INVESTIGATING SMALL-SCALE DYNAMICAL FEATURES
IN THE MESOPAUSE REGION

by

Deepak B. Simkhada

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Approved:

______________________________  ______________________________
Michael J. Taylor                  Bela G. Fejer
Major Professor                    Committee Member

______________________________  ______________________________
Jan J. Sojka                        Blake Crowther
Committee Member                    Committee Member

______________________________  ______________________________
David Peak                          Byron R. Burnham
Committee Member                    Dean of Graduate Studies

UTAH STATE UNIVERSITY
Logan, Utah

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Investigating Small-Scale Dynamical Features in the Mesopause Region

by

Deepak B. Simkhada, Doctor of Philosophy
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Major Professor: Dr. Michael J. Taylor
Department: Physics

Utilizing analyses of observational data, we performed a detailed study to investigate short-period atmospheric gravity waves and ripples, and instabilities in the mesopause region. Recent coordinated measurements from Haleakala Observatory, Maui, HI via airglow wave imaging, meteor wind radar, and Na wind-temperature lidar have provided a unique dataset for this study. Gravity waves generated in the lower atmosphere propagate energy upwards into the mesosphere and lower thermosphere region where they modulate the airglow emissions, and later break and deposit their momentum, causing significant perturbations in winds and temperatures. Ripples, on the other hand, are likely generated by localized shear instabilities in the background wind field, and are also observed frequently in airglow images.

First, we investigated the frequency of occurrence and climatology of ripples and their seasonal variation in propagation directions of motion. These novel results were then used to characterize the nature of the observed ripples and to identify individual Kelvin-Helmholtz instability events as their sources, which were consistent with theoretical computations. The results provided insight to better quantify seasonal wave anisotropy and instability dynamics in the mesopause.

Second, we investigated ground-relative and intrinsic properties of gravity waves that were Doppler-ducted or evanescent. The results were then compared
with an analytical model and numerical model simulations for the Doppler-ducted and evanescent waves, demonstrating significant differences in wave structures for apparently similar wave events. Results reinforced the need to better understand the dynamics of wave ducting and their importance for momentum and energy transports at upper mesosphere and lower thermosphere heights.

Third, we investigated the frequency of occurrence of frontal disturbances and the appearance of complementary intensities in OH and O$_2$ airglow emissions. We have analyzed ducting conditions of mesospheric bores. One bore event was trapped in a stable Doppler duct arising from wind structure, and other bore events were trapped within thermal inversion layers. Analysis confirmed that the measured ducted bore events were separately consistent with theoretical Doppler-ducted and thermal-ducted wave characteristics. These studies provide an extensive dataset for detailing the morphology and dynamics of mesospheric frontal events.
Though this dissertation is an individual work, I could never have explored the depths I have without the invaluable help, support, and guidance of several people. First of all, I would like to express my sincere gratitude to my major professor, Dr. Michael J. Taylor, for the support and supervision he has given in the completion of my dissertation. Throughout this period, he always encouraged me during my difficulties, and I am fortunate to have him as my advisor and mentor. I would also like to convey my sincere gratitude to my committee members, Dr. Jan J. Sojka, Dr. David Peak, Dr. Bela G. Fejer and Dr. Blake Crowther, for their decisive assessments and important feedback in the development of my dissertation. I am also fortunate to have the opportunity to work with a group of researchers in Dr. Taylor’s Airglow Imaging Lab. I am especially grateful to my co-worker, Dr. Jonathan B. Snively, for his insightful comments and helpful suggestions on my research work through numerous discussions. I would like to express my sincere thanks to my other co-worker, Dr. Pierre-Dominique Pautet, who selflessly helped me many times in airglow imaging data analysis and software-related problems. I would also like to extend my special thanks to our research collaborators, Dr. Dave Fritts and Dr. Brian Laughman of Colorado Research Associates, and Dr. Alan Liu, Dr. Gary Swenson, and Dr. Steven Franke of the University of Illinois at Urbana-Champaign.

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7.13 Plots showing the vertical profiles of (a) zonal wind, and (b) meridional wind for the night of October 21, 2003 ................................. 167
1.1 The Earth’s Atmosphere

The Earth’s neutral atmosphere is a relatively thin (≈100 km), spherical, gaseous layer surrounding the Earth that is retained by the force of gravity. The atmosphere is denser at the surface and gets progressively rarified with increasing altitude. The vertical distribution of temperature, pressure, density, and composition of the atmosphere constitutes atmospheric structure. The pressure and density of the atmosphere decrease exponentially with altitude. Solar radiative and internal chemical processes set up within the atmosphere create a temperature profile that varies significantly with altitude, creating four distinct regions: the troposphere, stratosphere, mesosphere, and thermosphere. The boundaries separating these atmospheric regions, called pauses, correspond to major turning points in the temperature profile. Figure 1.1 shows a typical vertical profile of the Earth’s atmospheric temperature derived from a large number of in-situ and remote sensing measurements and summarized in the global MSISE90 atmospheric model [Hedin, 1991]. The data are given for ≈20° N latitude and show that the atmospheric temperature varies typically by over 120 K from the troposphere to the mesopause region (up to 100 km altitude), with even higher temperature changes at summer polar latitudes. Solar cycle variations also cause significant changes in the temperature profile, particularly in the thermospheric region [e.g., Andrews et al., 1987; Hargreaves, 1992].

The troposphere is the lowest region of the Earth’s atmosphere, extending from the surface to a height of ≈10–15 km. The atmospheric temperature decreases uniformly with altitude at a nearly constant lapse rate of ≈6.5 K km\(^{-1}\) throughout this region. This is due primarily to radiative and adiabatic cooling effects near
the Earth-air interface via small-scale convection. The main heat source within the troposphere is the absorption of solar-induced infrared radiation from the Earth's surface by the greenhouse gases CO$_2$, CH$_4$, and H$_2$O, which act to trap radiant heat. The atmosphere within this region naturally becomes unstable due to the positive lapse rate, producing strong convection due to heating near the Earth's surface. The boundary between the upper troposphere and the stratosphere is called the tropopause and marks a turning point in the temperature profile from positive to negative lapse rate. The troposphere is therefore a rich source of convective (and other types) of vertically propagating perturbations, which can significantly modify the structure and dynamics of the other atmospheric regions. Atmospheric
waves, generated from tropospheric sources, propagate upwards into the rarified upper atmosphere, where wave amplitudes increase with height as the background density decreases.

The stratosphere is the second region of the atmosphere and extends from the tropopause to \( \sim 50 \) km. Due to variations in solar heating, the base of the stratosphere varies with latitude and the seasons. Throughout the stratospheric region, the atmospheric temperature increases with altitude up to about 270 K at around 50 km. This temperature reversal (inversion) is caused by the ozone layer \((O_3)\), which resides in the stratosphere (peak at \( \sim 30 \) km) and absorbs incident solar ultraviolet radiation, which is the primary source of heating in this region. Because of the negative lapse rate, the atmosphere is stably stratified and there is little convection or mixing within the stratosphere. This means that the air has greater static stability or greater buoyancy in the stratosphere. Various types of waves, such as atmospheric gravity waves and tides carry energy from the troposphere upward into the stratosphere and convey energy from this region up into the mesosphere. The waves and tides influence the flows of air and can cause regional heating in the stratosphere. The boundary between the stratosphere and the mesosphere is called the stratopause where the temperature stops increasing and reverts back to a positive lapse rate.

The mesosphere is the third region of the atmosphere and extends from about 50 to 85 km, where the temperature again decreases with altitude. The gravity waves, planetary waves and tides carry energy and transfer momentum from the troposphere and the stratosphere upward into the mesosphere. The dynamical processes of waves and their upward transport of momentum influence the mesospheric temperature. A temperature inversion layer is formed in the mesosphere where the temperature lapse rate reverses its sign from negative to positive. Possible causes responsible for
production of mesospheric inversion layers are atmospheric gravity wave and tidal activity and chemical heating. The region of inversion layer of enhanced temperature can be caused through the breaking of gravity waves associated with the regions of convective or shear instabilities [Meriwether and Gardner, 2000]. Gravity wave-tidal interaction is also believed to be the cause for temperature inversions in the mesosphere [Whiteway and Carswell, 1995].

The upper limit of the mesosphere is called the mesopause where the air temperature drops to its lowest values, reaching around $\sim$130–150 K at $\sim$85–90 km in the polar regions. A unique thermal feature of this region is that the summer mesopause is much cooler than its winter counterpart, despite the fact that it is continuously bathed in sunlight. This is due to the existence of a high-altitude summer-to-winter meridional circulation, which flows from the summer to the winter polar region. Rising air causes expansion and cooling, resulting in decreased temperatures at the summer mesopause, while strong downwelling air results in compression and warming at the winter mesopause. Atmospheric gravity waves (discussed below), are one of the most important dynamical processes in the atmosphere, and they play a key role influencing the wind and thermal structure of the mesopause region. In particular, the summer-to-winter circulation is thought to be due primarily to gravity wave dissipation in this region, which deposits momentum against the mean east-west flow, resulting in a residual north-south (meridional) circulation.

The thermosphere is the uppermost neutral region of the Earth’s atmosphere and extends from the mesopause ($\sim$85–100 km) up to $\sim$550 km height. Atmospheric temperature increases rapidly in the lower thermosphere (up to $\sim$700 K by $\sim$200–300 km altitude), and then remains fairly steady with increasing altitude. Temperatures in the upper thermosphere can range from $\sim$700 K to $\sim$1400 K or more, depending on the solar activity. The strong temperature inversion is caused by the absorption
of solar ultraviolet (UV) radiation by atomic oxygen and nitrogen, which is a major heating mechanism in the thermosphere. Winds and the overall circulation in the thermosphere are driven largely by the atmospheric tides and gravity waves. The neutral thermosphere overlaps in altitude with the ionosphere which extends from \( \sim 85 \text{ km} \) to well over \( \sim 550 \text{ km} \). Energetic ultraviolet and X-ray photons break electrons away from gas particles in the thermosphere, creating electrically-charged ions of atoms and molecules (primarily \( \text{N}_2 \) and \( \text{O}_2 \)). Moving ions, dragged along by collisions with the electrically neutral gases, produce strong electrical currents. Charged particles (electrons and protons) travel along the Earth's magnetic field lines collide and excite the constituent atoms and molecules, causing well-known optical auroral phenomena, which primarily occur at high-latitude thermospheric heights.

### 1.2 Atmospheric Airglow Emissions

It is known that there are several airglow emission layers in the upper mesosphere and lower thermosphere \([\text{Chamberlain}, 1961]\). The airglow is a very weak emission by chemical interactions of air molecules and atoms. Airglow is due to emissions from excited states formed by various processes in the upper atmosphere, such as the recombination of ions, which are photoionized by the sun during the day, luminescence caused by cosmic rays striking the upper atmosphere, and chemiluminescence caused mainly by oxygen and nitrogen reacting with hydroxyl ions. Some air molecules and atoms become excited by ultraviolet sunlight during the day, and de-excited to a lower-energy state at night, emitting light. It was first noticed by \textit{Angstrom} [1868]. Figure 1.2 shows a space shuttle picture of an airglow layer and the aurora above the limb of the Earth. Dayglow is the brightest airglow caused mainly by fluorescence processes as molecules and atoms are photodissociated and photoionized. Twilightglow is the most readily observable airglow from the ground
Figure 1.2. A picture taken from the space shuttle showing the visible airglow layer above the limb of the Earth. The aurora is red at the top [Courtesy of University of Alaska Fairbanks].

when only the upper atmosphere is illuminated by sunlight. Nightglow is the nocturnal phenomenon when the entire atmosphere is in darkness. The sinuous band of airglow layer seen in this picture is prominent, and is at an altitude of $\sim 100$ km. The aurora is red at the top.

Molecules giving rise to airglow emissions include sodium, atomic and molecular oxygen and nitrogen. Figure 1.3 shows a visible and near infrared spectrum of the nightglow emissions as recorded by a sensitive spectrometer [Broadfoot and Kendall, 1968]. The airglow emission-line and emission-band spectrum originate from energy level transitions in excited atoms and molecules at various heights in the Earth’s atmosphere. The airglow, together with aurora in the ionosphere, is emitted from the region at about 80 to 300 km in the atmosphere. It is caused by the absorption of incoming energetic solar ultraviolet and X-ray radiation, leading to atomic oxygen. Diffuse radiation, known as nightglow, extending from the near infrared to the far ultraviolet, is continuously emitted by the Earth’s atmosphere.
There are several prominent nightglow emissions which are all due to the photochemical processes. The nightglow consists of several different line and band emissions occurring at visible and near infrared (NIR) wavelengths in the MLT region. Numerous sounding rocket experiments have determined the OH emission layer, Na (589.2 nm) emission layer, O$_2$ emission layer, and OI (green line) emission layer at the mesopause heights. Figure 1.4 shows the normalized volume emission rate as a function of altitude for OI (green line), O$_2$, Na (D line) and OH nightglow emissions in the MLT region.

1.2.1 Green line OI emission

The brightest airglow line emission in the visible spectrum is the OI green line at 557.7 nm, from excited oxygen atoms in a layer peak at 96 km altitude. McLennan and Shrum [1925] first identified the green line emission as due to the
Figure 1.4. Normalized volume emission rates illustrating the different peak altitudes for the OI (555.7 nm), O₂ (0, 1), Na (589.2 nm) and NIR OH airglow emissions in the MLT region.

forbidden electronic transition (\(^1S_0 \rightarrow ^1D_2\)) of oxygen atoms. The green line emission originates from a single electronic transition and results in the line spectra. Chapman [1931] first proposed the three body atomic oxygen mechanism to explain the origin of this nightglow emission:

\[
O + O + O \rightarrow O_2^* + O^*. \tag{1.1}
\]

An alternative mechanism to explain the green line emission is given as [Barth, 1964]:

\[
O + O + M \rightarrow O_2^* + M, \tag{1.2}
\]
\[
O_2^* + O \rightarrow O_2^* + O(\,^1S\), \tag{1.3}
\]
\[
O(\,^1S) \rightarrow O(\,^1D) + h\nu (557.7 \text{ nm}), \tag{1.4}
\]
where M is a third body (major gas molecules, N\textsubscript{2} or O\textsubscript{2}), acting as a catalyst for two oxygen atoms to make excited O\textsubscript{2}.

1.2.2 NaD line emission

Another airglow emission, the sodium doublet line, is a yellow light from sodium atoms in a layer at 92 km. The yellow sodium emission is prominent at twilight, and due to photochemical processes in the night sky. The NaD radiation arises from the atomic transition Na (3\textsuperscript{2}P\textsubscript{3/2,1/2}-3\textsuperscript{2}S\textsubscript{1/2}), which consists of two closely spaced lines, D\textsubscript{1} and D\textsubscript{2}, at 589 and 589.5 nm, respectively. Chapman postulated the following sequence of nightglow emission reactions \cite{Chapman, 1939}:

\begin{align*}
\text{Na} + \text{O}_3 & \rightarrow \text{NaO} + \text{O}_2, \\
\text{NaO} + \text{O} & \rightarrow \text{Na} (^2\text{P}, ^2\text{S}) + \text{O}_2, \\
\text{Na}(^2\text{P}) & \rightarrow \text{Na}(^2\text{S}) + h\nu (589 \text{ nm}).
\end{align*}

In this process, the formation of NaO reacts with O to form Na in its excited \textsuperscript{2}P state, which then radiates to a ground \textsuperscript{2}S state, producing Na doublet emission.

1.2.3 O\textsubscript{2} emission

There are weak emissions from excited molecular oxygen. The peak of the O\textsubscript{2} atmospheric emission at 866–868 nm is \sim 94 km, and the emission height varies from 92 to 97 km. The mechanisms of excitation are three-body associations into the excited state with a third body M:

\begin{align*}
\text{O} + \text{O} + \text{M} & \rightarrow \text{O}_2^* + \text{M}, \\
\text{O}_2^* + \text{O} & \rightarrow \text{O}_2 ^1\Sigma + \text{O}, \\
\text{O}_2 ^1\Sigma & \rightarrow \text{O}_2 ^3\Sigma + h\nu.
\end{align*}

\textit{Krassovsky} [1958] considered the formation of O\textsubscript{2} molecules in excited vibrational levels in the ground state through the processes:

\begin{align*}
\text{O}_2 + \text{O} + \text{M} & \rightarrow \text{O}_3^* + \text{M},
\end{align*}
\[ O_3 + O \rightarrow O_2^* + O^*, \quad (1.12) \]
\[ O + O + M \rightarrow O_2^* + M. \quad (1.13) \]

### 1.2.4 OH emission

The red and infrared emissions from OH bands are the only example of emission from a vibrationally excited ground state. The peak of the OH (6, 2) Meinel emissions at 840–846 nm is around 87 km, and the OH emission is height-dependent, which varies from 85 to 90 km. The OH airglow spectrum is characterized by a number of isolated emission bands, each of which contains several discrete line emissions. Each band corresponds to a transition between two vibrational levels, with the emissions within each band being associated with rotational transitions [Herzberg, 1950]. The spectra determined by Meinel [1950] are due to rotational-vibrational transitions of the OH molecule in its ground state. The mechanism for OH excitation in the mesosphere was first proposed by Bates and Nicolet [1950] and further developed by Bates and Moiseiwitsch [1956]:

\[ O + O_2 + M \rightarrow O_3 + M, \quad (1.14) \]
\[ H + O_3 \rightarrow OH^* + O_2, \quad (1.15) \]
\[ OH^* \rightarrow OH + h\nu \ (3.34 \text{ eV}). \quad (1.16) \]

The energy released by reaction (3.34 eV) is sufficient to excite the ground-state OH to the ninth vibrational level. Krassovsky [1971] suggested an alternative mechanism involving vibrationally excited \( O_2 \):

\[ H + O_2^* \rightarrow OH^* + O. \quad (1.17) \]

There have been several measurements of wave structures imaged in the visible and NIR nightglow emissions from ground-based optical arrangements. Figure 1.5 shows examples of four airglow images of short-period gravity waves observed in OH, Na, O\(_2\) and OI emissions on June 5, 2002 from Bear Lake Observatory. The
most intense nightglow emission is the NIR OH emission, which is an extensively studied emission from ground level. The evident of wave structures imaged in the Na (D-line), O$_2$, and OI (green line) nightglow emissions is relatively less extensive.

1.3 Gravity Waves in the Atmosphere

In the Earth’s atmosphere, an air parcel has a tendency to move upwards or downwards in response to gravity and pressure-gradient forces. The atmospheric density gradient gives rise to a buoyancy force, which, together with gravity, provides restoring forces (potential energy) that act on parcels of vertically displaced air. Coherent and transverse oscillations resulting from the restoring force of buoyancy in the stable atmosphere are known as atmospheric gravity waves or buoyancy waves [e.g., Hines, 1960]. When a parcel of air is displaced, a small vertical distance upwards into a less dense region from its hydrostatic equilibrium position, gravity will act on the parcel to return it to its original position. The inertia (kinetic energy) of the returning parcel will cause it to overshoot its original position into a denser region where the buoyancy force will again act to move the parcel up. If the fluid stratification is stable, the parcel will start to oscillate at about the point of zero net force, at the frequency known as the Brunt-Väisälä frequency, $N$. The existence
of atmospheric gravity waves (GWs) was first proposed by Hines [1960] to explain numerous wave-like phenomena, such as those evident in noctilucent clouds and meteor trains, which have been observed in the atmosphere [Hines, 1974; Gossard and Hooke, 1975; Lighthill, 1978; Fritts and Alexander, 2003].

Gravity waves are the waves for which gravity and stable stratification, rather than rotation, are the dominant influences [Fritts and Werne, 2000]. Because of compression, gravity waves in the high-frequency limit are referred to as acoustic-gravity waves. In the low-frequency limit, this is termed an inertia-gravity wave. Atmospheric gravity waves are sometimes referred to as internal gravity waves because oscillations occur within the stratified atmosphere, analogous to oceanic internal waves. Ocean surface waves are external gravity waves, which propagate at the density discontinuity between water and air. In the atmosphere, internal gravity waves can travel vertically or obliquely to the density strata from the troposphere into the stratosphere and mesosphere [e.g., Sica and Russell, 1999; Taylor and Pendleton, 2003]. Internal waves produced within a layer bounded above and below by density or wind discontinuity surfaces can be reflected and propagate within a wave guide.

The mountain lee waves that arise from flow over mountains are one important example of a gravity wave. In the steady state lee waves are stationary with respect to terrain, but they are propagating relative to the mean air flow above the Earth’s surface. Figure 1.6 shows a visible satellite image of trapped lee waves extending downwind from the Hawaiian Islands. The trapped lee waves marked by evenly spaced cloud bands were observed on January 24, 2003. Mountain waves have been observed via satellite with infrared [Eckermann and Preusse, 1999; Preusse et al., 2002] and microwave [Wu and Jiang, 2002; Jiang et al., 2002, 2004] limb scanning instruments. These observations of temperature fluctuations provide strong indications that mountain waves induced by flow over the Andes Mountains may routinely
reach the stratosphere. Figure 1.7 shows the structure of a stratospheric mountain wave event observed from Atmospheric Infrared Sounder (AIRS) satellite radiance measurements and retrieved temperatures over the Antarctic Peninsula [Eckermann and Preusse, 1999; Alexander and Teitelbaum, 2007]. Smith et al. [2009] recently observed stationary mesospheric GWs over the Andes mountain range via airglow imaging, suggesting that mountain waves can indeed reach the MLT region under suitable wind and temperature conditions.

