these events and their effects on the observed wave propagation distribution.]
Figure 8.9: Intrinsic horizontal GW parameters and vertical wavelengths as a function of time obtained from ALOMAR during the winter season 2011/2012. (a) Intrinsic periods, (b) intrinsic phase speeds, (c) vertical wavelengths, and (d) combined MERRA and Saura zonal winds. Gray dots represent individual GW events and the solid lines represent monthly averaged wave parameters. Red contours indicate eastward winds while blue contours are westward winds.

Wavelengths, plot c, exhibit significant scatter during the period leading up to the SSW (i.e., from December to mid-January). During the SSW, these values decreased suddenly and appeared to rebound when the stratospheric winds returned to eastward (i.e., after the end of January), as shown in plot d. These results suggest that the SSW played a key role in the observed changes of the intrinsic phase speed and associated vertical wavelengths of the mesospheric GWs.

Building on the SSW results from PFRR, Figure 8.10a plots the GW blocking diagram for ALOMAR using combined MERRA and Saura radar winds together with the waves
observed during the 2012 SSW. During this SSW, a large number 36 GW events (similar to the total observed at PFRR), were detected and measured, again indicating a doubling in their occurrence frequency during the SSW. Figure 8.10, plots the vector blocking diagram for this SSW event. The 36 wave events show a strong preference for northwest propagation with relatively high observed phase speeds (blue, up to 70 m/s) opposite to the direction of the prevailing wind. This is also the preferred direction of motion for the waves observed at PFRR and ALOMAR during the winter months (as discussed earlier in this Chapter).

Figure 8.10 replots the PFRR blocking diagram during the 2012 SSW (Figure 8.8a) as a vector plot for comparison with the 2012 ALOMAR results (Figure 8.10). In this case, the majority of the waves observed at ALOMAR during the same SSW event propagated toward the northwest, consistent with the blocking to the southeast, but opposite to that observed at PFRR. Thus, while our earlier comparative study established similar propagation directions for the mesospheric GWs observed during the winter at ALOMAR and PFRR (5,000 km separation), the GW events that occurred during the SSW period are almost opposite to each other. This further emphasizes the importance of local background winds on the observed GW distributions at any given time.
8.5 Summary

In this chapter, the results of three GW climatology studies from 2011/2012 winter season using PFRR ASI, ALOMAR AMTM, and PFISR data have been compared to further investigate GW characteristics and propagation. These three studies allowed a novel comparison of short-period GWs observed at two widely spaced geographic locations and two different altitude regions. It was determined that the distribution (\( \lambda_z \), \( c_H \) and \( \tau \)) of the mesospheric waves from ASI and AMTM were very similar, as previously reported. However, it was most unexpected that the preferred directions of propagation were very similar at these sites separated by over 5000 km, suggesting common high latitude GW sources. Limited data from prior studies at two additional sites (Resolute Bay and Longyearbyen) further support this finding. In contrast, the thermospheric GW scale sizes were much larger than those of the mesospheric GWs, and their propagation headings, as measured by PFISR, showed dominant southeastward motion opposite to the primary mesospheric GW motions. This was most probably because the waves seen at both altitudes comprised different parts of the upward-propagating wave spectrum. The change in the wind direction in the thermosphere, and the associated wind filtering, enabled only the southeastward waves to propagate to higher thermospheric altitudes, consistent with our GW observations by PFISR.

Finally, chance observations of two SSW events offer further insight on the effects of induced local/regional changes in the winds on the prevailing GW propagation directions observed in the MLT. The number of wave events detected during SSWs was found to increase by a factor of two at each site and for each SSW event, again most probably due to the change in the background wind.
CHAPTER 9
SUMMARY/FUTURE RESEARCH SUGGESTIONS

9.1 Summary

The primary objective of this research is to investigate the characteristics and propagation of gravity waves (GWs) in the upper mesosphere (87 km) and thermosphere (100–300 km) at high northern latitudes where high-quality measurements, to date, have been limited. To do this, new data from three instruments an ASI, an AMTM, and an PFISR strategically located at two well-separated sites (160° in longitude), have been used to identify and characterize over 1200 discrete GW events.