Most of the excitation sources for the generation of the gravity waves lie in the troposphere and lower stratosphere [Taylor and Hapgood, 1988]. Two of the most important sources are orography and the convective system [Lu et al., 1984; Nastrom and Fritts, 1992; Alexander, 1996; Holton and Alexander, 1999; Fritts and Alexander, 2003; Alexander et al., 2004]. Several reported observations of short-period gravity waves are thought to be generated by convective forcing [Taylor and
Figure 1.7. Maps showing the structure of stratospheric mountain wave event observed at 43 km from Atmospheric Infrared Sounder (AIRS) satellite radiance measurements and retrieved temperatures over the Antarctic Peninsula [Courtesy of Alexander and Teitelbaum, 2007].

Edwards, 1991; Taylor et al., 1995a]. Such waves propagate into the MLT region, where their signatures can be detected readily in nightglow emissions [Moreels and Herse, 1977; Taylor and Hill, 1991; Swenson and Mende, 1994; Smith et al., 2000; Medeiros et al., 2003; Nakamura et al., 2003]. Other sources of gravity waves in the lower atmosphere include interaction between waves such as atmospheric tides, collisions of pressure fronts, wind shears, and various meteorological disturbances such as thunderstorms, frontal systems and jet-stream activity [e.g., Lu et al., 1984; Fritts and Nastrom, 1992; Tsuda et al., 1994; Preusse et al., 2001]. Unstable wind shears are believed to be common sources of vertically propagating gravity waves [Fritts and Luo, 1992]. Near the jet stream maximum, wind shears are very high,
providing favorable conditions of shear instability and generation of gravity waves. At higher altitudes they can be generated by the breakdown of large-scale wave motions [Fritts and Nastrom, 1992; Tsuda et al., 1994].

1.4 Gravity Wave Effects on the Atmosphere

Gravity waves interact with the atmosphere in many different ways. Hines [1960] first suggested that gravity waves might produce important dynamical effects in the middle atmosphere. Gravity waves propagating from the lower atmosphere cause fluctuations in the background wind, density and temperature, and short-period gravity waves can have a marked effect on the mean flow and thermal structure of the MLT region through wave energy and momentum transports [e.g., Lindzen, 1981; Holton, 1982; Garcia and Solomon, 1985].

1.4.1 Waves dissipation and background wind

The decrease in atmospheric density as the waves propagate upward without wave breaking or dissipation leads to the increase in amplitudes of oscillations with altitude in order to conserve kinetic energy [e.g., Gossard and Hooke, 1975]. That is, the wave kinetic energy per unit volume \( \frac{1}{2} \rho v^2 \) remains constant in the absence of dissipation where \( \rho \) is atmospheric density and \( v \) is the wind perturbation velocity due to the wave. In the linear theory, the velocity amplitude of displacements of the wave grows as \( e^{\frac{z}{H}} \) where \( \rho = \rho_0 e^{-\frac{z}{H}} \), \( z \) is altitude and \( H \) is the density scale height of an isothermal atmosphere. Because of this growth, wave amplitudes can eventually become large enough to produce breaking. When wave amplitudes cease to grow or becomes limited due to convective or shear instabilities, the energy is not conserved and the result is referred to as saturation of the waves [e.g., Fritts, 1984]. Under linear saturation theory, wave amplitude can grow only to the point where the total flow becomes convectively unstable and growth is balanced by wave
dissipation. Saturation theory assumes that the dissipation results in a divergence of wave stress, and wave breaking, resulting in turbulence [e.g., *Lindzen*, 1981; *Holton*, 1982; *Fritts and Alexander*, 2003].

The main consideration in the observational studies of atmospheric gravity wave effects in the MLT region is to determine their propagation characteristics such as vertically propagating or ducted, directionality, momentum and energy fluxes, and local and seasonal variability. Several observations and analyses have shown the ducting conditions of gravity waves [e.g., *Chimonas and Hines*, 1986; *Fritts and Yuan*, 1989a; *Isler et al.*, 1997; *Walterscheid et al.*, 1999, 2001; *Snively and Pasko*, 2003; *Snively et al.*, 2007]. When waves become evanescent, they have no vertical propagation and vertical momentum flux. Ducted waves consist of both upward and downward propagating waves whose momentum fluxes cancel each other. The momentum flux of waves would have no effect on the background atmosphere if there were no wave dissipation or breaking.

It is known that the induced variations in atmospheric wind speed away from the background average are due to the passage of atmospheric gravity waves [*Hines*, 1974]. The background wind can be influenced by gravity wave activity through a number of dissipative mechanisms including wave breaking, nonlinear interactions between gravity waves, wave-mean flow interactions, and damping by radiative and turbulence processes [Yeh and Liu, 1981]. The momentum transfer into the mean flow takes place when the gravity waves are broken or damped in the MLT region [e.g., *Lindzen*, 1981; *Fritts*, 1982, 1984].

Figure 1.8 shows gravity wave band and instability billow structures imaged in noctilucent clouds (NLC) display over Turku, Finland [Gadsden and Parviainen, 1995]. It is believed that the observed NLC structures are commonly due to upward-propagating and breaking gravity waves. In particular, vertical transport of horizon-
Figure 1.8. Photograph of noctilucent clouds (NLC) taken at Turku, Finland on 21-22 July, 1989 displaying billows across the long bands, and the billows exhibiting a wave-like structure with undulations [Courtesy of Gadsden and Parviainen, 1995].

tal momentum by gravity waves can play a crucial role in the global-scale circulation throughout the middle atmospheric region [e.g., Lindzen, 1981; Holton, 1983; Garcia and Solomon, 1985; Hamilton, 1996; Fritts and Alexander, 2003]. Wave dissipation causes divergence of momentum flux, which leads to local heating, turbulent diffusion, and accelerations of the local mean flow. The convectively generated gravity waves with large vertical wavelengths can play a major role in the flow of momentum into the MLT region [e.g., Alexander et al., 1995, 2004; Holton and Alexander, 1999; Horinouchi et al., 2002; Beres et al., 2004]. Dissipating gravity waves in the mesosphere and lower thermosphere (MLT) region exert a eastward (westward) gravity wave drag in the summer (winter) hemisphere, which is balanced by the coriolis torque of the meridional circulation. The adiabatic cooling (heating) associated with the rising (sinking) motion in the summer (winter) polar region balances the differential radiative heating between the summer and winter polar regions in the
mesopause [e.g., Andrews et al., 1987]. This mechanism causes an extreme cold summer mesopause, creating the conditions that permit ice formation. As gravity waves reach the mesopause region and break up, the input of their momentum in the summer polar mesopause can have a profound effect on noctilucent clouds (NLC) formation. Hines [1960, 1968] first proposed that structures seen in NLCs were the manifestation of internal gravity waves. Later modeling studies by [Fritts et al., 1993] indicated that the smaller-scale NLC billows that often accompany larger band structures are likely to be manifestations of evolving instability structures that are caused by the wave-breaking process in the mesopause region.

1.4.2 Temperature and airglow intensity perturbations

As gravity waves propagate, they induce significant density and temperature perturbations of the atmosphere. These perturbations in turn affect the photochemistry of the reactions, producing the airglow emissions through changes in concentration of the relevant atmospheric species. At the MLT heights, gravity waves create substantial perturbations in the intensity and rotational temperatures of several airglow emissions [e.g., Taylor et al., 1991, 2001; Garcia et al., 1997; Takahashi et al., 1999]. The gravity waves then appear as alternating bands of enhanced and depleted intensity that travel slowly across the night sky. Observations of the intensity variations of the OH, O$_2$ and OI (557.7 nm) airglow emissions provide information of wave parameters of passing gravity waves since the density and temperature vary with the wave’s intrinsic properties.

Atmospheric gravity wave induced airglow fluctuations of the hydroxyl OH Meinel bands are routinely observed using CCD imagers, operating in the visible and near infrared wavelength region [Swenson et al., 1990; Taylor and Hill, 1991; Taylor et al., 1995b; Hecht et al., 1995]. Swenson and Gardner [1998] have developed
analytic models to describe gravity wave induced perturbations in the high OH Meinel band emission and atomic Na density, and then predicted the fluctuations in OH intensity and rotational temperature. These constituents have been used to study the wave-induced perturbations in upper atmospheric density, temperature, and winds [e.g., Bills et al., 1991; She et al., 1991; Hecht et al., 1993] including estimates of momentum flux [e.g., Swenson et al., 1999; Espy et al., 2004, 2006; Suzuki et al., 2007, 2009, references therein]. Gravity wave-induced intensity and rotational temperature perturbations in the OH Meinel (6, 2) and O$_2$ Atmospheric band emissions have been investigated by using Mesospheric Temperature Mapper [Taylor et al., 2001]. Seasonal airglow intensity variations have also been observed for several different airglow emissions [Takahashi et al., 1992; Mulligan et al., 1995; Melo et al., 1999].

Theoretical and modeling studies have emphasized the understanding of the relation between the airglow perturbation and gravity wave [e.g., Walterscheid et al., 1987, 1999; Hickey, 1988; Tarasick and Shepherd, 1992a,b; Zhang et al., 1993; Hickey et al., 1993, 1997; Hines, 1997]. The numerical models have shown that the intensity perturbations observed by a ground-based imager depend primarily on the amplitude and vertical wavelength of the wave. The gravity waves typically have vertical wavelengths comparable to or larger than the layer thickness, and can propagate through the region [e.g., Gardner and Taylor, 1998]. The vertical wavelength and propagation direction can be measured from the phase information of perturbation observed in multiple layers. Liu and Swenson [2003] developed a model to examine the relation between gravity wave and airglow perturbations, as well as the phase and amplitude in observed OH and O$_2$ airglow layers.
1.5 Overview

In this dissertation, a study of small-scale features such as gravity waves, ripples and instabilities in the mesopause region is performed using unique Maui-MALT datasets. It is certainly known that the activity in the mesopause region can have significant effects on the climatology of other atmospheric regions. So study of dynamical features in this region is important to build a better understanding of the Earth’s atmosphere as a whole. The purpose of this dissertation is to investigate the propagation characteristics of gravity waves, and the climatology of ripples and associated instabilities. Gravity wave observations and measurements including analysis of airglow image data are described in Chapter 2. Chapter 3 of this dissertation describes the theory and data analysis methods used for this study. In Chapter 4, a climatology study of short-lived ripples with their seasonal anisotropy simultaneously observed in OH and O\textsubscript{2} airglow emissions is reported. Measurements are obtained using Mesospheric Temperature Mapper (MTM) image data from Maui, HI during two-year periods (2003–2004). In addition, measurements of anisotropy in the propagation direction of ripples are compared with gravity wave propagation direction. A case study of simultaneous occurrences of Kelvin-Helmholtz instability structures in OH and O\textsubscript{2} airglow layers due to dual anti-aligned shears is presented in Chapter 5. The Na lidar wind and temperature measurements smoothed to 100 m-15 min. spatial and temporal resolutions are used to investigate the dynamic instability associated with dual shears, and results are compared to the numerical simulation of the linking of Kelvin-Helmholtz instability at adjacent shear layers [Laughman et al., 2009]. In Chapter 6, the ducting and evanescence characteristics of small-scale gravity waves as determined by the meteor radar measurements of hourly background winds are examined. In two particular cases, analysis and modeling of Doppler-ducted and evanescent gravity waves observed in the OH and O\textsubscript{2}
airglow emissions are carried out in detail [Simkhada et al., 2009]. In Chapter 7, the characteristics of mesospheric bore-like events recorded in OH and \( \text{O}_2 \) emissions during 2003–2004 periods are presented. This is followed by a case study of Doppler-ducted and thermal-ducted bore-like events using meteor radar and Na lidar wind and temperature measurements. Finally, Chapter 8 summarizes the novel results in this dissertation research and the direction of possible further research.
CHAPTER 2

GRAVITY WAVES OBSERVATIONS AND MEASUREMENTS

2.1 Airglow Imaging

Airglow studies are frequently performed using satellite and ground-based instruments such as imagers, photometers and spectrometers. Airglow imaging is an important technique for investigating the horizontal morphology and dynamics of gravity waves in the mesopause region through airglow emissions. Emissions occur typically at 85–105 km and can be observed with low-light, all-sky imaging systems. Several airglow emission layers such as near infrared (NIR) OH Meinel band (peak at \( \sim 87 \) km), the NIR O\(_2\) (0, 1) Atmospheric band (peak at \( \sim 94 \) km), the visible wavelength OI (557.7 nm) green line (peak at \( \sim 96 \) km) and the NaD (589.2 nm) line (peak at \( \sim 90 \) km) have been widely utilized for the airglow imaging studies in the MLT region [Taylor et al., 1987, 1995b,c,d,e; Swenson and Espy, 1995; Swenson et al., 1995; Gardner et al., 1996; Taylor, 1997; Medeiros et al., 2001, 2004].

Figure 2.1 shows examples of short-period gravity wave and ripple structures in the NIR OH airglow emission imaged by the CCD camera on March 30, 1990.

![Figure 2.1](image)

**Figure 2.1.** Two examples showing images of near infrared OH wave structure recorded using a CCD imager from Haleakala Crater, Maui, HI during the ALOHA-90 Campaign on March 30, 1990. Two distinct types of wave motions are termed (a) bands, and (b) ripples [Taylor and Hill, 1991].
Figure 2.2. Examples showing (a) all-sky OH image of gravity wave structure recorded in the NIR OH emission on October 9, 1993 from an aircraft flying over the Pacific Ocean [Courtesy of G. R. Swenson, 1995], and (b) the ground map of all-sky OI and O$_2$ images observed on October 22, 1993 [Taylor et al., 1995b] from Maui, Hawaii during the ALOHA-93 Campaign.
literature describing observations of various airglow emissions and their importance for investigating the properties of gravity waves. During the ALOHA-93 Campaign, Taylor et al. [1995b] observed many extensive band and transient ripple events imaged in the four emissions, OI (557.7 nm), OH Meinel band, Na (589.2 nm), and O$_2$ atmospheric band (0, 1).

2.2 Radar and Lidar

A number of radar and lidar techniques have been employed to study gravity waves in the MLT region, including medium-frequency (MF) radar [e.g., Vincent and Fritts, 1987; Dowdy et al., 2001, 2007], incoherent-scatter radar (ISR) [e.g., Mitchell et al., 2002], mesosphere-stratosphere-troposphere (MST) radar [e.g., Wang and Fritts, 1990], and lidar [e.g., Senft and Gardner, 1991; She et al., 1991]. Most of the gravity wave observations in the MLT region have been made using either MF or MST radars [Manson and Meek, 1993; Vincent, 1994; Gavrilov et al., 1995; Nakamura et al., 1996]. The majority of studies of gravity waves in the middle atmosphere have been made primarily using Rayleigh and Na lidars to measure perturbations of temperature and density [Mitchell et al., 1991; Wilson et al., 1991; Senft et al., 1993; Meriwether et al., 1994; Collins et al., 1999]. Gravity wave momentum fluxes in the MLT region have been estimated using the MF radar [Fritts and Vincent, 1987; Murphy and Vincent, 1993], the VHF radar [Reid et al., 1988; Fritts and Yuan, 1989b], the MU radar [Nakamura et al., 1993; Gavrilov et al., 2000], the MST radar [Hitchman et al., 1992], the ISR radar [Fritts et al., 2006], and the Na lidar [Gardner and Liu, 2007] measurements. The co-located radar and lidar are very useful to compare data obtained from different observational methods, allowing more detailed interpretations of dynamical features in the atmosphere [Gardner, 1991; Hines, 1993; Taylor et al., 1995f; Gardner and Taylor, 1998]. Furthermore,
these observations have demonstrated the importance of coincident wind and temperature data for studying the effects of gravity waves and generation of instabilities in the MLT region. Because gravity waves can be observed under different background conditions, meteor radar and Na lidar wind and temperature measurements can be utilized to characterize the intrinsic properties and nature of propagation of gravity waves in the MLT region. Figure 2.3 displays examples of height profiles of meridional and zonal wind components measured simultaneously by the meteor radar and Na lidar in the mesopause region over Maui, HI, showing no significant differences in mean wind direction [Franke et al., 2005].

The meteor radar observations provide hourly zonal and meridional winds in the MLT region, which can readily be used for studying the tides, planetary waves, and gravity waves of periods 2–6 hours in this region [Kumar et al., 2007]. However, these hourly winds are not sufficient for studying short period gravity waves having periods less than an hour because of the limitations on their spatial and temporal resolution. Recently Hocking [2005] has shown that it is possible to obtain information on the fluctuating velocities corresponding to gravity waves using meteor radars. Few studies have used meteor radar wind to measure gravity wave activity, including wave variances and momentum fluxes in the MLT region [e.g., Hocking, 2005; Antonita et al., 2008; Clemesha and Batista, 2008; Mitchell and Beldon, 2009; Beldon and Mitchell, 2009]. The comparisons of meteor radar, wind together with airglow observations suggest that the seasonal directionality of gravity waves depends strongly on the prevailing winds at the mesopause region [Nakamura et al., 1993; Taylor et al., 1993; Takahashi et al., 1999; Walterscheid et al., 1999]. Airglow image data combined with simultaneous meteor radar wind measurements have been used to study propagation and Doppler ducting characteristics of gravity wave motions [Taylor et al., 1995c; Isler et al., 1997; Snively et al., 2007; Simkhada et al.,
Figure 2.3. Plots showing meridional and zonal wind profiles observed in the mesopause region over Maui, HI during the July 2002 and October 2003 campaigns. Lidar and radar wind profiles are marked with circles and asterisks, and the horizontal distance between vertical lines corresponds to 150 ms\(^{-1}\) [Franke et al., 2005].

2009]. Similarly, the nature of small-scale ripples and their generation mechanisms associated with local dynamic instabilities can be analyzed using radar wind of high resolution measurements.

Ground-based lidar measurements have been used to study temperatures and winds, providing another tool to probe the dynamics in the mesopause region [Swenson and Espy, 1995; Swenson and Liu, 1998]. The perturbations in density, temperature, and winds observed at the mesopause heights are caused by the propagating gravity waves [She et al., 1991]. Lidar studies of gravity wave temperature or density perturbations can provide information on seasonal variations in gravity wave activity [Mitchell et al., 1991, 1996; Whiteway and Carswell, 1995]. The generation of convective and Kelvin-Helmholtz billows, as well as the static stability and wind shear,
have been analyzed using lidar measurements of ambient wind and temperature in the MLT region \cite{Li_etal2005a, Hecht_etal2005, Li_etal2005, Sherman_She2006}. The lidar observations suggest the links between instabilities and temperature inversion layers, and in turn the possibility of upward propagating gravity waves and tidal structure. The gravity wave-tidal interaction can be the cause of temperature inversions in the mesosphere \cite{Whiteway_Carswell1995, Meriwether_etal1998, Liu_Hagan1998, Liu_etal2000, Williams_etal2002}. Furthermore, the lidar temperature measurements of high spatial-temporal resolution have also been utilized to study the thermal ducting conditions of small-scale gravity waves or mesospheric bores \cite{Smith_etal2003, She_etal2004}.

### 2.3 Maui-MALT Campaign and Instrumentation

The Air Force Office of Scientific Research (AFOSR) and the National Science Foundation (NSF) jointly supported a Campaign to study the Mesosphere and Lower Thermosphere (MALT), referred to as Maui-MALT. Measurements were obtained using a cluster of ground-based remote sensing instruments operated at the US Air Force AEOS facility located at the summit of Haleakala Crater, Maui, Hawaii (20.70°N, 156.30°W). These instruments include the Utah State University OH/O$_2$ band mesospheric temperature mapper, the University of Illinois Na wind/temperature lidar and meteor radar, and several other passive imaging instruments designed for the measurements of the MLT nightglow emissions. The Maui-MALT Campaign provided unique datasets, which make it possible to study gravity waves and tides and their induced airglow intensity and temperature perturbations in the MLT region at low-latitude. The measurements in this dissertation study are made by the Maui-MALT mesospheric temperature mapper, Na wind/temperature lidar, and meteor radar from Maui, Hawaii during 2002–2005.
2.3.1 Mesospheric temperature mapper (MTM)

The Utah State University Mesospheric Temperature Mapper (MTM) is a high-performance, solid state imaging system capable of measuring wave-induced fluctuations in the intensity and rotational temperatures of two upper mesospheric near infrared nightglow emissions: the OH (6, 2) Meinel band (peak altitude \( \sim 87 \) km), and the O\(_2\) (0, 1) Atmospheric band emission (peak altitude \( \sim 94 \) km), both of which exhibit well-defined layers with half-widths of \( \sim 8\text{-}10 \) km [e.g., Baker and Stair, 1988]. A high quantum efficiency CCD array coupled to a wide-angle telecentric lens system (90° field of view) is used to make narrow band (\( \Delta \lambda \sim 1.2 \) nm) emission measurements using a set of interference filters centered on the OH P1(2) and P1(4) lines (at 840 and 846.5 nm) and two well-defined regions of the O\(_2\) (0, 1) Atmospheric band (at 866 and 868 nm). Each emission is observed sequentially for 60 seconds followed by a background sky measurement at 857 nm, resulting in a \( \sim 5.5 \) minutes cadence and a zenithal footprint of \( \sim 0.9 \times 0.9 \) km at 90 km altitude. The high sensitivity and low noise characteristics of this system provide an exceptional capability for precise nocturnal measurements of the OH and O\(_2\) emission intensities (< 0.5% in 1 min.) and derived rotational temperatures (< 1–2 K in 3 min.). Further details of the MTM system and its operation are given in Pendleton et al. [2000] and Taylor et al. [2001, 2005]. Figure 2.4 shows the USU Mesospheric Temperature Mapper CCD imager developed by M. J. Taylor in 2001 at Haleakala, Maui. Figure 2.5 shows a schematic optical diagram of the CCD imager.

Here, the dataset obtained from the Mesospheric Temperature Mapper (MTM) is utilized to investigate horizontal characteristics of small-scale gravity waves, mesospheric fronts and ripples simultaneously observed in both OH and O\(_2\) airglow emissions. Figure 2.6 shows one example of the mesospheric frontal event recorded in OH emission at 07:33 UT and O\(_2\) emission at 07:36 UT using the MTM for July 4, 2003.
In addition, the MTM measurement is used for the estimation of the temperature and intensity perturbations in OH and O$_2$ airglow emissions induced by the passage of gravity waves. The mesospheric front observed during 07:21–07:51 UT (OH) and 07:24–07:53 UT (O$_2$) exhibited temperature and intensity perturbations in airglow emissions. Figure 2.7 shows band intensity and temperature profiles measured by the MTM from Maui, HI on July 4, 2003. The peaks of temperature and intensity
Figure 2.6. Example of flat-fielded and unwarped images of mesospheric frontal event observed by the Mesospheric Temperature Mapper (MTM) from Maui, Hawaii on July 4, 2003 (a) in the OH emission at 07:33 UT, and (b) in the O$_2$ emission at 07:36 UT.

Figure 2.7. Plots show the profiles of OH and O$_2$ (a) band intensities, and (b) temperatures measured on July 4, 2003 using the Mesospheric Temperature Mapper (MTM) from Maui, Hawaii.

perturbations induced by the front occurred at $\sim$07:30–08:00 UT.

2.3.2 Meteor radar

In this dissertation, the coincident meteor radar measurements of horizontal hourly winds from 80–100 km are utilized to obtain intrinsic parameters of gravity waves in the mesopause region. Figure 2.8 shows vertical profiles of meteor radar
Figure 2.8. Plots show the vertical profiles of meteor radar (a) zonal (solid line) and meridional (dashed line) winds, and (b) zonal (solid line) and meridional (dashed line) wind shears measured from Maui, Hawaii at 06:00 UT on August 7, 2004.

horizontal winds and wind shears measured from Maui, Hawaii at 06:00 UT on August 7, 2004. The system of the University of Illinois (UIUC) meteor radar is a SKiYMET radar [Hocking, 2001] operating at 40.92 MHz. A single three-element Yagi antenna directed toward the zenith is used to illuminate meteor trails. Meteor trail reflections are coherently detected on five three-element Yagi antennas oriented along two orthogonal baselines, with one antenna in the center of the array common to both baselines. On each baseline the outer antennas are separated from the center antenna by 1.5 and 2.0 wavelengths, respectively. This configuration provides enough redundancy to unambiguously determine the azimuth and elevation of meteor echoes, and angular resolution for position determination. Wind vector composed of eastward and northward components are estimated from the trail positions and Doppler shifts using a weighted least squares fit, where the vertical wind is assumed to be negligible. The wind vector fit is based on echoes collected within 1 hour time bins. Height resolution of the estimated winds is determined by a triangular height weighting function with a 3 km half width and a base width of 6 km. The center of the
height weighting function is stepped in 1 km increments to provide wind estimates at 1 km height intervals throughout the 80–100 km range. The detailed procedures are described by Franke et al. [2005]. The standard format provides hourly winds data in the altitude range 80–100 km, sampled every 1 km.

2.3.3 Na wind/temperature lidar

In this dissertation, lidar wind and temperature datasets are used to study instability structures and propagating characteristics of waves in the thermal ducts in the mesopause region. Figure 2.9 shows vertical profiles of Na lidar horizontal winds and temperature measured from Maui, Hawaii at 08:15 UT on October 28, 2003. The University of Illinois (UIUC) Na wind/temperature lidar is coupled with the steerable large aperture (3.67 m) astronomical telescope. A 0.6-m diameter portion of the telescope primary mirror is used to project the laser beam, while the remainder is used for collecting backscattered light and for focusing it onto the detector. Backscattered photons were collected by the primary mirror, and returned through the same optical path to the lidar receiver. The lidar system measures the line-of-sight (LOS) Na density, Doppler temperature, and Doppler wind profiles in the 80–105 km height range. The lidar is pointed to the zenith (Z) and 30° off the zenith to the north (N), east (E), south (S) and west (W) in the ZNEZSW sequence. This six-position sequence is completed about every 12 minutes so the complete wind and temperature field can be derived every 6 minutes. The raw photon data were collected at a high vertical range resolution of 24 m with a 90-second integration time to derive temperature, radial wind, and Na density from the ratio of the photon counts at three different frequencies. The raw lidar data are binned to vertical resolution of 480 m in zenith profiles and 417 m in off-zenith profiles. Temperature is derived in every lidar beam position. Horizontal wind at
Figure 2.9. Plots show the vertical profiles of Na lidar (a) zonal (solid line) and meridional (dashed line) winds, and (b) temperature measured from Maui, Hawaii at 08:15 UT on October 28, 2003.

Each direction (NSEW) is derived only from off-zenith LOS wind profiles, while the zenith LOS wind gives the vertical wind. The off-zenith LOS horizontal wind profiles are derived from the following relationships [Li et al., 2005b]:

\[
V_E = u_E \sin \theta + w \cos \theta, \tag{2.1}
\]

\[
V_W = -u_W \sin \theta + w \cos \theta, \tag{2.2}
\]

\[
V_N = v_E \sin \theta + w \cos \theta, \tag{2.3}
\]

\[
V_S = -v_E \sin \theta + w \cos \theta, \tag{2.4}
\]

where \(V, u,\) and \(v\) are the LOS wind, the zonal wind, and meridional wind for the E, W, N, and S lidar beam positions. The zenith angle, \(\theta = 30^\circ\) and \(w\) is the vertical wind. The zonal and meridional winds are calculated as the LOS wind measurements divided by \(\sin \theta\) and ignore the vertical wind. Therefore, a complete ZNEZSW cycle includes six temperature profiles, two zonal wind profiles, and two meridional wind profiles, having temporal resolution of 2 minutes for temperature and 6 minutes for horizontal winds. The processed data were then spatially and temporally smoothed.
using a 2-D Hamming window with full widths of 1 km and 0.5 hour. The data are interpolated onto a uniform grid with 100 m vertical resolution and 15 min. temporal resolution. A more detailed description of the Maui-MALT Na lidar system is given by Franke et al. [2005].

2.4 Analysis of Airglow Image Data

An airglow imaging method was applied to analyze short-period (< 1 hour) gravity wave and ripple structures in the Mesospheric Temperature Mapper (MTM) images of the airglow emissions. The technique involves flat fielding, spatial calibration, and geographical projection to determine the horizontal wave parameters. To investigate wave patterns, the raw MTM imagery were processed separately for each OH and O$_2$ emission using well-developed flat-field functions in order to correct for non-uniformities in the image brightness due to the MTM optics and to variations in the CCD pixel sensitivity. This is a standard analysis technique derived from averaging together a large number of images at each filter wavelength. The data were then normalized using the appropriate zenith emission intensity to create a time series of images of uniform brightness across the camera field of view, thereby enhancing the detection and measurements of any wave-structures present. The flat-fielded OH and O$_2$ data were then calibrated using the background star-field [e.g., Hapgood and Taylor, 1982], geographically orientated and projected onto a uniformly sampled 180 km $\times$ 180 km rectangular grid [e.g., Garcia et al., 1997; Pautet and Moreels, 2002] assuming nominal emission altitudes of $\sim$87 km and $\sim$94 km respectively. This was the maximum useable grid size and encompassed almost all data within the camera’s 90° circular field of view.

The horizontal wavelength and propagation direction of each wave event was then computed using a three-dimensional Fast Fourier Transform (FFT) spectral
analysis technique to determine their unambiguous spectrum typically using a time series of 4 (or 8) images [Garcia et al., 1997; Coble et al., 1998]. The apparent phase speeds of waves were determined separately by measuring the horizontal displacement of selected wave crests as a function of time. Data from both the P₁(2) and P₁(4) time series (alternately sampled at ∼1 and ∼5 min. intervals) were used to obviate possible aliasing problems due to the overall 5.5 minutes cadence time of the measurement. A similar technique was used to quantify the O₂ wave motions. Finally, the onset time and duration of each event was determined using movie clips.

2.4.1 Band-type wave

Band events appear as a train of wave fronts in airglow images, and are generally assumed to be quasi-monochromatic gravity waves originating from the lower atmosphere and propagating up to the MLT region. These extensive, long-lasting band-type wave events have been attributed to freely propagating, evanescent and ducted short-period gravity waves [Chimonas and Hines, 1986; Taylor et al., 1995c; Isler et al., 1997; Walterscheid et al., 2001; Snively et al., 2007; Simkhada et al., 2009]. Figure 2.10 shows examples of band-type wave structures imaged in the OH and O₂ emissions on April 24, 2004 during the MAUI-MALT Campaign. As shown in Figure 2.10a, the images of a 90° circular field of view were flat-fielded in order to correct for non-uniformities in the image brightness. In Figure 2.10b, the flat-fielded OH and O₂ images were calibrated using the background star field in order to orient geographically [Hapgood and Taylor, 1982]. The calibrated flat-fielded images were then unwarped by projecting onto a uniformly sampled 180 km × 180 km rectangular grid [Garcia et al., 1997] as shown in Figure 2.10c.

The horizontal characteristics of short-period gravity waves and their temporal evolution, have been observed by using mesospheric airglow data observations
Figure 2.10. Sequential airglow images showing band-type wave structures simultaneously recorded in OH and O₂ emissions at 10:38 and 10:41 UT, respectively, on April 24, 2004 during the Maui-MALT Campaign. Each image observed from the Mesospheric Temperature Mapper (MTM) was (a) flat-fielded, (b) calibrated, and (c) unwarped onto a 180 km × 180 km geographical grid.

reveal that the short-period gravity waves exhibit typical horizontal wavelengths of a few to several tens of kilometers and horizontal phase velocities of \( \sim 20–100 \, \text{ms}^{-1} \), resulting in observed periods of typically \( \sim 10–20 \) minutes (< 1 hour). Most of the airglow image measurements reported a clear seasonal variation of the horizontal propagation direction of the waves [e.g., Taylor et al., 1997; Nakamura et al., 1999; Medeiros et al., 2004]. The anisotropy of the wave propagation direction observed mainly in summer and winter was due to wave filtering by the stratospheric and mesospheric winds [Taylor et al., 1993; Medeiros et al., 2003].

Figure 2.11 shows the first observed bore event in the mesospheric airglow emissions using the all-sky imager from Maui, Hawaii on October 10, 1993. This
spectacular gravity wave event was reported by Taylor et al. [1995d] in the Hawaiian airglow during the ALOHA-93 Campaign and was later termed a mesospheric bore [Dewan and Picard, 1998]. This disturbance appeared as a leading sharp front followed by several wave crests that progressed coherently across the sky. In similar events, dramatic changes in airglow intensity and temperature are often accompanied with the passage of the leading front. We have reported several number of bore-like events and their horizontal characteristics and occurrence of frequency at OH and O$_2$ emission altitudes using the MTM imaging system from Maui, Hawaii during the Maui-MALT Campaign. Bore-like events observed from the MTM exhibit average horizontal wavelength of 21 km and horizontal phase speed of 46 ms$^{-1}$, resulting in an average observed period of 8 min. Figure 2.12 shows examples of bore-like wave structures observed in OH and O$_2$ emissions on September 20, 2003 during the MAUI-MALT Campaign. This mesospheric bore event is characterized by sharp bright (OH) and dark (O$_2$) leading fronts followed by a growing number of trailing wave crests. There have been other reports of the mesospheric bores observed in air-
Figure 2.12. Series of airglow images showing bore-like structures simultaneously observed in OH and O\textsubscript{2} emissions at 10:06–10:12–10:17 UT and 10:09–10:15–10:20 UT, respectively, on September 20, 2003 during the Maui-MALT Campaign. Each flat-fielded image observed from the Mesospheric Temperature Mapper (MTM) was calibrated and unwarped onto a 180 km \times 180 km geographical grid.


2.4.2 Ripple-type wave

Ripples are small-scale structures, seen in airglow images, which are believed to be manifestations of instability features generated in situ \cite{Peterson1979, Peterson1983, Taylor1990, Taylor1991, Swenson1994}. Figure 2.13 shows examples of ripple structures imaged in OH and O\textsubscript{2} emissions on May 10, 2004 during the MAUI-MALT Campaign. Each image was (a) flat-fielded to correct the non-uniform intensity in the image, (b) calibrated to get a geographic orientation, and (c) unwarped onto a 180 km \times 180 km rectangular grid.
Figure 2.13. Sequential airglow images showing ripple structures simultaneously recorded in OH and O\textsubscript{2} emissions at 11:04 and 11:07 UT, respectively, on May 10, 2004 during the Maui-MALT Campaign. Each image observed from the Mesospheric Temperature Mapper (MTM) was (a) flat-fielded, (b) calibrated, and (c) unwarped onto a 180 km × 180 km geographical grid.

Ripples, having quite different characteristics, are easily distinguished from bands. Ripples are much smaller in spatial extent and exhibit considerably shorter lifetimes (<50 min.). Several measurements indicate that the short-lived ripples observed in the airglow images exhibit typical wavelengths ∼5–15 km and phase speeds of ∼20–50 ms\textsuperscript{−1}, resulting in observed periods of ∼5–10 minutes [Taylor et al., 1997; Nakamura et al., 1999; Medeiros et al., 2004].

A combination of observational and theoretical studies provides that the small-scale ripples are generated by localized dynamical instabilities [Taylor and Hapgood, 1990; Hecht et al., 2001b, 2005; Li et al., 2005a] or by a convective instability [Fritts et al., 1997; Hecht et al., 1997; Yamada et al., 2001] in the mesopause.
region. The dynamic instability is caused by wind shear, whereas the convective instability is caused by a superadiabatic temperature lapse rate. Gravity waves and tidal perturbations can significantly influence the background wind and temperature structures causing instabilities in the MLT region. Dynamic and convective instabilities are known to be associated with gravity wave breaking, suggesting that ripple events are signatures of gravity waves breaking at and near the airglow emission layer [e.g., Fritts et al., 1993; Hecht et al., 1997; Horinouchi, 2004; Li et al., 2005a].
CHAPTER 3
THEORY OF WAVES AND INSTABILITIES IN THE ATMOSPHERE

3.1 Buoyancy Oscillations and Atmospheric Stability

An air parcel, when displaced vertically from its position of equilibrium in a stable atmosphere, will tend to return to its original state by a series of movements, which usually take the form of oscillations. The buoyancy difference between the parcel and its environment will produce a restoring force, accelerating the parcel back toward its equilibrium position. The adiabatic oscillations of an air parcel about its equilibrium position in a stable atmosphere are referred to as buoyancy oscillations [Holton, 1992]. For an atmosphere in static equilibrium the net forces acting on an air parcel are balanced by the pressure-gradient force and the force of gravity. We consider the vertical motion of an air parcel at pressure $p$, temperature $T$, density $\rho$ and specific volume $v$ at altitude $z$, all equal to the values for the surroundings. Assuming the atmosphere to be in hydrostatic equilibrium [Houghton, 1986; Holton, 1992], we have

$$dp = -\rho g dz.$$  \hspace{1cm} (3.1)

From the first law of thermodynamics applied to unit mass for adiabatic motion of the parcel, the quantity of heat $dq = 0$, we get

$$c_p dT - v dp = dq = 0,$$  \hspace{1cm} (3.2)

where $c_p$ is specific heat of air at constant pressure. Substitution of $dp$ from equation (3.2) and $\frac{1}{v} = \rho$ gives

$$-\frac{dT}{dz} = \frac{g}{c_p} = \Gamma(z).$$  \hspace{1cm} (3.3)

Here $\Gamma$ denotes the rate of decrease of temperature with height and is called the adiabatic lapse rate. The dry adiabatic lapse rate, $\Gamma_d = -\frac{dT_d}{dz}$, is the rate of temperature decrease with height for a parcel of dry air rising under adiabatic conditions. For dry air, $\Gamma_d = 9.8 \, \text{K km}^{-1}$ and $c_p = 1005 \, \text{J kg}^{-1} \, \text{K}^{-1}$ [Andrews, 2000].
In order to derive the characteristic frequency of buoyancy oscillations, the air parcel is considered to be displaced vertically upward. Suppose an air parcel is displaced a distance $\delta z$ from equilibrium. The upward buoyancy force on the parcel at height $z + \delta z$ is given by

$$F_b = -g v (\rho_p - \rho_e), \quad (3.4)$$

where $v$ is the volume of the parcel there, and $\rho_p$ and $\rho_e$ are air parcel density and environmental (or ambient) air density. Figure 3.1 shows a schematic diagram of an air parcel displaced a height $\delta z$ from equilibrium position at height $z$ [Andrews, 2000]. By Newton’s second law, this force can be equated to the mass of the parcel times its acceleration, i.e.,

$$F_b = \rho_p v \frac{d^2(\delta z)}{dt^2}. \quad (3.5)$$

Since change in pressure for air parcel and its surroundings is the same, we apply the ideal gas law, $p = \rho RT$, where $R$ is the gas constant per mole, to introduce temperature of the air parcel ($T_p$) and temperature of the ambient air mass ($T_e$). Equating (3.4) and (3.5), we get an equation of the form

$$\frac{d^2(\delta z)}{dt^2} = -g \frac{(\rho_p - \rho_e)}{\rho_p} = -g \frac{(T_e - T_p)}{T_e}. \quad (3.6)$$

Let $T_a (= T)$ be the temperature of environment at initial position, $z$. Consider the parcel temperature to be initially equal to the temperature of the environment at $z$. At the displaced position ($z + \delta z$), the parcel temperature has increased to $T_p$. According to the adiabatic lapse rate, the parcel temperature is

$$T_p = T_a + \left( \frac{dT_a}{dz} \right) \delta z = T_a - \Gamma_a \delta z. \quad (3.7)$$

On the other hand, the environmental temperature at height ($z + \delta z$) is

$$T_e = T_a + \left( \frac{dT_a}{dz} \right) \delta z = T_a - \Gamma_a \delta z. \quad (3.8)$$
Here \( a \) and \( d \) are the environmental lapse rate (ELR) and the dry adiabatic lapse rate (DALR). If \( a = d \), there is a temperature difference between the displaced air parcel and its surroundings. Substituting \( T_p \) and \( T_e \) from equations (3.7) and (3.8) in equation (3.6), we get

\[
\frac{d^2 (z)}{dt^2} = \frac{g}{T_a} (d - a) \ z \quad (3.9)
\]

If the air parcel is displaced vertically and released, then its motion is described by the general solution of the differential equation (3.9) \([Nappo, 2002]\) as

\[
z(t) = Re \ Ae^{iNt} + Be^{-iNt} \quad (3.10)
\]

where the frequency of oscillation, \( N \), is given by

\[
N = \frac{g}{T_a} (d - a) \quad (3.11)
\]

This natural frequency of the oscillation, \( N \), of the displaced air parcel is referred to as the Brunt-Väisälä frequency or buoyancy frequency, with corresponding Brunt-
Väisälä period $\tau_B = \frac{2\pi}{N}$ [Hines, 1960]. Then, re-writing equation (3.9) gives

$$\frac{d^2(\delta z)}{dt^2} + N^2 \delta z = 0,$$

(3.12)

where static stability is given by

$$N^2 = \frac{g}{T_a} (\Gamma_d - \Gamma_a) = \frac{g}{T_a} \left(\frac{dT_a}{dz} - \frac{dT_d}{dz}\right).$$

(3.13)

Figure 3.2 illustrates a schematic diagram of vertical stability in terms of temperature, showing the dry adiabatic lapse rate (DALR), and the stable and unstable environmental lapse rates (ELRs). This graph shows the motion of a parcel of air moving adiabatically under the influence of a buoyancy wave. When the temperature of a dry air parcel, $T_d >$ temperature of the environment, $T_a$, the parcel becomes warmer and less-dense than the environment, the displacement will continue to grow and the parcel will be positively buoyant (buoyant force in direction of displacement). For unbounded growth of the displacement, $\Gamma_d < \Gamma_a$, is called a superadiabatic lapse rate, and $N^2 < 0$ ($N$ is imaginary). When this condition is present, the atmosphere is to be statically unstable. When temperature of a dry air parcel, $T_d <$ temperature of the environment, $T_a$, the parcel becomes cooler and denser than the environment and the parcel will sink back down to its initial level and be negatively buoyant (buoyant force is opposite the displacement). For the parcel oscillation, $\Gamma_d > \Gamma_a$ and $N^2 > 0$ ($N$ is real). When this condition is present, the atmosphere is to be statically stable or stably stratified. Gravity waves are able to propagate vertically only in stable atmospheric conditions. The maximum frequency of vertically propagating gravity waves is equal to the Brunt-Väisälä frequency. The buoyancy frequency (period) is larger (smaller) in regions of higher atmospheric stability, while it decreases (increases) as the atmosphere becomes more unstable when $\Gamma_a$ approaches $\Gamma_d$. In the mesosphere, typically the Brunt-Väisälä frequency ($N$) = 0.02 s$^{-1}$, leading to a buoyancy period of about 5 minutes [Nappo, 2002].
Figure 3.2. Schematic diagram of vertical stability in terms of temperature, showing the dry adiabatic lapse rate (DALR), and the stable and unstable environmental lapse rates (ELRs) [e.g., Andrews, 2000].