Imaging and radar measurements were made over a three year period (2010–2013) at the Poker Flat Research Range (PFRR) Alaska, while the AMTM measurements were used from ALOMAR, northern Norway, during a single winter 2011–2012. These data enabled a novel comparative study of the longitudinal variability in the wave characteristics at MLT heights, and the first investigation of GW climatologies at two adjacent regions in the upper mesosphere and thermosphere. The capability of PFISR to measure the MSTIDs during both the summer and winter months has provided essential new knowledge on the full seasonal thermospheric GW climatology.

The ALOMAR data, together with available radar and model wind data, have been used to investigate the intrinsic properties of the waves, including their distributions and vertical wavelengths. In both the mesosphere and thermosphere data, clear evidence was found for critical level filtering (by the background winds) on the observed wave motions. The net effect of the background wind caused significant changes in the wave azimuthal distribution, intrinsic phase speeds, and vertical wavelength spectrum during the winter season 2011/2012.

Investigations of the mesospheric GWs at PFRR and ALOMAR (separated by 5000 km) showed remarkably similar wave propagation distributions during the same time period. At each site, westward wave motion dominated throughout the winter months, mainly
comprising low phase speed waves, while the eastward propagating waves exhibited higher phase speeds, generally exceeded their blocking regions. Wave propagation to the northwest was most common and constituted relatively low phase speed waves (<40 m/s). A second wave preference to the northeast comprised significantly higher phase speed (>40 m/s) waves. A third, less prominent peak to the southeast, contained intermediate phase speeds.

In contrast, the comparison of the mesospheric wave distributions with the larger-scale thermospheric MSTID events revealed a completely opposite preference for the observed wave propagation distribution, with strong southeastward motion at thermospheric heights during the extended winter season (October to March). Indeed, very few thermospheric waves propagated towards the northwest, which was the preferred direction for the winter-time mesospheric waves. This was most probably because the waves seen at both altitudes comprised different parts of the upward propagating wave spectrum due to the change in the wind direction and the associated wind filtering enabling only the southeastward waves to propagate to higher thermospheric altitudes, consistent with our GW observations by PFISR.

Finally, chance observations of two SSW events offer further insight on the effects of induced local/regional changes in the winds on the prevailing GW propagation directions observed in the MLT. The number of wave events detected during SSWs was found to increase by a factor of two at each site and for each SSW event, again most probably due to the change in the background wind.

9.2 Future Research

The results from the PFRR ASI, ALOMAR AMTM, and PFRR PFISR have allowed us to perform a detailed investigation of high-latitude, short-period GWs at different geographic and attitudinal locations. Much has been learned about the propagation properties and the important role of the middle atmospheric winds on the wave propagation distributions. However, there are still significant unknowns concerning the sources of the waves observed in the mesosphere and thermosphere. Important new studies should include detailed ray tracing analysis of individual events to determine their dominant sources.
Also, the discovery of unusual electron density profiles during these analyses may pro-
vide new evidence for GW breaking in the mesopause region and seeding of secondary GWs
that can propagate to high thermospheric altitudes. These unusual events appear as a
discontinuity in the vertical extent of the thermospheric waves as measured by PFISR on
25 September 2010 as shown in Figure 9.1. On this occasion, the discontinuity in the phase
fronts occurred at an altitude of \(\sim 250\) km. As many as 50 candidate secondary wave events
have been identified in the PFISR data as shown in Figure 9.2. The climatology of these
secondary waves will be investigated as part of a future study.

![Figure 9.1: Band-passed relative electron density perturbations obtained by PFISR from
25 September 2010 illustrating a possible secondary GW event.](image)
Figure 9.2: Horizontal wave parameters obtained from PFISR as a function of time with gray dots representing all MSTID events and red dots representing possible secondary wave events. (a) Periods, (b) horizontal wavelengths, (c) horizontal phase speeds, and (d) azimuths.
REFERENCES


Frissell, N. A., J. B. H. Baker, J. M. Ruohoniemi, R. A. Greenwald, A. J. Gerrard, E. S. Miller, and M. L. West (2016), Sources and characteristics of medium-scale traveling


Herzberg, G. (1951), The atmospheres of the planets (with plates III and IV), *J. Royal Astro. Soc. Canada*, 45, 100.


Nielsen, K. (2007), Climatology and case studies of mesospheric gravity waves observed at polar latitudes, PhD dissertation, Utah State University, U.S.A.


Rawer, K. (1957), The ionosphere: its significance for geophysics and radio communications, Burns & Oates.