Purely vertical displacements of air parcels due to waves occur when the wave train phase lines are only moving horizontally. However, gravity waves almost always propagate at an angle to the vertical within frequency \( N \), and so the uid parcels will be displaced at an angle to the vertical. Consider the parcel of air displaced upward from its equilibrium position along a sloping surface inclined at an angle \( \phi \) to the horizontal. Figure 3.3 shows a diagram of an air parcel displaced upward along a sloping surface under the in uence of gravity and buoyancy forces [Andrews, 2000]. If the air parcel moves a vertical distance \( z \) upwards from its equilibrium position and hence a distance \( s = \frac{z}{\sin \phi} \) along the slope, it will then experience a downward buoyancy force (per unit mass) \( F_b = N^2 z \). The component of buoyancy force (per unit mass) up the sloping surface is \( N^2 \sin \phi \ z \). Therefore, acceleration of the parcel up the slope is given by

\[
\frac{d^2 s}{dt^2} = \frac{d^2}{dt^2} \frac{z}{\sin \phi}
\] (3.14)
and Newton’s second law of motion gives

\[
\frac{d^2 z}{dt^2} \sin \phi + N^2 \sin \phi \cdot z = \frac{d^2 (z)}{dt^2} + 2 N^2 \sin^2 \phi \cdot z = \frac{d^2 (z)}{dt^2} + 2 z = 0 \quad (3.15)
\]

This implies an oscillation whose angular frequency \( \omega \) is given by

\[
\omega^2 = N^2 \sin^2 \phi = N^2 \frac{k^2}{k^2 + m^2} \quad (3.16)
\]

This is the dispersion relation for a hydrostatic gravity wave in the Boussinesq approximation, where \( k \) and \( m \) are horizontal and vertical wave numbers. The basis of this approximation is that there are flows in which the variation in density is neglected except where it is coupled with gravity to produce buoyancy force. The parcel executes an oscillation at the frequency \( \omega \) that depends only on the static stability measured by the buoyancy frequency \((N)\). Therefore, the range of gravity wave frequencies extends from 0 to \( N \) depending on the angle of propagation relative to the horizontal plane, \( \phi \) [Holton, 1992; Nappo, 2002].
### 3.2 Richardson Number and Instability Formation

The existence of a vertical shear in the horizontal wind disturbs atmospheric flow. Shear or dynamic instability is induced by large vertical shears of horizontal wind with low static stability. The instabilities of the atmosphere can be characterized by considering the square of the Brunt-Väisälä frequency, $N^2$, and the gradient Richardson number, $Ri$ [Richardson, 1920; Gossard and Hooke, 1975]. This dimensionless number, $Ri$, is defined as the ratio of the stabilizing effect of buoyant force to the destabilizing effect of wind shear or ratio of buoyant production to shear production, i.e.,

$$ Ri = \frac{N^2}{\left( \frac{du}{dz} \right)^2 + \left( \frac{dv}{dz} \right)^2}, \quad (3.17) $$

where $u$ and $v$ are zonal and meridional winds, and the $\frac{dU}{dz} = \sqrt{\left( \frac{du}{dz} \right)^2 + \left( \frac{dv}{dz} \right)^2}$ is the total vertical shear of the horizontal winds. In terms of potential temperature, $\theta$, the static stability is $N^2 = \frac{g}{\theta} \frac{d\theta}{dz}$ and the equation (3.17) can be written as

$$ Ri = \frac{\frac{g}{\theta} \frac{d\theta}{dz}}{\left( \frac{dU}{dz} \right)^2}. \quad (3.18) $$

The Richardson number represents the ratio of the work required to interchange vertically adjacent air parcels against the acceleration of gravity, to the kinetic energy available to do the work. Consider the work done in the adiabatic exchange of two neighboring parcels separated by $\Delta z$ in a medium with a potential temperature gradient, $\frac{\Delta \theta}{\Delta z}$ [Gossard and Hooke, 1975]. The work done is given by

$$ \Delta W = \rho g \left( \frac{\Delta \theta}{\theta} \right) \Delta z, \quad (3.19) $$

where $\Delta \theta = \theta_{z+\Delta z} - \theta_z$. The kinetic energy available to do the work due to the velocity difference $\Delta U$ between the parcels is given by

$$ \Delta E = \frac{1}{2} \rho \left[ U^2 + (U + \Delta U)^2 - \frac{1}{2} (U + U + \Delta U)^2 \right] = \frac{1}{4} \rho (\Delta U)^2, \quad (3.20) $$
\[ \frac{\Delta W}{\Delta E} = \frac{\rho g (\frac{\Delta \theta}{\theta}) \Delta z}{\frac{1}{4} \rho (\Delta U)^2} = 4 g \frac{\Delta \theta}{\theta} (\Delta U)^2, \]  

(3.21)

\[ \frac{1}{4} \frac{\Delta W}{\Delta E} = Ri. \]  

(3.22)

So, for stability, \( 4 \rho g \Delta z \frac{\Delta \theta}{\theta} > (\frac{dU}{dz})^2 \) or \( Ri > \frac{1}{4} \frac{g \Delta \theta}{(\Delta U)^2} \) [Gossard and Hooke, 1975].

There are two instability conditions: (1) when \( Ri \) is \( 0 < Ri < \frac{1}{4} \), dynamic instability occurs due to either the presence of large wind shear or a moderate wind shear associated with a large negative value of vertical temperature gradient but less than 9.5 K km\(^{-1}\) at the mesopause height to overcome the stabilizing buoyant force, and (2) when \( Ri \leq 0 \) with negative \( N^2 \), convective instability occurs due to the presence of superadiabatic temperature gradient. Hines [1971] found that even when \( Ri > \frac{1}{4} \), the shear flows must be inherently less stable. Although \( Ri \leq \frac{1}{4} \) is a necessary condition to produce dynamic instability in a fluid undergoing shear, it is far from a sufficient condition [Gossard and Hooke, 1975].

The Kelvin-Helmholtz (K-H) instability can occur when velocity shear or sufficient velocity difference is present at the interface between two adjacent flow regions. Figure 3.4 shows a schematic diagram of time evolution of the K-H instability at the interface of two fluid layers of different velocities and densities [Graf and Mortimer, 1979]. Consider two layers of the fluid flow at different velocities (\( U_1 \) and \( U_2 \)) and densities (\( \rho_1 \) and \( \rho_2 \)), where a lighter (warmer) fluid is moving to the right and a heavier (colder) fluid is moving to the left, which tend to create a wave of wavelength \( \lambda \). Internal gravity waves can propagate along the interface separating these two stratified fluid layers. When a vertical shear exists across the interface, the internal wave may grow in time and break in the vicinity of the interface, leading to K-H instability [Chapman and Browning, 1997]. The K-H instability results from enhanced velocity shears and/or a local minimum of the static stability either in
the mean flow or associated with low-frequency wave motions. But, the convective instability occurs in a region of negative static stability and minimum shear through the action of gravity wave motions of large amplitudes [Fritts and Rastogi, 1985].

3.3 Linear Gravity Waves Theory

The buoyancy force acting on an air parcel and its subsequent ability to generate an oscillatory motion is strongly dependent on the stability of the surrounding atmosphere. The atmosphere is taken as a complex medium of varying background temperature, density, and neutral wind field through which the waves propagate. Atmospheric gravity waves can be described with a simple linear theory that treats them as small departures from a stably stratified background state varying only in the vertical. In simplest form, the equation of motion can be solved analytically, starting from the continuity, momentum and energy equations.
3.3.1 Linear gravity wave equations

The equations of motion for atmospheric gravity waves (GWs) are derived here from linear theory. The Taylor-Goldstein equation of linear gravity waves for the vertical perturbation velocity \( w_z \) can be written as

\[
\frac{d^2 w_z}{dz^2} + \left[ \frac{N^2}{(c - u_0)^2} + \frac{1}{(c - u_0)} \frac{d^2 u_0}{dz^2} - \frac{1}{H(c - u_0)} \frac{du_0}{dz} - k^2 - \frac{1}{4H^2} \right] w_z = 0. \tag{3.23}
\]

This is the second order linear, irrotational and inviscid Taylor-Goldstein equation after neglecting compressibility effects and assuming solutions periodic in time and the horizontal direction. The first term on the left in the brackets is called the buoyancy term, the second term is called the curvature term, and the third term is called the shear term. The fourth term is also called the nonhydrostatic term, and the last term has no specific name. Full details derivation of linear gravity wave equations are given in Appendix A. If the coefficient in brackets is constant, then equation (3.23) yields simple plane-wave solutions of the form:

\[
w'(x, z, t) = w_z e^{\frac{i}{\hat{m}} e^{i(k x - \omega t)}}. \tag{3.24}
\]

If there is no background wind shear, then the WKB solution to the Taylor-Goldstein equation (3.23) for vertical wave number, \( m \), is given by [Nappo, 2002] as

\[
m^2 = \frac{N^2}{(c - u_0)^2} + \frac{1}{(c - u_0)} \frac{d^2 u_0}{dz^2} - \frac{1}{H(c - u_0)} \frac{du_0}{dz} - k^2 - \frac{1}{4H^2}, \tag{3.25}
\]

where \( N \) is the Brunt-Väisälä frequency, \( c \) is observed phase speed, \( u_0 \) is the background wind in the wave propagation direction, and \( H \) is the scale height.

Figure 3.5 illustrates the wave fronts and wave vectors for a two-dimensional wave showing the relationship between the horizontal and vertical wave parameters. Consider a two-dimensional plane-wave solution of the form:

\[
w'(x, z, t) = Re \{ \hat{w} e^{i(k x + m z - \omega t)} \}, \tag{3.26}
\]
Figure 3.5. The wave fronts and wave vectors for a two-dimensional wave. Adapted from Taylor [1985].

where the horizontal and vertical wave numbers are \( k = \frac{2}{z} \) and \( m = \frac{2}{z} \), and the phase angle is \( \tau = kx + mz \). The lines of constant \( \tau \) determine the spatial orientation of the wave fronts. Figure 3.6 illustrates a diagram of downward progression of phase and upward progression of wave energy, as well as the growing amplitude of the wave with altitude [Hargreaves, 1992]. Airglow wave analysis typically provides horizontal wavelengths and phase speeds in the propagation direction of the short-period (< 1 hr) gravity waves, which are then used to estimate the observed periods and the vertical wave parameters. The angle, \( \sigma \), made by the wave vector (\( K \)) with the horizontal is

\[
\sigma = \tan^{-1} \left( \frac{\rho_z}{\rho_x} \right) = \tan^{-1} \left( \frac{m}{k} \right) \tag{3.27}
\]
Figure 3.6. The downward progression of phase and upward progression of energy of gravity waves showing wave amplitude increases with height and particle motion. Adapted from Hines [1968].

The magnitude of the wave vector is
\[ |\vec{K}|^2 = k^2 + m^2, \]  
(3.28)

the wavelength (\(\lambda\)) is given by
\[ \frac{1}{\lambda^2} = \frac{1}{\lambda_x^2} + \frac{1}{\lambda_z^2} = \frac{(k^2 + m^2)}{(2\pi)^2}, \]  
(3.29)

and the phase speed in the direction of \(\theta\) is
\[ c = \frac{\omega}{2\pi \lambda} = \frac{\omega}{K}. \]  
(3.30)

The horizontal and vertical phase speeds are
\[ (c_x, c_z) = \left(\frac{\omega}{k}, \frac{\omega}{m}\right). \]  
(3.31)

For the zero background wind, the dispersion relation is
\[ \omega = \frac{Nk}{(k^2 + m^2)^{1/2}} = N \cos \theta. \]  

(3.32)

The horizontal and vertical group velocities in this case are given by

\[ (c_{gx}, c_{gz}) = \left( \frac{\partial \omega}{\partial k}, \frac{\partial \omega}{\partial m} \right) = \left( \frac{Nm^2}{(k^2 + m^2)^{3/2}}, \frac{-Nmk}{(k^2 + m^2)^{3/2}} \right), \]  

(3.33)

and using \( c_z \) and \( \omega \) from equations (3.31) and (3.32) in equation (3.33) for \( c_{gz} \), we obtain

\[ c_{gz} = -c_z \sin^2 \theta. \]  

(3.34)

Since \( \sin^2 \theta \) is always positive, \( c_z \) and \( c_{gz} \) must be in opposite sign [Nappo, 2002]. Therefore, a wave motion energy propagating upward must have a downward phase progression and be parallel to lines of constant phase.

### 3.3.2 Gravity wave ducting theory

An upward-propagating internal gravity wave may encounter a level where the background flow characteristics such as the Brunt-Väisälä frequency, \( N \), or background wind, \( u_0 \), change quickly with altitude. When such gradients give rise to wave reflection, gravity waves may become trapped in well-defined ducts, allowing propagation over large horizontal distances with little attenuation [e.g., Isler et al., 1997; Walterscheid et al., 2001; Hecht et al., 2001a].

Consider a simplified form of the Taylor-Goldstein equation for the vertical velocity field, \( w_z \), from equation (3.23) after neglecting curvature and shear terms:

\[ \frac{d^2 w_z}{dz^2} + \left[ \frac{N^2}{(c - u_0)^2} - k^2 - \frac{1}{4H^2} \right] w_z = 0. \]  

(3.35)

The dispersion relation with vertical wave number, \( m \), is given by

\[ m^2 = \frac{N^2}{(c - u_0)^2} - k^2 - \frac{1}{4H^2}. \]  

(3.36)

Figure 3.7 illustrates an idealized wind profile and corresponding altitude plot of \( m^2 \) where wind with maximum speed, \( u_{0max} \), defines a ducting region \((m^2 > 0)\)
bounded by evanescent regions \((m^2 < 0)\) on either side of the ducting region. In equation (3.36), both \(N\) and \(u_0\) are slowly varying functions of altitude. However, for very narrow velocity jets or \(u_0\) varying rapidly, the use of WKB solutions may not be justified [Isler et al., 1997]. If \(N^2\) or \(u_0\) or both together vary with altitude so as to make \(m^2 < 0\), a region of evanescence exists, causing reflection of wave energy. This leads to two kinds of wave ducting: thermal ducting and Doppler ducting, both of which can be present at any instant [e.g., Fritts and Yuan, 1989a]. A duct due to temperature gradient or variation in the Brunt-Väisälä frequency associated with superadiabatic temperature lapse rate is called a thermal duct [Francis, 1975]. A duct due to variation in the mean wind is called a Doppler duct. Doppler ducting was first considered by Chimonas and Hines [1986], and is typically created by a jet-like wind field when it exists in the same direction as wave propagation, resulting in a peak of vertical wave number squared, \(m^2\). In the evanescent regions, there exists \(\frac{N^2}{(c/u_0)^2} < k^2\) because of large negative background wind.

Figure 3.8 illustrates a freely propagating wave incident on an evanescent region \((m^2 < 0)\) and transmitted into a freely propagating region \((m^2 > 0)\), and the reflected wave at the upper boundary \((m^2 < 0)\), which propagates towards the lower
Figure 3.8. Wave pattern propagation in the ducting region bounded by two evanescent regions. Adapted from Nappo [2002].

reflecting level where it is again reflected upward. If the upward and downward waves constructively interfere, then the wave becomes trapped between the two reflecting levels. Some of the trapped wave energy can transmit through the evanescent regions.

3.3.3 The critical level

If the atmosphere exhibits a uniform wind field, the change in the frequency of wave for an observer moving with the basic flow (source of the wave) is known as the Doppler-shifted wave frequency or the intrinsic frequency, \( \nu_k = (c + u_0) k \), where \( \nu_k \) is the wave frequency observed in a fixed coordinate system, \( c \) is the apparent phase speed of the wave and \( u_0 \) is the component of the background wind in the direction of the wave propagation [e.g., Chimonas and Hines, 1986; Houghton, 1986]. The presence of horizontal background wind with velocity equal to a wave s
apparent horizontal phase velocity \((c - u_0)\) can provide a critical level for an upwardly propagating wave, with Doppler-shifted frequency and intrinsic phase speed approaching zero, where the vertical wave number squared, \(m^2\), in equation (3.25) approaches infinity.

The attenuation of gravity waves may be caused by the absorption of the waves at a critical level or by the breaking of the waves themselves at a different altitude. As the wave approaches the critical level from below, the wave-perturbation horizontal velocity grows as \((z - z_c)^{-1/2}\), where \(z_c\) is the height of the critical level \([\text{Gossard and Hooke, 1975}]\). This condition can eventually lead to gravity wave dissipation or breaking. When a gravity wave approaches a critical level, its vertical wavelength \(\lambda_z\) and vertical group velocity \(c_{gz}\) will vanish, at which point the waves no longer propagate vertically upward.

Figure 3.9 shows the propagation of a wave packet upward toward a critical level located at \(z_c\). As the vertical group velocity approaches zero, the wave fronts become horizontally oriented at \(z_c\). If the background wind is stable, the gravity waves encounter a critical level and will be absorbed into the mean flow at or just below this altitude. Gravity wave energy and momentum will be dissipated into the background flow \([\text{Hines and Reddy, 1967; Jones and Houghton, 1971}]\). This way of blocking waves is called critical level filtering which can cause significant anisotropy in the gravity waves’ propagation direction in the MLT region. Since the stratospheric jet is easterly (westerly) in the summer (winter) hemisphere, only eastward (westward) propagating gravity waves with phase speeds less than the magnitude of the jet velocity can reach the MLT region. There have been several measurements of mean wind variations \([\text{e.g., Manson et al., 1991; Nakamura et al., 1996}]\) and gravity wave filtering by the mean winds at lower altitudes \([\text{e.g., Taylor et al., 1993; Medeiros et al., 2003}]\). Gravity waves propagating through this region
Figure 3.9. The propagation of a wave packet upward toward a critical level located at $z = z_c$. Adapted from Gossard and Hooke [1975].

may also become unstable and break, depositing their energy into the mean flow, and leading to turbulent motions [Booker and Bretherton, 1967; Hines, 1968; Fritts et al., 2006]. Therefore, small-scale, short-period gravity waves can have a marked effect on the mean flow and thermal structure of the MLT region through wave energy and momentum transports and depositions [Holton, 1983; Garcia and Solomon, 1985; Hamilton, 1996].
CHAPTER 4
CLIMATOLOGY OF RIPPLE–TYPE INSTABILITY STRUCTURES IMAGED IN THE MESOPAUSE REGION OVER MAUI, HAWAII

4.1 Introduction

Imaging observations of airglow emissions in the mesosphere and lower thermosphere (MLT) region (altitude range \(\sim\)80–100 km), have become a standard technique for investigating the horizontal spatial and temporal properties of atmospheric gravity waves that are frequently detected in the vicinity of the mesopause. Over the past 30 years gravity waves with short periods (<1 hour) and small horizontal wavelengths (<100 km) have been observed from a number of sites at low, mid- and high latitudes, helping establish their global presence in the MLT region [Peterson and Kieffaber, 1973; Moreels and Herse, 1977; Peterson, 1979; Armstrong, 1982; Peterson and Adams, 1983; Takahashi et al., 1985; Taylor et al., 1987, 1993, 1995a,b,c,d,e,f, 1997; Tarasick and Hines, 1990; Taylor and Hapgood, 1990; Taylor and Hill, 1991; Taylor, 1997; Swenson and Mende, 1994; Swenson and Espy, 1995; Swenson et al., 1999; Gardner and Taylor, 1998; Nakamura et al., 1999, 2003; Hecht et al., 2000; Yamada et al., 2001; Ejiri et al., 2001, 2002; Smith et al., 2003; Shiokawa et al., 2003; Wrasse et al., 2003; Medeiros et al., 2003]. As a result of these studies, distinct spatial and temporal properties have emerged that suggest the existence of two types of short-period wave events, often termed bands and ripples. Bands are extensive, long-lasting (several hours) quasi-monochromatic wave patterns that exhibit horizontal wavelengths of a few to several tens of kilometers and horizontal phase speeds up to 100 ms\(^{-1}\) [e.g., Taylor et al., 1997]. These are the wave patterns that have been attributed to freely propagating (or ducted) short-period gravity waves [e.g., Taylor et al., 1987, 1995b,c, 1997; Isler et al., 1997; Walterscheid et al.,]
1999]. Ripples on the other hand are transient (<45 min. duration), small-scale (horizontal wavelengths typically <16 km) wave patterns [Peterson, 1979; Peterson and Adams, 1983; Taylor and Hapgood, 1990] that do not exhibit a clear relationship between their horizontal wavelengths and observed periods, as found for band events [Taylor et al., 1997], suggesting they have a different origin.

The apparent similarity in morphology and dynamics of ripple events with billow waves seen in polar mesospheric clouds [Fritts et al., 1993], and their small scale-sizes, strongly suggests that they are generated in situ within the MLT region by similar type source mechanisms [e.g., Fogle and Haurwitz, 1969; Taylor et al., 1984; Clairemidi et al., 1985]. More recent studies of ripples, including coordinated observations with lidar and radar wind and temperature measurements, have established that they are generated mainly by localized regions of wind shear creating dynamic instabilities, or by strong gradients in the background temperature field which can create convective-type instabilities [e.g., Taylor and Hapgood, 1990; Taylor and Hill, 1991; Swenson and Mende, 1994; Taylor et al., 1995f, 1997; Hecht et al., 1997, 2000, 2001b, 2005; Nakamura et al., 1999; Yamada et al., 2001; Hecht, 2004; Li et al., 2005].

Gravity wave bands are the most frequent wave patterns observed in the MLT airglow emissions. Ripple events are also quite common, but are less well documented, to date. Although there is some overlap in their horizontal wavelength scales, ripple events are easily distinguishable from propagating short-period gravity waves in three main ways: (a) the instability structures are short-lived (<45 min.), (b) they usually extend over much smaller geographical areas (typically <10⁴ km²) than bands which can exceed 0.5×10⁶ km², and (c) they all exhibit very short horizontal wavelengths (observed range ~5–16 km). These details have been known for some time [e.g., Taylor and Hill, 1991; Taylor et al., 1995b,f, 1997; Swenson and
It has also been noted that most (but not all) ripple events exhibit apparent horizontal speeds of < 60 ms$^{-1}$ [e.g., *Taylor et al.*, 1997; *Medeiros et al.*, 2004, 2007]. These characteristics result in a very narrow range of observed periods for ripple events, extending from the nominal Brunt-Väisälä period of $\sim$5 minutes to $\sim$10 minutes [e.g., *Taylor et al.*, 1997; *Nakamura et al.*, 1999; *Hecht et al.*, 2005; *Medeiros et al.*, 2004, 2007].

While there have been several in-depth investigations of individual ripple events identifying them with convective [*Hecht et al.*, 1997, 2000; *Li et al.*, 2005] and dynamical [*Hecht et al.*, 2001b, 2005; *Li et al.*, 2005a] type instabilities, as far as we are aware there are only four studies focusing on their seasonal properties and climatology [*Taylor et al.*, 1997; *Nakamura et al.*, 1999; *Medeiros et al.*, 2004, 2007]. The results of these studies are summarized in Table 4.1. Using all-sky measurements of three nightglow emissions OH, OI (557.7 nm) and Na (589.2 nm) *Taylor et al.* [1997] were the first to look into the variability of ripple characteristics. Measurements were made over a limited period (August-October, 1994) from equatorial Brazil (2.3° S, 45° W) as part of the NASA-INPE Guara Campaign. They compared 25 ripple events establishing mean values of 11.7 km and 44 ms$^{-1}$ for their horizontal wavelength and apparent horizontal speed, yielding a restricted range for their observed periods of $\sim$4–8 min. They also showed that ripples were often present in only one of the three emissions imaged consistent with their in-situ formation over a limited height range as later measured by *Ejiri et al.* [2002]. The ripples were also found to exhibit similar propagation directions ($\sim$NE) to an ensemble of gravity wave bands (31 events), but due to the limited observation period ($\sim$2 months) seasonal investigations were not possible.

*Nakamura et al.* [1999] reported on 18 months of ripple measurements (November 1996–May 1998) captured using a narrower field of view ($\sim$93°) OH imager lo-
Table 4.1. Data summary of ripple characteristics from airglow image data.

<table>
<thead>
<tr>
<th>Observation Site</th>
<th>Emission</th>
<th>Events</th>
<th>$\lambda$ (km)</th>
<th>$c$ (m/s)</th>
<th>$\tau$ (min.)</th>
<th>Duration</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Alcantara, Brazil (2.3° S, 45° W)</td>
<td>OH/Na/O$_2$</td>
<td>25</td>
<td>11.7 (6-17.7)</td>
<td>44 (23-86)</td>
<td>5.1 (2.7-7.5)</td>
<td>2 months</td>
<td>Taylor et al., 1997</td>
</tr>
<tr>
<td>Shigaraki, Japan (35° N, 136° E)</td>
<td>OH</td>
<td>175</td>
<td>&lt;17.5</td>
<td>10-60</td>
<td>5-10</td>
<td>18 months</td>
<td>Nakamura et al., 1999</td>
</tr>
<tr>
<td>Paulista, Brazil (23° S, 45° W)</td>
<td>OH/OI/O$_2$</td>
<td>102</td>
<td>13 (10-15)</td>
<td>27.2 (10-60)</td>
<td>8.4 (6-10)</td>
<td>12 months</td>
<td>Medeiros et al., 2004</td>
</tr>
<tr>
<td>Cariri, Brazil (7° S, 36° W)</td>
<td>OH/OI/O$_2$</td>
<td>525</td>
<td>14 (5-20)</td>
<td>29 (15-50)</td>
<td>9.6 (4-12)</td>
<td>51 months</td>
<td>Medeiros et al., 2007</td>
</tr>
<tr>
<td>Maui, Hawaii (20.7° N, 156.3° W)</td>
<td>OH/O$_2$</td>
<td>488</td>
<td>11 (6-18)</td>
<td>27.4 (20-50)</td>
<td>6.7 (5-10)</td>
<td>24 months</td>
<td>This study</td>
</tr>
</tbody>
</table>

cated at the MU radar facility at Shigariki, Japan (35° N, 136° E). This study defined ripples as wave events exhibiting horizontal wavelengths $< 17.5$ km. From a total of 175 such events they determined similar ranges for the ripple characteristics as previously reported by Taylor et al. [1997] (no mean values were given). Ripples were observed most frequently during the winter months but exhibited no preference in their observed direction of motion. Moreover, no clear change of propagation direction with season was found suggesting a ripple distribution close to omni-directional. However, the number of ripple measurements during the summer months was somewhat comprised by limited observing conditions.

The two Brazilian studies reported by Medeiros et al. [2004, 2007] are the most comprehensive to date. Using all-sky imager data from Cachoeira Paulista (23° S, 45° W) and later from Cariri (7° S, 36° W), they investigated ripple signatures in the OH, OI and O$_2$ nightglow emissions. The Cachoeira-Paulista study was performed over a one year period (October 1998–September 1999) and used the same all-sky camera employed for the Guara campaign study. In total, 102 ripple events were identified, yielding similar mean values for their horizontal wavelength (13 km) and observed period (8.4 min.), but a significantly lower average speed of (27.2 ms$^{-1}$),
compared with the limited observations of Taylor et al. [1997]. Again, no clear seasonal variation for the propagation direction of the ripples was found. In their most recent publication, Medeiros et al. [2007] reported on over 4 years of all-sky data (October 2000–December 2004) obtained from Cariri. All 51 months of data were averaged together to gain the best insight into seasonal ripple behavior (total 525 events). Their observed ripple characteristics were almost identical to those found in their earlier study from Cacheoira-Paulista (see Table 4.1), however, a strong anisotropy in the direction of motion of the ripples was clearly discerned with ripple propagation preferentially to the north or south. In particular, ripples were more common during the winter months and exhibited a strong preference for northward and southward motion. While during the summer months the ripple motions were predominantly northwards. Throughout their summed year east-west ripple motions were much more restricted.

In this study, we present new results on the frequency of occurrence, characteristics and seasonal variability of ripple events using two years of high-quality OH and O$_2$ nightglow measurements obtained by the Utah State University Mesospheric Temperature Mapper (MTM). The observations were made from the summit of Haleakala Crater, Maui (2970 m altitude) which afforded excellent night-sky observing conditions. This rich data set has enabled us to establish strong evidence for seasonal changes in their observed motions, as well as evidence for cyclic variations in their frequency of occurrence. These results provide new knowledge on the climatology of ripples present in the low-latitude mesopause region.

In the following Sections 4.2, we present example ripple data together with a description of our method of analysis used to determine their parameters. This is followed by a detailed summary of the seasonal results in Section 4.3 which focuses on several key aspects of the ripples. Discussion in Section 4.4 compares and contrasts
our results with previous reports and investigates the suggestion of Taylor et al. [1997] and Medeiros et al. [2004, 2007] that there may be a causal relation between gravity wave bands and ripples. These results are then summarized in Section 4.5.

4.2 Observations and Analysis

As part of the Maui-MALT program the MTM was operated autonomously for a five-year period (November 2001–December 2006) from the Air Force AMOS facility located at the summit of Haleakela Crater, Maui HI (20.7° N, 156.3° W, altitude 2970 m). MTM observations were made from dusk (~6:00 UT) to dawn (~16:00 UT), for solar depression angles > 12°, centered on the new moon period resulting in ~22 nights of observations per month. The excellent observing conditions resulted in over 1000 nights of high-quality measurements providing an important data set for investigating seasonal variability in mesospheric temperature and gravity waves. Here we utilize two consecutive years (2003–2004) of the MTM intensity data to determine the climatology (occurrence frequency and horizontal parameters) of a large variety of small-scale ripple events imaged in both the OH and O\textsubscript{2} nightglow emission layers at low-latitudes.

To investigate even faint ripple patterns the raw MTM imagery were flat-fielded separately for each emission of a circular 90° field of view (OH and O\textsubscript{2}) in order to correct for non-uniformities in the image brightness. The flat-fielded OH and O\textsubscript{2} data were then calibrated using the background star-field in order to get a geographic orientation [e.g., Hapgood and Taylor, 1982]. The calibrated flat-fielded images were then unwarped by projecting onto a rectangular grid of size 180 km × 180 km [García et al., 1997; Pautet and Moreels, 2002].

The horizontal wavelength and propagation direction of each ripple event was then computed using a three-dimensional Fast Fourier Transform (FFT) spec-
tral analysis technique to determine their unambiguous spectrum using a time series of typically 4 images [Garcia et al., 1997; Coble et al., 1998]. As most ripple patterns were spatially localized and usually comprised several well-defined wave crests their apparent speeds were determined separately by measuring the horizontal displacement of selected wave crests as a function of time. The observed periods were then computed from measured wavelengths and apparent speeds. Finally, the duration of the ripple events was estimated by observing the sequence of OH and O$_2$ images in ∼5.5-minute intervals of time. Due to their high coherence the typical uncertainties in the ripple measurements were ±1 km for their horizontal wavelength, ±3 ms$^{-1}$ for their observed motion, and <4° for their direction of motion.

Data from the 8th of January 2003 provide good examples of the quality of the ripples imaged in the OH and O$_2$ emissions from Maui, as well as their properties and variability. On this night three distinct ripple events were detected in the overhead sky at Maui together with some gravity wave activity. Here we focus on describing the ripple events. Figure 4.1 contains example unwarped flat-fielded airglow images of the three ripple events that were observed during the course of this night (from 06:37 UT to 16:01 UT). From the 9.4 hours of clear sky measurements ripples were recorded for a total 81 minutes in the OH data and 55 minutes in the O$_2$ data. Initially, there was very little wave activity present in either emission. However, around 08:05 UT an isolated ripple event was detected in both emissions in the NE quadrant of the imager field. By 08:19 UT at least eight ripple structures were clearly imaged in the O$_2$ emission (Figure 4.1a) while they appeared fainter and less distinct in the OH emission (Figure 4.1b). In both emissions the ripples exhibited a mean horizontal wavelength of ∼13 km and moved uniformly towards the north-west (azimuth ∼334° N (as indicated by the arrow), with an apparent speed of ∼37 ms$^{-1}$, indicating an observed period of ∼6 minutes. This event lasted for ∼17 minutes and
Figure 4.1. Examples of flat-fielded and unwarped images of ripple events observed on January 8, 2003. The top panel shows images of ripple events in the O$_2$ emission at (a) 08:19 UT, (c) 10:20 UT, and (e) 13:56 UT. The bottom panel shows images of ripple events in the OH emission at (b) 08:16 UT, (d) 10:17 UT, and (f) 13:53 UT.

was most prominent in the O$_2$ emission.

About 90 minutes later a second more extensive set of ripples was observed clearly in both emissions. This event lasted for $\sim$40 minutes and is shown in Figures 4.1c and 4.1d as it appeared at 10:20 UT in the O$_2$ emission and at 10:17 UT in the OH emission. The ripples comprised a similar number of wave crests as the first event, but were more spatially extensive and complex in their structure. The motion of this event was towards the east (at $\sim$83°N azimuth) significantly different from the first ripple event while its horizontal wavelength ($\sim$11 km), speed ($\sim$35 ms$^{-1}$), and observed period ($\sim$5.3 min.) were quite similar to the first event. Thereafter significant larger-scale band-type gravity wave structuring developed in
both emissions that persisted for the rest of the night and was most prominent in the OH emission. Around 13:35 UT a third well-defined ripple event was clearly observed in the OH emission, but was barely detectable in the O\textsubscript{2} emission. This event, duration $\sim$23 minutes, was much smaller in area than the previous two events and is shown in Figures 4.1e and 4.1f for comparison. The direction of motion of this ripple event towards the $\sim$W (azimuth 286° N) was again distinct from the two previous events but its wave characteristics were quite similar (horizontal wavelength $\sim$11 km, horizontal speed $\sim$29 ms$^{-1}$, observed period $\sim$6.2 min.).

For comparison, Figure 4.2 shows the evolution of a very extensive set of fine-scale ripples that developed over Maui on August 21, 2004 during a $\sim$70 minutes period, significantly longer than a typical ripple event. The ripples were prominent in both emissions but close inspection of the data indicate considerable evolution of the wave display suggesting the sequential development of more than one ripple event in adjacent areas within the camera field and exhibiting slightly different wave characteristics. Figure 4.2 shows three snap-shot images illustrating the evolution of this display observed simultaneously in the O\textsubscript{2} (upper panel) and OH (lower panel) emissions. Initially a set of over 10 well-defined wave crests developed in the upper left quadrant of the imager field around 09:30 UT. This situation is depicted in the O\textsubscript{2} emission at 09:45 UT (Figure 4.2a) and in the OH emission at 09:42 UT (Figure 4.2b). The bright diagonal band crossing the images is the Milky Way and at this time the ripples are evident to the left of the Milky Way, with no other structures present in the camera field. This ripple event lasted for $\sim$35 minutes and exhibited a horizontal wavelength of $\sim$8 km and moved towards the NE (azimuth 39° N) with an apparent speed of 23 ms$^{-1}$ (yielding an observed period of $\sim$6 min.).

Around 10:06 UT this ripple event began to fade and a second set of ripples started to grow in the lower half of the camera field. This is illustrated in the O\textsubscript{2} and
Figure 4.2. Examples of flat-fielded and unwarped images of typical ripple event observed on August 21, 2004. Panes (a), (c), and (e) depict images in the $O_2$ emission at 09:45 UT, 10:09 UT, and 10:26 UT. Panes (b), (d), and (f) depict images in the OH emission at 09:42 UT, 10:06 UT, and 10:23 UT, respectively.

OH images of Figure 4.2c and 4.2d, respectively. By $\sim$10:25 UT (Figures 4.2e and 4.2f) the second event is well developed, containing over 15 crests rotated $\sim$6° further around towards the NE (azimuth 45° N) that the earlier set of ripples. Their average horizontal wavelength ($\sim$7 km) and apparent speed ($\sim$21 ms$^{-1}$) were also slightly smaller (yielding an average observed period of $\sim$5.7 min.), but these differences are not significant within the uncertainty of our measurements. A summary of the measured characteristics of the ripple events described in Figures 4.1 and 4.2 is given in Table 4.2. These data illustrate well the properties and variability of the ripples imaged in the OH and $O_2$ emissions during both the summer and winter periods. The exceptionally long duration of this ripple event ($\sim$70 min.) is singular in our
Table 4.2. Data summary of observed horizontal parameters with duration of the ripples as determined from the image data: wavelength (\(\lambda\)), direction of propagation (\(\theta\)), apparent speed (\(c\)), and observed period (\(\tau\)).

<table>
<thead>
<tr>
<th>Date</th>
<th>Time (UT)</th>
<th>Emission</th>
<th>(\lambda) ±1 km</th>
<th>(\theta) ±3 deg.</th>
<th>(c) ±2 m/s</th>
<th>(\tau) ±0.5 min.</th>
</tr>
</thead>
<tbody>
<tr>
<td>01/08/2003</td>
<td>08:04-08:21</td>
<td>OH</td>
<td>13.2</td>
<td>335</td>
<td>36</td>
<td>6.1</td>
</tr>
<tr>
<td></td>
<td>08:07-08:24</td>
<td>O(_2)</td>
<td>13</td>
<td>334</td>
<td>37</td>
<td>5.9</td>
</tr>
<tr>
<td></td>
<td>10:00-10:41</td>
<td>OH</td>
<td>11.1</td>
<td>82.2</td>
<td>35</td>
<td>5.3</td>
</tr>
<tr>
<td></td>
<td>10:03-10:38</td>
<td>O(_2)</td>
<td>11</td>
<td>83</td>
<td>34</td>
<td>5.4</td>
</tr>
<tr>
<td></td>
<td>13:35-13:58</td>
<td>OH</td>
<td>10.8</td>
<td>286</td>
<td>29</td>
<td>6.2</td>
</tr>
<tr>
<td>08/21/2004</td>
<td>09:31-10:06</td>
<td>OH</td>
<td>8.2</td>
<td>39.2</td>
<td>23.2</td>
<td>5.9</td>
</tr>
<tr>
<td></td>
<td>09:34-10:09</td>
<td>O(_2)</td>
<td>8.1</td>
<td>39</td>
<td>23</td>
<td>5.9</td>
</tr>
<tr>
<td></td>
<td>10:06-10:41</td>
<td>OH</td>
<td>7.3</td>
<td>45.1</td>
<td>21.2</td>
<td>5.7</td>
</tr>
<tr>
<td></td>
<td>10:09-10:44</td>
<td>O(_2)</td>
<td>7.2</td>
<td>45</td>
<td>21.1</td>
<td>5.7</td>
</tr>
</tbody>
</table>

dataset and our analyses suggest the development of at least two events. However, it is possible that this ripple event may have resulted from an unusually long duration instability that evolved significantly in time and location.

Before presenting results on the seasonal ripple characteristics we first investigate the detection of ripples in the overhead sky in either the OH, O\(_2\) or jointly in both emissions for 2003 and 2004 as shown in Figure 4.3. A total of 261 events were observed in 2003 with 76 events in OH emission, 67 events in O\(_2\) emission and 118 events present in both emissions. Figures 4.3a and 4.3b are histogram plots of the monthly number of ripple events observed in either the OH or the O\(_2\) emissions only. Both plots show significant variability during the course of the year but, on average, the number of ripple events observed independently in the OH or the O\(_2\) emission was \(\sim\)6 events per month. In contrast, Figure 4.3c shows that the number of ripple events observed simultaneously in both the OH and O\(_2\) emission layers was significantly higher at \(\sim\)10 events per month. Figures 4.3d, 4.3e and 4.3f show the corresponding plots for the 2004 ripple dataset when a smaller total number of 227 events were recorded of which 101 were imaged in both emissions. With adjustment
Figure 4.3. Histogram plots in the left column showing the distributions of monthly ripple events observed (a) in the OH emission, (b) in the O$_2$ emission, and (c) in both the OH and O$_2$ for 2003. Histogram plots in the right column showing the distributions of monthly ripple events observed (d) in the OH emission, (e) in the O$_2$ emission, and (f) in both the OH and O$_2$ emissions for 2004.
for the winter months (when the observing conditions were more limited), they show very similar ripple event rates for OH, O$_2$ and OH/O$_2$ to those of 2003. Thus, our dataset comprises a total of 488 independent events of which almost 50% were observed to occur simultaneously in both emissions while independent events in either the OH or O$_2$ emissions occurred with approximately equal probability of $\sim 25\%$.

4.3 Climatology Results

In this section we present results of our statistical analyses of these ripple events imaged over two consecutive years (2003 and 2004), focusing on their occurrence frequency, lifetimes, wave characteristics, and seasonal variability.

4.3.1 Seasonal occurrence frequency

To investigate any seasonal variability in ripple occurrence, we have summed the OH and O$_2$ data into monthly bins counting individual events in either emission but, as discussed above, when the same ripple event occurred simultaneously in both OH and O$_2$ emissions it was counted as a single event as shown in Figure 4.4. Figure 4.4a plots the total number of ripple events as a function of month of year for 2003. Ripples were observed in every month but the number of detections varied significantly (from $\sim 10$ to 35) events with a distinct peak during the early summer months and minima during both the spring and fall. As the ripples observations are subject to clear sky observing conditions Figure 4.4a also plots the number of clear nights as function of month (dashed line). The observing conditions at Haleakala were very good throughout 2003 with typically $\sim 24$ nights of data per month (total 284 nights). Due to limitations from moonlight our observations varied from $\sim 3$–11 hours, with the majority of measurements $> 6$ hours duration/night. Figure 4.4b plots the fractional ratio of the two datasets shown in Figure 4.4a to more clearly assess the variability of ripples as function of season. The resultant plot shows a
Figure 4.4. The top panel plots show frequency of occurrence of ripple events (solid line) and total number of observation nights (dashed line) recorded by month in (a) 2003 and (c) 2004. The bottom panel plots show the ratio of number of ripple events per month to the total number of clear image nights for (b) 2003 and (d) 2004.

distinct peak during the early summer months (May–June) with minima around the spring and fall equinox periods.