APPENDIX
### Figure A.1: Copyright permission for use of Figure 3.4. [Peterson and Kieffaber, 1973a].
CURRICULUM VITAE

Michael Ray Negale

Education

**Ph.D., Physics**, Utah State University, Logan, UT, 2018  
Concentration: Upper atmospheric dynamics and variability  
Dissertation Title: Investigating the Climatology of Mesospheric and Thermospheric Gravity Waves at High Northern Latitudes  
Major Advisor: Michael J. Taylor, Ph.D.

**B.S., Physics**, Utah Valley University, Orem, UT, 2011

**B.S., Mathematics**, Utah Valley University, Orem, UT, 2011

**A.S., Computer Science**, Utah Valley University, Orem, UT, 2008

Experience

**Graduate Research Assistant**, 2011–2017  
Utah State University, Logan, UT  
Dissertation Research: Study of the motions of the upper atmosphere (>85 km altitude) using passive optical and radar instrumentation located in central Alaska and northern Norway.

**Graduate Teaching Assistant**, 2011–2017  
Utah State University, Logan, UT  
Courses: Physics for Scientists and Engineers 1 & 2

**Instructor**, 2016–2017  
Utah State University, Logan, UT  
Course: Physics for Scientists and Engineers Lab 1

**Instructor**, Spring 2017  
Utah State University, Logan, UT  
Course: Physics of Technology

**Instructor**, Summer 2016  
Utah State University, Logan, UT  
Course: Physics for Scientists and Engineers 2

**Summer Internship**, 2011  
Space Dynamics Laboratory, Logan, UT  
Research Focus: Motions of the upper atmosphere/ionosphere (>100 km) using satellite data.
Undergraduate Teaching Assistant, 2010–2011
Utah Valley University, Orem, UT
Course: Physics for Scientists and Engineers Lab 1

Undergraduate Teaching Assistant, 2010–2011
Utah Valley University, Orem, UT
Course: Acoustics, Mechanics, Linear Algebra, Calculus 1 & 2, Abstract Algebra.

Summer Internship, 2009
Argonne National Laboratory, Argonne, IL
Research Focus: Calibration of surface meteorological instrumentation.

Publications


Research Activities

Research Visit: SRI Mike Nicolls Collaboration (Data Analysis), Jul. 2015.

Fellowships and Scholarships

National Science Foundation Graduate Research Fellowship, 2013–2015
Gene Adams Endowed Graduate Fellowship, 2017
Howard Blood Endowed Graduate Fellowship, 2016, 2017
Navajo Nation Graduate Fellowship, 2011–2013
American Indian Center Graduate Fellowship, 2011–2013
Navajo Nation Chief Manuelito Undergraduate Scholarship, 2008–2011
American Indian Services Undergraduate Scholarship, 2008–2011
Utah Valley University Multicultural Undergraduate Scholarship, 2007–2008
Awards and Honors

Winds of Change Cover Article (2014)
APS Meeting of the Four Corners Best Student Poster Award (2012)
Graduated UVU Magna Cum Laude in Physics and Mathematics (2011)
UVU Deans List (2008–2011)
Outstanding UVU Student of the Year in Physics (2010)
Outstanding UVU Student of the Year in Mathematics (2009)
Graduated UVU with high honors in Computer Science (2008)

Public Outreach

Mentor: Native American Summer Mentorship Program (NASMP), Summer 2015, 2016, 2017

USU Eclipse Team: Rigby, ID, 2017

Judge: Heritage Elementary STEM Fair, 2017

Judge: American Indian Science and Engineering Society (AISES) virtual science fair (NAIVSEF), 2016

Judge: Bridgerland Science Fair (K-12), 2015, 2016

Presented: Short talks on research at USU for Native Aggie Day, 2015

Reviewer: AISES Scholarship reviewer, 2015

Graduate Assistant: Physics Day at Lagoon, 2011–2017

Judge: Hillcrest Elementary Science Fair, 2015

First Author Research Presentations


Negale (2017), Seasonal characteristics of thermospheric gravity waves in the high latitude observed with the Poker Flat Incoherent Scatter Radar, Physics Colloquium, Logan, UT, Oct. 2017

Negale et. al., (2017), Seasonal propagation characteristics of MSTIDs observed at high-latitude over central Alaska using the Poker Flat ISR, CEDAR Workshop, Santa Fe, NM, Jun. 2017.


Negale, (2015), Thermospheric wind observations obtained from the Poker Flat Incoherent Scatter Radar, USU Student Research Symposium, Logan, UT, Apr. 2015.