Figures 4.4c and 4.4d show the results of a similar analysis applied to the 2004 ripple dataset. Variable weather conditions during the first quarter of the year (indicated by the dashed line) limited the image measurements, but thereafter the number of observation nights per month (20 nights per month) was very similar to that of 2003 (total 243 nights). The number of ripple events per month (Figure 4.4c) shows a somewhat broader peak during the summer period, with minima in
the spring and fall. This result is also evident in Figures 4.4d, which plots the ratio of the ripple observations per month to the total number of clear nights. The similarity between this plot and the data of 2003 (Figure 4.4b) is evident. Both plots exhibit significant peaks in ripple occurrence during the summer months with lowest number of events occurring during the spring and fall equinoxes, and somewhat reduced ripple activity during the winter months.

To further investigate this apparent seasonal relationship, Figure 4.5a plots the ratio of number of ripple events per month (solid curve in Figure 4.4a) to the total number of clear sky hours in each month for 2003 (typically ~165 hours per month). A similar plot is also shown in Figure 4.5b for 2004 (typically ~143 hours per month). Both of these plots indicate a mean monthly occurrence frequency for ripple events of ~13% per hour of observation throughout the year (for our limited 180 km x 180 km zenith field of view). However, they also show significant variability about this mean, especially during the early summer months, when their occurrence rate increases to > 20% per hour and reduces to a minima of ~6–10% per hour during the equinox periods. Together with Figure 4.4 these results provide new evidence

Figure 4.5. Plots show the frequency of occurrence of the ratio of number of ripple events per month to the total number of hours of clear image nights in each month for (a) 2003 and (b) 2004.
for a seasonal variability in the occurrence frequency of ripples.

4.3.2 Ripple lifetimes and nocturnal occurrence rate

From the literature it is well known that ripple events exhibit relatively short lifetimes, typically < 45 min., which is usually much less than the observed durations of freely propagating (and/or ducted) gravity waves imaged in the mesospheric airglow emissions. To investigate the range of ripple durations more quantitatively, Figure 4.6a plots the number of ripple events verses their observed duration in min. The plot separately shows all the OH and O$_2$ ripple durations summed in 5 min. bins over the 2003–2004 period (over 345 events each). In both emissions the ripple durations ranged from ~15 minutes up to ~50 minutes with a strong peak (~60%) of ripples exhibiting lifetimes of ~15–20 min. The occurrence frequency then decreases rapidly to < 20% in the next sample bin, 20–25 min. This plot clearly shows a marked preponderance for relatively short duration ripples (< 25 min.) at low-latitudes, and that longer period events occurred, but were far less common. To investigate any seasonal variability in the duration of ripple events, Figure 4.6b plots their average

![Figure 4.6](image_url)

**Figure 4.6.** Histogram plots showing the distribution of (a) number of ripple events as function of ripple duration in minute, and (b) average duration (minute) of ripples per month recorded in OH and O$_2$ emissions for a two-year period (2003–2004).
duration as a function of month summed over the 2003–2004 period. The distribution shows little or no seasonal variation with the mean monthly event duration of \( \sim 22 \) minutes. As expected almost identical results was obtained for both the OH and \( \text{O}_2 \) ripple datasets.

To investigate how often ripples are seen, Figure 4.7 plots the number of clear sky nights (which we have defined earlier as typically \( > 6 \) hours of cloudless observations per night), versus the number of ripple events imaged either in the OH, \( \text{O}_2 \) or jointly in both emissions. The results from 2003 and 2004 are plotted separately, but both yield near identical distributions. For example, the plot shows that out of a total of 284 clear nights in 2003 (Figure 4.7a) one or more ripple events were observed on 127 nights indicating a detection rate of \( \sim 45\% \). A similar value of \( \sim 40\% \) was found for 2004 (Figure 4.7b). The plot also shows that of the 127 nights when ripples were detected in 2003 they were most frequently seen only once during the night \( (\sim 54\%) \), while for two events/night the percentage was significantly lower at \( \sim 21\% \). In contrast, the probability of observing more than two events remained good at \( \sim 25\% \). The maximum number of ripple events detected within our zenith

![Clear Sky Nights Vs Events](image1)

![Clear Sky Nights Vs Events](image2)

**Figure 4.7.** Plots show the number of clear sky nights as function of number of ripple events detected per night (a) in 2003, and (b) in 2004.
centered field of view (\(\sim 90^\circ\)) was eight per night. These results show that although ripple events are relatively common they certainly do not occur on all clear nights at a given site.

4.3.3 Ripple wave characteristics

Figure 4.8 summarizes the measured distributions of the ripple horizontal wavelengths, horizontal speeds and their observed periods. For this study we have chosen to include measurements of all of the ripple displays as recorded separately in the OH and O\(_2\) emissions during 2003 (OH 194 events; O\(_2\) 185 events) and 2004 (OH 166 events; O\(_2\) 162 events). The top panel (Figure 4.8a) compares the distribution of horizontal wavelengths in 2 km bins for both years. The OH and O\(_2\) distributions are similar in shape to each other, but differ slightly from year to year. This said, they all exhibited a relatively narrow range of ripple wavelengths ranging from \(\sim 6\) to 18 km. In 2003, horizontal scales of \(\sim 8\)–12 km were most frequent, whereas in 2004 the range of ripple wavelengths was slightly broader \(\sim 8\)–14 km. The mean horizontal wavelength was \(\sim 11\) km in both years. Figure 4.8b show the corresponding distributions of the ripple horizontal speeds for 2003 and 2004, plotted at 5 ms\(^{-1}\) intervals. Both histogram plots show a strong tendency for horizontal speeds in the 20–30 ms\(^{-1}\) range (average 27.4 \(\pm\) 0.5 ms\(^{-1}\)) with maximum values up to 50 ms\(^{-1}\), and no events with observed speeds < 20 ms\(^{-1}\).

The resulting distributions of observed periods of the ripple events for 2003 and 2004 are shown in the bottom panel (Figure 4.8c), binned at 1-minute intervals. In both years the distributions of ripple periods exhibited a relatively narrow range extending from \(\sim 5\)–10 min. and similar peak values at \(\sim 6\)–7 min. The average ripple periods for OH and O\(_2\) data combined was 6.4 \(\pm\) 0.1 min. for 2003 and 6.9 \(\pm\) 0.1 min. for 2004. These results agree very well with previous measurements by Taylor et al.
Figure 4.8. Histogram plots showing the distribution of (a) horizontal wavelengths, (b) horizontal speeds, and (c) observed periods for ripples as a function of number of ripple events observed in OH and O\textsubscript{2} emissions for 2003 and 2004.
[1997], Nakamura et al. [1999], and Medeiros et al. [2004, 2007] who reported similar values for the average values (and observed ranges) for the horizontal wavelengths, horizontal speeds and derived periods of ripple events imaged in the MLT nightglow emissions from equatorial and low-latitude sites, as summarized in Table 4.1.

4.4 Discussion

Our ripple analyses constitute the most comprehensive study of ripple characteristics to date. The data set are rich in information and comprise two consecutive years enabling much more detailed studies than have previously been possible. Furthermore the combination of OH and O$_2$ data has provided new insight into the occurrence and properties of ripples at two separate altitudes $\sim$87 and $\sim$94 km.

The results of the seasonal investigation on the ripple occurrence frequency were most unexpected, as to our knowledge this has not previously been quantified. The very good observing conditions afforded at the summit of Haleakala and the subsequent large number of ripple events measured has made this study possible. Our analysis first looked at the monthly number of ripple events as a function of season (Figure 4.4) for the total number of events (OH and O$_2$ combined) and strong peaks were found in both the 2003 and 2004 results indicating highest occurrence of ripples during the summer months with lowest number of events during the spring and fall equinoxes, and somewhat reduced ripple activity during the winter months. This result was confirmed in Figure 4.6 which showed similar seasonal behavior for the number of events/month normalized by the number of clear sky observing hours. These results provide new evidence for a seasonal variability in the occurrence frequency of ripples. Prior studies by Nakamura et al. [1999] and Medeiros et al. [2007] suggested that ripples were more common during the winter months but no account of the observing conditions was presented.
It is well known in the literature that ripples are relatively common events, but detailed information on their nocturnal occurrence and lifetimes is generally lacking. Our analysis, based on a field of view of 90° centered on the zenith, shows that ripples are seen on about 45% of nights and that their durations are typically < 20 min. On some active nights several ripple events are observed; however, it is more normal to detect only one or two events. This said, the frequency of occurrence on ripple active nights is expected to be significantly higher (∼ factor 4) for all-sky (180°) airglow data due to the much larger sample area.

As anticipated, measurements of the characteristics of ripple events (horizontal wavelength, horizontal speed, and inferred period) show similar results to prior studies, as listed in Table 4.1. However, seasonal studies of the direction of motion of the ripples are few. Figure 4.9 plots yearly summaries of the ripple propagation headings at Maui for 2003 and 2004 in the form of polar block diagrams with 15° interval bins. This format has been adopted by several previous investigations of gravity wave anisotropy [e.g., Taylor et al., 1993, 1997; Medeiros et al., 2004, 2005]. In Figures 4.9a and 4.9b the OH and O2 data are plotted separately for 2003. It is immediately clear that the ripples exhibited a strong anisotropy in their preferred directions of motion, which was very similar in both emissions. In particular, about 75% of events exhibited a northward (meridional) component of motion with almost equal preference for east-west (zonal) motion. A similar situation prevailed in 2004 with dominant northward ripple propagation in both emissions, but their zonal components of motion were mainly eastward as shown in Figures 4.9c and 4.9d.

To investigate seasonal contributions to this anisotropy, Figure 4.10 re-plots the OH data for 2003 (Figure 4.9a) dividing it into four seasons, with summer and winter each comprising four months of data (May, June, July and August) and (November, December, January and February), while spring and fall each contained
two months of data (March and April) and (September and October), respectively. Figure 4.10a and 4.10b show the summer and winter seasons which comprised a similar number of ripple events (∼72 each), and for comparison the data are plotted on the same scales. During the summer months (Figure 4.10a) the data reveals a strong focus for ripple propagation towards the north-east (azimuth range ∼30–75° N), with very few events at other azimuths. In contrast, the winter data (Figure 4.10b) show a distinct preference for westward ripple motions, but over a boarder
Figure 4.10. Polar block diagrams showing the azimuthal distribution of ripple events observed in the OH emission for (a) summer, (b) winter, (c) spring, and (d) fall seasons of 2003.

The lack of eastward propagation during winter produces a distinct anisotropy between the summer and winter ripple motions. The spring and fall data (Figures 4.10c and 4.10d) contained significantly lower numbers of ripples events and for clarity are plotted on a smaller scale. During the spring-time focused northeastward and westward ripple motions occurred, and detailed inspection of the data suggests that they were essentially a combination of the late winter and early summer distributions. The fall data (Figure 4.10d) contained only 17 events and no anisotropy was evident.
in these limited data.

To further investigate the strong summer-winter differences in the ripple headings observed in the OH data we have also studied the seasonal variability of the O$_2$ ripple motions observed during 2003. The results are given in Figure 4.11 and reveal a very similar situation with dominant northeastward motion during the summer months and persistent westward motion during the winter time. This result is not unexpected as mentioned earlier $\sim$50% of the ripple events were common in both emissions. Nevertheless, the clear similarity between these two separate datasets provides convincing evidence of strong seasonal differences in the ripple alignment and headings from summer to winter during 2003.

Based on these findings we have also investigated seasonal differences in the 2004 OH and O$_2$ ripple dataset as shown in Figures 4.12 and 4.13, respectively. The results are essentially the same as reported for 2003 with highly focused northeastward motion during the summertime, and strong westward motion during the winter months. However, the lower number of ripple events observed during the wintertime (mainly due to poor weather conditions) created a more restricted azimuthal range for the 2004 dataset. The effects of this on the winter azimuthal distribution will be discussed later.

As discussed earlier evidence for seasonal variations in the propagation directions of ripple is very sparse, and of the three long-term datasets reported in the literature (summarized in Table 4.1), only one has determined evidence of seasonally related anisotropy. Medeiros et al. [2007] analyzed over four years of all-sky image data from Cariri, Brazil ($\sim$7° S) comprising a mixture of 525 ripple events observed in the OH, O$_2$ and the OI (557.7 nm) green line emission (at $\sim$96 km altitude). While their observed ripple characteristics were almost identical to those reported herein and previously in the literature, they found compelling evidence for
Figure 4.11. Polar block diagrams showing the azimuthal distribution of ripple events observed in the emission for (a) summer, (b) winter, (c) spring, and (d) fall seasons of 2003.

a strong anisotropy in the direction of motion of ripple events. In particular, ripples were more common during the winter months and exhibited a strong preference for northward and/or southward motion. While during the summer months the ripple motions were predominantly northwards. Throughout their summed year, east-west ripple motions were much more restricted. The results of their summer and winter study are shown in Figure 4.14, which is plotted in the same format as our Maui results (Figures 4.10–4.13), to permit easy comparison. During the summertime both datasets (Cariri and Maui) show strong preference for northward ripple propagation
and also a significant eastward component of motion (that is stronger in the Maui data). In contrast, the wintertime datasets show an almost anti-correlation with strong north-south motions prevailing at Cariri while the Maui data are overwhelmingly westward, with only limited north-south components of motion. Importantly, our Maui data show a distinct seasonal change from summer to winter during two consecutive years, whereas the four-year averaged data from Cariri reveal strong north-south ripple motions throughout the year and seasonal changes in direction are not easily discerned. Possible reasons for these differences are discussed below.
Figure 4.13. Polar block diagrams showing the azimuthal distribution of ripple events observed in the $\text{O}_2$ emission for (a) summer, (b) winter, (c) spring, and (d) fall seasons of 2004.

In their investigation of wave climatology over South America Medeiros et al. [2007] also plotted azimuthal distributions of short-period (< 1 hour) gravity wave events. As discussed earlier, these are spatially extensive, long-lasting (several hours) quasi-monochromatic wave patterns (often termed bands), that are mainly attributed to the passage of freely propagating gravity waves or ducted waves within the airglow layers [e.g., Taylor et al., 1987, 1995b,c, 1997; Isler et al., 1997; Walterscheid et al., 1999]. Seasonal variations in gravity wave propagation is expected due to changes in the dominant tropospheric source locations during the year and
to the cumulative effects of background winds which can filter out waves (termed critical layer filtering), especially those with low phase speeds as they propagate upwards through the intervening stratosphere and lower mesosphere [e.g., Taylor et al., 1993]. Medeiros et al. [2007] determined similar shaped azimuthal distributions for the band motions (range $\sim$10–70 ms$^{-1}$), as compared with the ripples, again with strong north-south preference throughout the year. This prompted them to comment on the general similarity of ripple and band wave characteristics (with the exception of duration and spatial extent), and the possibility of common causes for both types of wave events. While their concluding statement goes against recent studies that have definitively shown that some ripple events are generated in situ by short-lived convective or dynamical instabilities, the strong similarities between the ripple and band azimuthal distributions is an important finding and bears further study.

To test this hypothesis we have extended our ripple analysis to include measurements of quasi-monochromatic gravity wave events that were also imaged by the MTM during the same period as our ripple study. Figure 4.15 contains four polar
plots showing the vector velocity of 168 distinct band events that were recorded during the summer and winter of 2004. Most of these events were common in both OH and O$_2$ emissions (154 events). This format provides more details on the individual events and clearly shows a strong seasonal dependency of the propagation directions of bands. While seasonal changes in band motions have been previously reported from a number of sites around the world [e.g., Taylor et al., 1997; Nakamura et al., 1999, 2003; Suzuki et al., 2004; Medeiros et al., 2004; Pautet et al., 2005; Nielsen et al., 2009], our results from Maui are particularly distinct. During the summertime nearly all of the waves (>100 events) exhibited velocity vectors, that ranged in azimuth from 0–120°, indicating a very strong preference for generally northeastward wave propagation. In contrast, during the winter months (~56 events) the preferential propagation direction was bimodal in form, with well-defined central wave azimuths towards the southwest and the northeast. Close inspection of these results revealed a further distinction in the wave propagation headings: with >75% of waves imaged during November–December period propagating towards the southwest, while over 80% of the waves imaged during January and February progressed towards the northwest. Thus, the bimodal nature of the winter wave results is due mainly to a change in dominant direction of motion during the course of the winter, and has also been observed at mid-latitudes in 2002 from Bear Lake Observatory [Simkhada et al., 2006]. Systematic changes in the dominant wave azimuth during the course of the winter season have also recently been reported by Nielsen et al. [2009] using OH data from Halley Station, Antarctica (76° S).

Comparison of these band results with the seasonal ripple measurements shown in Figures 4.10–4.13 (for both 2003 and 2004) reveals a remarkably consistent pattern. During the summer months the ripple and the band events predominantly progressed in the same direction towards the ~northeast. During the winter
months a similar situation appears to occur with the ripples exhibited a dominant westward motion but over a broader azimuth range that encompasses the observed bi-modal band motions. However, further inspection of our results shows that the situation during the winter months is more complex and revealing. Figure 4.16 re-plots the 2004 ripple and band azimuthal data of Figures 4.12 and 4.15 for the OH emission only. The data have been divided into two periods November–December and January–February to better investigate their relationships. During November–
Figure 4.16. Polar plots in the left panel show the vector velocity of ripples recorded in the OH emission for (a) November–December, and (b) January–February of 2004. Polar plots in the right panel show the vector velocity of bands recorded in the OH emission for (c) November–December, and (d) January–February of 2004.

In December (Figure 4.16a) the majority of the ripples exhibited motions towards the northwest, while at the same time the bands (Figure 4.16c) show mainly motion towards the southwest, i.e., approximately orthogonal to the ripple motions. Likewise, during the January–February period the majority of the ripples (Figure 4.16b) moved towards the ∼southwest while the bands (Figure 4.16d) progressed towards the ∼northwest. Thus, our data also show a clear relationship between the band and ripple motions; however, they appear to be co-aligned in summer but near or-
thogonal in winter. The reasons for this are currently under investigation but may have their roots in the instability origin of the ripples.

Given the fact that ripples are transient, spatially localized wave patterns it is most probable that their sources are distinct from the much larger and longer lasting band events. Indeed, our present studies presented in Chapter 5 utilize coincident airglow imaging and lidar wind/temperature measurements to show that some small-scale ripple structures are locally generated by dynamic type instabilities. This result is good agreement with other recent studies of ripples, and argues against the statement of Medeiros et al. [2007] that ripples and bands are of similar origin. However, the fact that our results also show good agreement with results of Medeiros et al. [2007] argues that they are related possibly by the following mechanism.

It is now well established that the atmosphere in the mesopause region may become unstable due to the passage of large scale tides and gravity waves through this region and subsequent perturbation of the background wind and temperature field. Taylor and Hapgood [1990] discussed the possibility of ripple formation by the chance superposition of waves that can lead to temporary but strong wind shears that result in the generation of Kelvin-Helmholtz (dynamical) instabilities. This work was inspired by studies of noctilucent clouds which also show similar ripple-type waves termed billows [Witt, 1962; Fogle and Haurwitz, 1969]. Recent airglow and lidar observations by Hecht et al. [2005], Li et al. [2005a], and Li et al. [2005] have revealed that strong wind shears are capable of causing dynamical instabilities resulting in ripples at MLT heights. A second type of instability is also known to occur. Convective instabilities are associated with temporary steepening of the atmosphere temperature gradient producing a superadiabatic lapse rate [Li et al., 2005b]. Fritts et al. [1993] and Hecht et al. [1997, 2000] have shown that breaking of large-scale gravity waves or wave-wave interaction can produce this type of insta-
bility. Indeed, observations by Hecht et al. [1997] and numerical modeling by Fritts et al. [1997] suggest that convective instabilities induced by a larger scale gravity wave can also be a significant source of small-scale ripples.

These studies have shown that ripples produced by wind shear (K-H instabilities) are naturally orientated perpendicular to the wind shear vector. As such they would tend to be aligned parallel with the wave-crests of the larger-scale perturbing wave. In contrast, modeling studies of instability structures produced by convective overturning appear near perpendicular to the perturbing wave fronts [Fritts et al., 1997]. Using this knowledge we suggest that the apparent alignment of the Maui ripple events with the short-period gravity wave motions during the summer months would indicate a preference for the wind shear mechanism for their generation, whereas the near-perpendicular alignment of the bands and ripple motions during the winter months suggests a preference for convective type instabilities. Our analyses have identified this apparent relationship and while this hypothesis is mainly speculation at this stage it does provide a plausible explanation for the observed similarities between band and ripple motions as reported by Medeiros et al. [2007] and our Maui results presented herein.

We can investigate this hypothesis a little further using published measurements of convective and dynamical instabilities obtained by lidar. Zhao et al. [2003] used high resolution Na wind-temperature data obtained from the Starfire Optical Observatory (SOR) in New Mexico (35° N) to gain first insight into summer and winter variability of instabilities in the MLT region. Using 32 nights of data they determined the probabilities of convective and dynamic instabilities were maximum in the mid-winter with almost equal values of ∼11% and minimum is the summer when the average probabilities were lower at ∼6% for both mechanisms. These results suggest that ripples should be more common during the winter months but does
not help to distinguish between their possible mechanisms. Li et al. [2005b] used 19 nights of Na wind-temperature lidar data from Maui (20.7° N) obtained during campaigns (2002–2005) as part of the Maui-MALT program. They found probabilities of convective and dynamical instabilities over the MLT region (85–100 km) were 2.9% and 10.5%, respectively. As our ripple data were obtained as part of the same program (but over more limited period 2003–2004) these lidar results are most applicable. Although the results of Li et al. [2005b] were not seasonal they do strongly suggests that dynamical instabilities were much more common over Maui than convective instabilities. Analysis of our OH and O\textsubscript{2} ripple data shows that ripples were seen with almost equal numbers during the summer (96 events) and wintertime (94 events) for 2003. Invoking our hypothesis we would expect to see more ripples during the summer period (assuming similar observing conditions) than the wintertime, as the summer data were considered to be associated with dynamical instabilities. However, this was not the case. Analysis of the 2004 ripple data shows significantly more ripples during the summertime (109 events) than the winter period (56 events), which agrees qualitatively with our hypothesis but as discussed earlier the reduced number of ripple (and band) events during the winter months appear to be due to poorer sky observing conditions in 2004.

Most recently, Sherman and She [2006] reported a seasonal investigation using one year (May 2002–April 2003) of Na wind-temperature lidar data obtained at mid-latitudes from Ft. Collins, CO (41° N). Using a total of 63 nights of lidar data taken during the course of the year they determined a summertime probability of instability occurrence of 4.3%, and a wintertime probability of 7.3%. These results are in qualitative agreement with Zhao et al. [2003] but the probabilities for summer and winter are both somewhat lower in magnitude. Sherman and She [2006] also reported on the seasonal occurrence of the type of instability. They found that dynamical
instabilities were more frequent throughout the year (mean 4.4%) compared with convective instabilities (mean 0.9%). These data again suggest that ripples should be more common during the summer and that they are most frequently due to strong wind shears (typically $> 40 \text{ ms}^{-1} \text{km}^{-1}$).

The significant variation in the results of lidar studies may be due to the latitude of the sites, as the MLT wind field is strongly influenced by tides. For example, the diurnal and semi-diurnal tides exhibit strong latitudinal variations. Unfortunately, the lidar studies at Maui were limited and a seasonal study of the occurrence rates of dynamical and convective instabilities was not possible. Our ripple measurements indicate clear seasonal variability in their preferred directions of motion and also show that ripples are more frequently observed during the summer (Figures 4.4 and 4.5) with a somewhat reduced activity during the wintertime in general agreement with the lidar measurements. Furthermore, it is common wisdom that although the mean state and tides are responsible for preparing the background atmosphere for instabilities (of either kind) to develop, it is the superposition of a short-period (band-type) gravity wave that cause the atmosphere to temporarily become unstable [Hecht et al., 1997; Liu et al., 2004]. Thus, the apparent association of ripples with bands is not unexpected. Our hypothesis which links summertime ripples to dynamical instabilities and wintertime to convective instabilities also seems reasonable based on the observed correlative motions of the ripple and band datasets. However, the relatively low occurrence rate of convective instabilities during the winter months (as determined by the lidar measurements) raises questions concerning their potential seasonal role in ripple formation. We are currently investigating these results further. A direct comparison of the ripple and lidar data for two dynamical instability events is presented in the following Chapter 5.
4.5 Summary

An extensive two year study of well-defined mesospheric ripple events observed from Maui, HI, has yielded important new information on their seasonal occurrence frequency, as well as anisotropy in their directions of motion suggesting a strong causal relationship with short-period gravity waves imaged during the same time. A total of 488 events were measured in 2003 and 2004 of which 219 ripple events were identified in both OH and O$_2$ image data. The spatial and temporal characteristics of the ripple events: horizontal wavelengths, horizontal speeds and observed periods were investigated and found to be typically $\sim 6$–18 km, $\sim 20$–50 ms$^{-1}$ and $\sim 5$–10 min., respectively, with mean values of 11 km, 27.4 ms$^{-1}$ and 6.7 min., in good agreement with previous studies performed at a number of different sites at equatorial, mid- and high latitudes.

Independent investigations of the ripple occurrence frequency on the OH and O$_2$ emissions revealed new evidence for a strong seasonal variability that was evident in both years and indicated highest occurrence rate during the summer months, with somewhat reduced ripple activity during the winter, and lowest number of events during the spring and fall equinoxes. To our knowledge this is the first measurement of seasonal variability in ripple activity. When the data were subdivided into the four seasons, detailed studies of their propagation headings has revealed a remarkably strong preference for northeastward motion during the summer months and westward motion during the wintertime. This pattern was observed for both consecutive years establishing a distinct seasonal change in ripple motions from summer to winter. These results build on the finding of Medeiros et al. [2007] who also reported evidence of anisotropy in ripple headings observed from Brazil.

Comparison of our ripple results with short-period gravity waves measured over Maui during the same period has revealed an intriguing relationship where rip-
ules observed during the summer months progressed in the same direction (towards the northeast) as the larger-scale bands. However, during the winter months the ripple motions were predominantly westward and near orthogonal to the bands. Based on recent studies which have shown that ripples are generated in situ in the MLT region mainly by convective or dynamical-type instabilities, we have hypothesized that the different relationships between bands and ripples may reflect different dominant instability processes summer and winter. New coordinated wind-temperature lidar and airglow imager seasonal studies are needed to establish the underlying cause of this relationship.
CHAPTER 5
SIMULTANEOUS OCCURRENCES OF KELVIN-HELMHOLTZ
INSTABILITY STRUCTURES IN TWO AIRGLOW LAYERS

5.1 Introduction

Small-scale instability features, known as ripples, generated in situ are ever-present in mesopause airglow layers, as reported by numerous airglow imager observations [Peterson, 1979; Taylor and Hapgood, 1990; Taylor and Hill, 1991; Swenson and Mende, 1994; Taylor et al., 1995f, 1997; Hecht et al., 1997, 2001b, 2005; Nakamura et al., 1999; Medeiros et al., 2004; Hecht, 2004; Li et al., 2005]. Several measurements indicate that they are easily distinguishable from propagating gravity waves [Taylor and Hill, 1991; Swenson and Espy, 1995; Taylor et al., 1997; Nakamura et al., 1999; Medeiros et al., 2004, 2007]. The ripples appearing in airglow images are typically limited in their spatial and temporal extent, with $\sim$4–16 phase fronts, horizontal separations of $\sim$5–15 km, and durations of less than an hour. Observational and theoretical studies suggest that ripples are generated in situ by local dynamical or convective instabilities [Fritts et al., 1993, 1997; Hecht et al., 1997, 2000, 2001b, 2005; Li et al., 2005a]. We note, however, that it is sometimes difficult to distinguish ripples indicative of instability dynamics from small-scale gravity waves (GWs) that are ducted at a stability maximum (and vertically evanescent above and below), and which have very different sources and exhibit very different dynamics [Fritts and Alexander, 2003].

Theoretical and numerical studies in the last two decades have provided a basic understanding of GWs and shear flow instability dynamics for idealized flows [e.g., Klaassen and Peltier, 1985, 1991; Klostermeyer, 1991; Fritts et al., 1993, 1994, 1996a,b, 1998; Andreassen et al., 1994, 1998; Palmer et al., 1994, 1996; Vanneste,
1995; Lombard and Riley, 1996; Sonmor and Klaassen, 1997; Smyth, 1999; Werne and Fritts, 1999, 2001; Achatz, 2005, 2007]. Only more recently, however, have we begun to address the two- and three-dimensional instability dynamics for more realistic Reynolds numbers and accompanying superpositions of motions occurring on multiple scales [Fritts et al., 2006, 2009a,b]. These studies have demonstrated clearly that instability dynamics are far more complex and diverse than previously believed. These studies have also shown that the traditional views of dynamical and convective instability thresholds at local Richardson numbers of \( Ri = 0.25 \) and 0, respectively, are at best qualitative guides to instability tendency and character and often do not adequately describe the occurrence of instability, even in highly idealized flows [Fritts et al., 2006, 2009a]. Nevertheless, numerical studies and theory provide better guidance, and have yielded more convincing results in the case of Kelvin-Helmholtz shear instability [Gossard and Hooke, 1975; Fritts and Rastogi, 1985; Hecht et al., 2001b, 2005; Li et al., 2005a, and references therein] for which instability orientations, wavelengths, growth rates, and the occurrence and nature of secondary instability are largely determined by the initial shear flow [Werne et al., 2005].

Our main purpose is to describe the characteristics of ripples from the results of airglow imaging analysis and their alignment and association with local wind shears in two airglow layers. We have observed small-scale, spatially localized ripples simultaneously in the mesopause airglow layers OH (87±3 km) and \( O_2 \) (94±3 km) from Maui, Hawaii (20.7° N, 156.3° W). Here, the Na lidar wind and temperature measurements were analyzed to examine the nature of the ripples associated with local instabilities in OH and \( O_2 \) airglow layers. The observation and data analysis is described in Section 5.2. Results and discussion of two observed ripple events, together with shear instabilities with data obtained from Na wind-temperature lidar as well as their comparison with numerical model case studies, are presented in
Section 5.3. Then these are followed by the summary and conclusions in Section 5.4.

5.2 Observations and Data Analysis

The USU Mesospheric Temperature Mapper (MTM) was utilized to sequentially observe ripple events in both OH and O$_2$ airglow emissions. The exposure time for each image observation was 60 seconds with two observations per emission, plus a background image, resulting in a cycle time of $\sim$5.5 min. The raw MTM images were processed separately for each emission (OH and O$_2$) using well-developed flat-field functions in order to correct the non-uniform intensity in the image. The data were then normalized using the appropriate zenith emission intensity to create a time series of images of uniform brightness across the camera’s 90° circular field of view. Next, the flat-fielded images were calibrated using the background star field in order to get a geographic orientation [Hapgood and Taylor, 1982]. The calibrated flat-fielded images were then unwarped by projecting them onto a uniformly sampled 180 km x 180 km rectangular grid [Garcia et al., 1997; Pautet and Moreels, 2002].

In order to determine the horizontal wavelength and propagation direction of ripple events, we used a 3-D Fast Fourier Transform (FFT) spectral analysis to process unambiguous spectrum using a time series of images [Garcia et al., 1997; Coble et al., 1998]. The apparent horizontal speeds of ripples were determined separately by measuring the horizontal displacement of selected wave crests as a function of time. The observed periods were then computed from measured wavelengths and apparent speeds. We have estimated the duration of the ripple events by observing the sequence of OH and O$_2$ images in $\sim$5.5-minute intervals of time.

The summary of characteristics of ripple events is tabulated in Table 5.1. A total of 15 ripple events were recorded on 10 nights of coincident lidar wind-temperature measurements during 2002–2005. During this period, 13 events were
Table 5.1. Airglow image data summary of measured horizontal parameters of ripple events: duration, wavelength ($\lambda$), direction of propagation ($\theta$), speed ($c$), and observed period ($\tau$) on the nights of coincident lidar measurements.

<table>
<thead>
<tr>
<th>Date</th>
<th>Time (UT)</th>
<th>Emission</th>
<th>$\lambda$ ±1 (km)</th>
<th>$\theta$ ±3 (deg.)</th>
<th>$c$ ±3 (m/s)</th>
<th>$\tau$ ±0.7 (min.)</th>
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<td>12:03-12:20</td>
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<td>10.2</td>
<td>292</td>
<td>32</td>
<td>5.3</td>
</tr>
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<td></td>
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<td>O$_2$</td>
<td>10.1</td>
<td>292</td>
<td>32</td>
<td>5.3</td>
</tr>
<tr>
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<td>12:12-12:29</td>
<td>OH</td>
<td>9.1</td>
<td>6</td>
<td>27</td>
<td>5.6</td>
</tr>
<tr>
<td></td>
<td>12:15-12:32</td>
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<td>9.3</td>
<td>6.5</td>
<td>28</td>
<td>5.5</td>
</tr>
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<td>10:47-11:05</td>
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<td>12.1</td>
<td>43</td>
<td>32</td>
<td>6.3</td>
</tr>
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<td>25</td>
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<td>26</td>
<td>6.7</td>
</tr>
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<td>302</td>
<td>26</td>
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<td>301</td>
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<td>316</td>
<td>24</td>
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<td>14.2</td>
<td>136</td>
<td>36</td>
<td>6.6</td>
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<tr>
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<td>08:13-08:36</td>
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<td>5</td>
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<td>7.8</td>
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<td>7</td>
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<tr>
<td></td>
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<td>OH</td>
<td>14</td>
<td>133</td>
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<td>6</td>
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<td>280</td>
<td>29</td>
<td>7.2</td>
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</table>

recorded in both OH and O$_2$ emissions, with only one event in the OH emission and one event in the O$_2$ emission. The horizontal wavelengths, horizontal speeds, and observed periods of these ripples were typically $\sim$7–14 km, $\sim$25–36 ms$^{-1}$ and $\sim$5–8 min., respectively. The typical measurement errors calculated as standard
deviation for the horizontal wavelength, horizontal speed, propagation direction and observed period were 1 km, 3 ms\(^{-1}\), 3 and 0.7 min., respectively. The temporal and spatial characteristics of these ripple events were consistent with other measurements [Peterson, 1979; Peterson and Adams, 1983; Taylor and Hapgood, 1990; Taylor and Hill, 1991; Taylor et al., 1995f, 1997; Nakamura et al., 1999; Hecht et al., 2001b, 2005; Medeiros et al., 2004; Li et al., 2005].

Figures 5.1 and 5.2 show two examples of ripple events observed simultaneously in OH and \(O_2\) airglow emissions on the nights of July 7, 2002 and March 8, 2005. Figure 5.1 contains at-Øelded and unwarped images of the July 7 ripple event.

**Figure 5.1.** Examples of at-Øelded and unwarped images of ripple events observed on July 7, 2002 (a) in the OH emission at 12:12 UT and 12:17 UT, and (b) in the \(O_2\) emission at 12:15 UT and 12:20 UT.
Figure 5.2. Examples of flat-fielded and unwarped images of ripple events observed on March 8, 2005 (a) in the OH emission at 13:08 UT and 13:14 UT, and (b) in the O$_2$ emission at 13:11 UT and 13:17 UT.

observed in the OH emission at 12:12–12:17 UT and in the O$_2$ emission at 12:15–12:20 UT. We have observed the event for 17 minutes from 12:12 to 12:29 UT in the OH emission and from 12:15 to 12:32 UT in the O$_2$ emission. The ripple event of a similar orientation pattern with ~6–9 wave crests in OH and O$_2$ emissions traveled with speeds of 27 and 28 ms$^{-1}$, respectively, towards north at ~6$^\circ$ as measured clock-wise from north. The event exhibits equal horizontal wavelength of ~9 km, yielding apparent periods of 5.6 and 5.5 min., respectively, in both the OH and O$_2$ airglow layers. The arrows in each panel indicate the direction of motion of the ripple.

We have recorded second ripple event for 41 minutes from 13:02 to 13:43 UT
in the OH emission and from 13:05 to 13:46 UT in the O\textsubscript{2} emission on March 8, 2005. Figure 5.2 displays unwarped flat-fielded images of this event observed in the OH emission at 13:08–13:14 UT and in the O\textsubscript{2} emission at 13:11–13:17 UT. The ripple fronts with \(~8-10\) wave crests recorded in both the OH and O\textsubscript{2} emissions were traveling with speeds of 28 and 29 ms\(^{-1}\) towards the west at \(~276°\) and \(~280°\), respectively, as measured clockwise from north. The horizontal wavelengths of this event were measured to be \(~10\) km and \(~12.5\) km, exhibiting an observed period of 6 and 7.2 min. The arrows indicate the direction of motion of the ripple.

Figure 5.3 shows ground maps of the July 7, 2002 ripple event observed at Haleakala, Maui (a) in an OH emission at 12:17 UT, and (b) in an O\textsubscript{2} emission at 12:20 UT. The projected ripple wave crests onto the map exhibited approximately an E-W orientation of the ripple event heading nearly towards the north. Figure 5.4 shows ground maps of the second ripple event observed at Haleakala, Maui in (a) the OH emission at 13:14 UT, and (b) in the O\textsubscript{2} emission at 12:17 UT on March 7, 2005. For this ripple event, the projected wave crests onto the map exhibited a \(~\text{-}N-S\) orientation heading nearly towards the east.

The distinct features, dynamic and convective instabilities, are determined

\begin{figure}[h]
\centering
\includegraphics[width=\textwidth]{figure5.3.png}
\caption{Ground maps showing orientation of the ripple event observed on July 7, 2002 at Haleakala, Maui for (a) the OH emission at 12:17 UT, and (b) the O\textsubscript{2} emission at 12:20 UT.}
\end{figure}
Figure 5.4. Ground maps showing orientation of the ripple event observed on March 8, 2005 at Haleakala, Maui for (a) the OH emission at 13:14 UT, and (b) the O₂ emission at 13:17 UT.

using the UIUC Na lidar wind-temperature datasets. By measuring vertical profiles of temperature and horizontal wind components smoothed to 100 m and 15 min. resolutions, the Richardson number, \( Ri \) can be computed at each altitude and used to assess Kelvin-Helmholtz instability in the mesopause region. Errors in the calculation of \( N^2 \) and \( Ri \) are primarily due to uncertainties in temperature and wind measurements. In order to reduce measurement uncertainties, the wind and temperature profiles need to be smoothed. However, stronger vertical and temporal smoothing can reduce the vertical gradients in temperature and wind. For the derivation of the temperature gradient, the Brunt-Väisälä frequency squared, the total wind shear and \( Ri \) values, smoothing of temperature and wind data using a running mean over 5 points between 80 and 100 km altitude were applied.

The properties of the instabilities can be quantified by considering the squared of the Brunt-Väisälä frequency, \( N^2 \) and the gradient Richardson number, \( Ri \) [Richardson, 1920; Gossard and Hooke, 1975]. The Brunt-Väisälä frequency, \( N \) is defined as

\[
N^2(z) = \frac{g}{T(z)} \left( \frac{dT(z)}{dz} - \frac{dT_d}{dz} \right),
\]

where \( g \), \( T(z) \), \( \frac{dT(z)}{dz} \) and \( \frac{dT_d}{dz} \) are acceleration due to gravity, atmospheric temperature,
temperature gradient, and adiabatic lapse rate. In the mesopause region, adiabatic lapse rate and $g$ are taken to be $-9.5 \text{ K km}^{-1}$ and $9.5 \times 10^{-3} \text{ km s}^{-2}$. The local atmosphere is to be convectively unstable if $N^2 < 0$ where the atmospheric lapse rate is larger than the adiabatic lapse rate. Under this condition, an air parcel displaced vertically will tend to continue moving away from its equilibrium position, resulting in small perturbations, which can be amplified in an unstable atmosphere.

The existence of vertical shear in horizontal wind disturbs atmospheric flow. Shear or Kelvin-Helmoltz instability is induced by large vertical shears of the horizontal winds with low static stability. Kelvin-Helmoltz instability can be measured by the gradient Richardson number, $Ri$, which is defined as

$$Ri = \frac{N^2}{\left(\frac{du_x}{dz}\right)^2 + \left(\frac{du_y}{dz}\right)^2},$$

where $u_x$ and $u_y$ are zonal (positive to east) and meridional (positive to north) winds. If the wind velocity vector, $V$, is the vector sum of the zonal and meridional winds, then vertical shear of the horizontal winds is given by $V = \sqrt{\left(\frac{du_x}{dz}\right)^2 + \left(\frac{du_y}{dz}\right)^2}$. Here the wind is measured as a function of altitude, and the direction of the shear vector is the vector difference of the winds at two different altitudes. The Brunt-Väisälä frequency squared, $N^2$, is a measure of the static stability of the atmosphere, while the dimensionless number, $Ri$, is a measure of the stability of the stratified shear flow.

There are two instability generations: (1) if $Ri$ is $0 < Ri < 0.25$ [Miles, 1961], dynamic instability occurs due to either the presence of a large wind shear or moderate wind shear accompanied by a large negative value of vertical temperature gradient ($< 9.5 \text{ K km}^{-1}$) to overpower the stabilizing buoyant forces, and (2) if $Ri = 0$, the convective instability occurs due to the presence of a superadiabatic temperature gradient [e.g., Fritts and Rastogi, 1985].
5.3 Results and Discussion

The observed ripple structures are usually regarded to exist at the height of the center of typical airglow emissions OH (∼87 km) and O$_2$ (∼94 km). However, recent satellite observations and analytical approach have clarified that there are significant variations in airglow peak emission heights [e.g., Shepherd et al., 1995; Yee et al., 1997; Melo et al., 2000; Nakamura et al., 2005]. Kubota et al. [1999] and Ejiri et al. [2002] determined the height of ripple structures observed in the all-sky airglow images and showed that the ripple heights are not at the typical airglow center height, but could be significantly higher or lower than the center height of the airglow emissions. By noting the height ranges in OH and O$_2$ emission layers as being 87 ± 3 km and 94 ± 3 km, we have determined instability signatures where shear vectors were aligned perpendicular to the phase fronts of ripple events observed on the nights of July 7, 2002 and March 8, 2005. Theoretical studies have shown that the phase fronts of K-H billows were typically perpendicular to the wind shear vector (speed and rotational shear) in the unstable layer [Gossard and Hooke, 1975].

5.3.1 July 7, 2002 event

The observed ripple event was characterized in terms of local instability with lidar measurements of horizontal wind and temperature on July 7, 2002 at 12:15 UT. The $N$-squared and Richardson number profiles were derived from the temperature and wind gradients at 12:15 UT when the ripples were observed in the two airglow layers. Figure 5.5 shows the horizontal winds and wind shears profiles of 100 m-15 min. vertical-temporal resolution. As shown in Figure 5.5a, the meridional wind was significantly changed in comparison to the zonal wind in order to create the wind shears and dynamical instabilities in both airglow layers. The total wind shears were measured from zonal and meridional wind shears profiles as shown in Figure 5.5b.
Figure 5.5. Plots show the vertical profiles of (a) zonal (blue-solid) and meridional (red-solid) winds, and (b) zonal (blue-solid) and meridional (red-solid) wind shears at 12:15 UT on July 7, 2002.

The wind shear measurements allowed for the estimation of the alignment of the ripple fronts relative to the wind shear direction and the dynamic instability regions at altitudes between 84 and 100 km. For this analysis, we estimated the measurement uncertainty as the standard deviation of the mean as 3.2 ms$^{-1}$ for horizontal wind and 1 ms$^{-1}$km$^{-1}$ for wind shear.

Figure 5.6 depicts the vertical shears at different altitudes derived from lidar wind measurements. The wind shear vectors at the altitudes of two airglow layers at 12:15 UT, July 7, 2002 were calculated from the change of the horizontal wind vectors with altitude displayed in Figure 5.6a. For the July 7 event, we have chosen the altitude ranges of 90.0–90.7 and 96.8–97.1 km in two airglow layers where the anti-aligned wind shear vectors were nearly orthogonal to the ripple fronts. The local directional wind shears of magnitudes 39.1 and 21 ms$^{-1}$km$^{-1}$ at 90–90.7 and 96.8–97.1 km, respectively, were heading nearly towards north and south. The results suggest that two airglow layers were coupled by anti-aligned wind shears where small-scale instability structures were generated. The characteristics of the small-scale structures, including their alignment with respect to the wind shear vectors
Figure 5.6. Plots show vertical wind shears at different altitudes between 84 and 100 km derived from the change of the horizontal wind vectors with altitude (a) at 12:15 UT, July 7, 2002, and (b) at 13:00 UT, March 8, 2005. The wind shear vectors are nearly perpendicular to the fronts of ripples at 90–90.7 km (red arrow) and 96.8–97.7 km (blue arrow) in Figure a, and at 86.5–86.7 km (red arrow) and 91.8–92.2 km (blue arrow) in Figure b.

were consistent with those of K-H billow fronts perpendicular to the direction of shears reported by Hecht et al. [2001b, 2005] and Li et al. [2005], suggesting the small-scale structure ripple was caused by the dynamical instability.

The line plots of temperature, vertical temperature gradient, Brunt-Väisälä frequency squared, and the Richardson number are displayed in Figure 5.7. The temperature profile in Figure 5.7a shows two higher temperature peaks of 187 K at 86.3 km and 198 K at 93.1 km. This profile also depicts two altitude regions exhibiting minimum mesopause temperatures of 168 and 172 K located at altitudes of 90.9 and 98.3 km below and above the peak of the inversion layer at 93.1 km. The valley-to-peak amplitude of inversion layer between 90.9 and 93.1 km was obtained as 30 K, providing dynamically unstable regions below the regions of cooling at 90.9 km and 98.3 km with temperature lapse rates less than the adiabatic lapse
Figure 5.7. Plots show the vertical profiles of (a) temperature, (b) temperature gradient, (c) Brunt-Väisälä frequency squared, $N^2$, and (d) Richardson number, $Ri$, measured at 12:15 UT on July 7, 2002. The vertical red lines in Figure c indicate the convective instability limit, $N^2$, and the vertical line in Figure d is the dynamic instability limit ($Ri = 0.25$).

This inversion layer typically displays a vertical width of 7 km. We have estimated the uncertainty of 0.7 K as the standard deviation of the mean of the temperature between 84 and 100 km. The uncertainties of $N^2$ and $\frac{dT(z)}{dz}$ due to error in temperature were $0.3 \times 10^4 \text{s}^{-2}$ and 0.6 K km$^{-1}$, respectively. The results of temperature gradient, stability parameter, total wind shear, shear direction, the Richardson number, $Ri$, and instability for the July 7, 2002 event at 12:15 UT are tabulated in the Table 5.2.
Table 5.2. Data summary of temperature gradient \( \frac{dT(z)}{dz} \), static stability \( N^2 \), total wind shear \( V \), shear direction as measured clockwise from north \( \phi \), the Richardson number \( Ri \), and instability determined on July 7, 2002 at 12:15 UT.

<table>
<thead>
<tr>
<th>Altitude (km)</th>
<th>( \frac{dT_a}{dz} ) (K/km)</th>
<th>( N^2 ) ( \times 10^{-4} )s(^{-2} )</th>
<th>( V ) (m/s/km)</th>
<th>( \phi ) (deg.)</th>
<th>( Ri )</th>
<th>Instability</th>
</tr>
</thead>
<tbody>
<tr>
<td>85.0-85.1</td>
<td>-3.6</td>
<td>3.02</td>
<td>36.8</td>
<td>55</td>
<td>0.22</td>
<td>Dynamic</td>
</tr>
<tr>
<td>86.8-87.3</td>
<td>-5.9</td>
<td>1.84</td>
<td>36.2</td>
<td>152</td>
<td>0.14</td>
<td>Dynamic</td>
</tr>
<tr>
<td>87.4-87.5</td>
<td>-8.3</td>
<td>0.61</td>
<td>19.4</td>
<td>215</td>
<td>0.16</td>
<td>Dynamic</td>
</tr>
<tr>
<td>88.4-88.7</td>
<td>-6.8</td>
<td>1.47</td>
<td>27.8</td>
<td>345</td>
<td>0.19</td>
<td>Dynamic</td>
</tr>
<tr>
<td>90.0-90.7</td>
<td>-5.1</td>
<td>2.44</td>
<td>39.1</td>
<td>4</td>
<td>0.16</td>
<td>Dynamic</td>
</tr>
<tr>
<td>95.7-96.0</td>
<td>-5.9</td>
<td>1.86</td>
<td>30.6</td>
<td>143</td>
<td>0.19</td>
<td>Dynamic</td>
</tr>
<tr>
<td>96.8-97.1</td>
<td>-8.9</td>
<td>0.33</td>
<td>21</td>
<td>197</td>
<td>0.09</td>
<td>Dynamic</td>
</tr>
</tbody>
</table>

In order to confirm the occurrence of K-H instability, we have derived \( N^2 \) from temperature profile using the equation (5.1). More detailed structures of \( \frac{dT(z)}{dz} \) and \( N^2 \) can be seen in Figures 5.7b and 5.7c where their vertical structures were significantly consistent. K-H instability is likely to occur when there is a wind shear of nominal value of 40 ms\(^{-1}\)km\(^{-1}\) with static stability, \( N^2 \), of \( 4 \times 10^{-4} \) s\(^{-2} \) in the mesosphere and lower thermosphere (MLT) region [Klostermeyer, 1980; Hirota et al., 1983], though the \( N^2 \) depends on the temperature gradient.

It is noted that the moderate wind shear below 40 ms\(^{-1}\)km\(^{-1}\) for the small positive value of \( N^2 \) associated with a relatively large negative value of temperature gradient (\(< 9.5 \) Kkm\(^{-1}\)) was sufficient to generate K-H instability. The large wind shear of 39.1 ms\(^{-1}\)km\(^{-1}\) with \( N^2 \) of \( 2.44 \times 10^{-4} \) s\(^{-2} \) and a temperature gradient of \(-5.1 \) Kkm\(^{-1}\) was enough to provide the \( Ri \) value less than 0.25, leading to K-H instability at 90–90.7 km. Also the moderate wind shear of 21 ms\(^{-1}\)km\(^{-1}\) with \( N^2 \) of \( 0.33 \times 10^{-4} \) s\(^{-2} \) and a temperature gradient of \(-8.9 \) Kkm\(^{-1}\) was able to lower the \( Ri \) value to less than 0.25, again leading to K-H instability of 96.8–97.1 km. The profile of the \( Ri \) number derived from the total vertical wind shear and the \( N^2 \) data using equation (5.2) is shown in Figure 5.7d. The combination of vertical temperature
gradients and wind shears exhibited $R_i$ numbers of 0.10 and 0.16 at 90–90.7 and 96.8–97.1 km altitudes, respectively.

### 5.3.2 March 8, 2005 event

To further characterize the ripple and associated instabilities, the vertical temperature gradient and wind gradient profiles were derived from the temperature and horizontal wind profiles on the night of March 8, 2005 at 13:00 UT. Then $N$-squared and the Richardson number were calculated from the temperature and wind gradient profiles at the time ripples occurred in two airglow layers. Figure 5.8 shows the zonal and meridional winds and wind shears profiles of 100 m-15 min. vertical-temporal resolutions. As shown in Figure 5.8a, there were sharp changes in zonal and meridional winds in speed and direction in both airglow layers. The total wind shears were calculated from zonal and meridian wind shear profiles as shown in Figure 5.8b. The key feature of the horizontal winds was the presence of large wind shears between 91.8 and 93.4 km, mainly due to zonal wind shear that produced K-H instability. The wind shear measurements between 83 and 100 km allowed estimating the alignment of the ripple phase fronts with respect to the wind shear direction. The maximum allowable measurement uncertainties as the standard deviation of the mean between 83 and 100 km was about 4.2 ms$^{-1}$ and 3.1 ms$^{-1}$ for zonal and meridional winds, respectively. The uncertainty in total wind shear between 83 and 100 km was estimated at 1.2 ms$^{-1}$ km$^{-1}$.

The wind shear vectors computed from the change of the horizontal wind vectors with altitude at 13:00 UT, March 8, 2005 are shown in Figure 5.6b. Since shear vectors were rotated, two oppositely aligned local wind shears (speed and rotational) of magnitudes $\sim 21$ ms$^{-1}$ km$^{-1}$ at 86.6–86.7 km and $\sim 52.2$ ms$^{-1}$ km$^{-1}$ at 91.8–92.2 km, were nearly orthogonal to the ripple fronts, which were again consistent with
Figure 5.8. Plots show the vertical profiles of (a) zonal (blue-solid) and meridional (red-solid) winds, and (b) zonal (blue-solid) and meridional (red-solid) wind shears at 13:00 UT on March 8, 2005.

those of K-H billow fronts perpendicular to the direction of shear [Hecht et al., 2001b, 2005; Li et al., 2005], suggesting the ripple was caused by dynamical instability. The wind shear of magnitude $\sim 21 \text{ ms}^{-1}\text{km}^{-1}$ was heading towards the east at $285^\circ$, while the wind shear magnitude of $\sim 52.2 \text{ ms}^{-1}\text{km}^{-1}$ was heading towards the west at $112^\circ$ as measured clockwise from north. These anti-aligned wind shear vectors coupled two airglow layers where small-scale instability features were generated.

The line plots displayed in Figure 5.9 show the detailed structures of temperature, vertical temperature gradient, Brunt-Väisälä frequency squared, and the Richardson number. The temperature profile in Figure 5.9a shows two peaks of maximum temperatures of 207 and 223 K at 84.7 and 92.2 km, which then again decreased to minimum mesopause temperatures of 185 and 184 K located at altitudes of 89.2 and 97.6 km. The valley-to-peak amplitudes of lower and upper inversion layers were $\sim 22$ and $\sim 39$ K with dynamical (and convective) unstable regions below the cooling regions at 89.2 and 97.6 km, respectively. The upper and lower inversion layers typically display the vertical widths of $\sim 8$ and $\sim 6$ km, respectively. The measurement uncertainty as the standard deviation of the mean was 1 K for the
temperature between 83 and 100 km. The measurement uncertainties of $N^2$ and temperature gradient, due to error in temperature, were $0.4 \times 10^{-4}$ s$^{-2}$ and 0.9 K km$^{-1}$.

The results of the temperature gradient, static stability parameter, total wind shear, shear direction, and the Richardson number, and instability for the March 8, 2005 event at 13:00 UT are tabulated in the Table 5.3.

We have derived $\frac{dT(z)}{dz}$ and $N^2$ from the temperature profile using equation (5.1) in order to determine the possible occurrence of K-H instability. More detailed
Table 5.3. Data summary of temperature gradient \( \frac{dT(z)}{dz} \), static stability \( N^2 \), total wind shear \( V \), shear direction as measured clockwise from north \( \phi \), the Richardson number \( Ri \), and instability determined on March 8, 2005 at 13:00 UT.

<table>
<thead>
<tr>
<th>Altitude (km)</th>
<th>( \frac{dT}{dz} ) (K/km)</th>
<th>( N^2 ) ( (10^{-4}s^{-2}) )</th>
<th>( V ) (m/s/km)</th>
<th>( \phi ) (deg.)</th>
<th>( Ri )</th>
<th>Instability</th>
</tr>
</thead>
<tbody>
<tr>
<td>85.0-85.4</td>
<td>-5.7</td>
<td>1.77</td>
<td>32</td>
<td>318</td>
<td>0.17</td>
<td>Dynamic</td>
</tr>
<tr>
<td>86.5-86.7</td>
<td>-7.5</td>
<td>1.01</td>
<td>21</td>
<td>285</td>
<td>0.23</td>
<td>Dynamic</td>
</tr>
<tr>
<td>87.3-87.5</td>
<td>-10.4</td>
<td>-0.46</td>
<td>17.8</td>
<td>319</td>
<td>-0.14</td>
<td>Convective</td>
</tr>
<tr>
<td>91.8-92.2</td>
<td>5.0</td>
<td>6.19</td>
<td>52.2</td>
<td>112</td>
<td>0.22</td>
<td>Dynamic</td>
</tr>
<tr>
<td>92.3-93.0</td>
<td>-2.3</td>
<td>3.08</td>
<td>41.3</td>
<td>125</td>
<td>0.18</td>
<td>Dynamic</td>
</tr>
<tr>
<td>93.1-94.4</td>
<td>-7.8</td>
<td>0.72</td>
<td>32.7</td>
<td>139</td>
<td>0.07</td>
<td>Dynamic</td>
</tr>
<tr>
<td>94.5-94.9</td>
<td>-9.9</td>
<td>-0.16</td>
<td>24.2</td>
<td>145</td>
<td>-0.03</td>
<td>Convective</td>
</tr>
<tr>
<td>95.0-95.6</td>
<td>-8.8</td>
<td>0.33</td>
<td>15.9</td>
<td>143</td>
<td>0.12</td>
<td>Dynamic</td>
</tr>
<tr>
<td>95.7-96.3</td>
<td>-11.5</td>
<td>-0.98</td>
<td>4.4</td>
<td>165</td>
<td>-8.97</td>
<td>Convective</td>
</tr>
</tbody>
</table>

Structures of temperature gradient and \( N^2 \) are plotted in Figures 5.9b and 5.9c, where their vertical structures are seen to be significantly consistent. The magnitude of \( N^2 \) changes according to the change in the temperature gradient with altitude, so that the dynamically unstable regions occurred with wind shears below the 40 ms\(^{-1}\)km\(^{-1}\) based on the negative value of temperature gradient \((<9.5 \text{ K km}^{-1})\). The moderate wind shear of \( \sim 21 \text{ ms}^{-1}\text{km}^{-1} \) for the corresponding small positive \( N^2 \) of \( 1.01 \times 10^{-4} \text{ s}^{-2} \) with a temperature gradient of \(-7.5 \text{ K km}^{-1}\) at 86.5–86.7 km was sufficient to produce an \( Ri \) value less than 0.25, leading to Kelvin-Helmholtz instability. The strong wind shear of \( \sim 52.5 \text{ ms}^{-1}\text{km}^{-1} \) with \( N^2 \) of \( 6.2 \times 10^{-4} \text{ s}^{-2} \) and a temperature lapse rate of 5 K km\(^{-1}\) produced a dynamically unstable region at 91.8–92.2 km.

The results suggest that not only the strong wind shear but also the moderate wind shear less than 40 ms\(^{-1}\)km\(^{-1}\) with small positive \( N^2 \) leads to dynamic instability. The profile of the Richardson number, \( Ri \), derived from \( N^2 \) and total vertical wind shear profiles using equation (5.2) is shown in Figure 5.9d. The \( Ri \) values of 0.23 and 0.22 were derived at 86.5–86.7 and 91.8–92.2 km due to the presence of moderate and strong wind shears, leading to K-H instabilities.
Because of the superadiabatic lapse rate that exceeds the adiabatic lapse rate (> 9.5 K km\(^{-1}\)) in the MLT region, convective instabilities are likely to occur with \(Ri\) values less than zero. It is found that there were convectively unstable layers at 87.3–87.5 and 94.5–94.9 km between the dynamic instability layers indicating \(Ri\) values of −0.14 and −0.03. By averaging the temperature profile of 100 m-15 min. vertical and temporal resolution, the uncertainty in temperature can be reduced, so that the negative value of vertical gradient of temperature can be decreased and in turn convective instability region may become stable with larger value of \(N^2\).

5.3.3 Numerical model simulation

Numerical simulation was employed to examine apparent simultaneous occurrences of Kelvin-Helmholtz (K-H) instability due to dual shear in two airglow layers. The incompressible Navier-Stokes equations subject to the Boussinesq approximation were utilized for direct numerical simulations (DNS) of Kelvin-Helmholtz shear instability growth [Werne and Fritts, 2001; Fritts et al., 2009a,b]. The Navier-Stokes equations are given in Appendix B. This simulation was performed using the code in deep domain studies instead of very shallow domains [Laughman et al., 2009]. This work was a collaborative effort with Dave C. Fritts and Brian Laughman [2009] of NorthWest Research Associates, CoRA Division.

The goal of these modeling efforts is to reproduce the case study of the July 7, 2002 event with the Kelvin-Helmholtz instability structure with observed horizontal speed of \(\sim 27\) ms\(^{-1}\), wavelength of \(\sim 9\) km and propagation direction of \(\sim 6^\circ\) towards the north. Additionally, the numerical studies presented two plausible coupling mechanisms for this event. The vertical resolution of the lidar wind and temperature data along the 30\(^\circ\) off-zenith line of sight (LOS) was 417 m. Modeled wind profiles include the observed, smoothed and extended off-zenith LOS wind profile and
small vertical scale perturbations to this observed profile. The potential temperature profile used in the model was based on the observed, smoothed and extended temperature profile; the static stability \( N^2 \) vertical profile was computed from the temperature data using equation (5.1) and then smoothed. The mean potential temperature profile is then computed from the smoothed \( N^2 \) profile according to

\[
\theta(z) = \frac{\bar{\theta}}{g} \int N^2(z)dz. \tag{5.3}
\]

The potential temperature is the anti-derivative of the static stability \( N^2 \), so

\[
\theta(z) = \frac{\bar{\theta}}{g} [0.0001z + (0.0017)(0.8) \tanh((z - 91.5)/0.8)], \tag{5.4}
\]

where \( \bar{\theta} \) is the vertically averaged value of an assumed, idealized, potential temperature and \( g \) is the acceleration of gravity. The analytical form of the small-scale perturbation added to the observed mean wind is given as

\[
u'(z) = \frac{18}{\pi} \sin[2\pi(z - 97)] \times \frac{1}{2} [\tanh(z - 97 + 2.5) - \tanh(z - 97 - 2.5)]. \tag{5.5}
\]

Figure 5.10 shows the vertical profiles of the static stability \( N^2 \), mean wind with the addition of the small-scale perturbation \([U(z)]\), and the resulting Richardson number \((Ri)\) at the time \(\sim12:18\) UT of the ripple occurrence. \( N^2 \) units are \(s^{-2}\), \(U(z)\) units are \(\text{ms}^{-1}\), and the altitude is plotted in units of km. The small-scale perturbation changes the Richardson number at many more altitudes than the original unperturbed profile. As shown in Figure 5.10c, the Richardson numbers at altitudes \(\sim89.0\) and \(\sim96.0\) km were \(>0.25\) and \(<0.25\) with corresponding wind shears of 32.0 and \(-26.3\) \(\text{ms}^{-1}\)\(\text{km}^{-1}\), indicating two oppositely aligned K-H shears at two layers.

Figure 5.11 demonstrates the first coupling mechanism. K-H billows arise at the upper shear and produce an enhanced \(N^2\) response in the nearby high stable layer. Figure 5.11 displays time evolution of the perturbation potential temperature
Figure 5.10. Plots show the vertical profiles of (a) computed Brunt-Väisälä frequency squared, \( N^2 \), (b) smoothed and extended mean wind created by the addition of the small-scale perturbation, \( U(z) \), and (c) Richardson number, \( R_i \), at \(~12:18\) UT for July 7, 2002.

Field for the observed wind and temperature profiles in Figure 5.10. Vertical and horizontal axes are plotted in units of km and the time is in minute. Panels (a) and (f) show an early time in the evolution and the end of evolution, and panels (c) to (e) are an intermediate time in the evolution. As shown in Figure 5.11, the K-H shear instabilities with a shear of \(-26.3\) ms\(^{-1}\)km\(^{-1}\) appeared at the upper layer \((\sim 96\) km\). The model results show that an adjacent \(N\)-squared layer \((\sim 91\) km\) was able to enhance the response produced by upper shear K-H instability to give the appearance of a dual shear response. The response in the lower layer is the effect of the enhanced sensitivity of the strongly stratified region to vertical displacements, while the vertical displacements are caused by the upper shear layer. We note that throughout the evolution the response in the \(N\)-squared duct was strongly locked to the higher altitude K-H billows.

In order to confirm the possibility of the second coupling mechanism, the dual shear response, we have used a different perturbation to the observed wind data to enhance the shears, as well as the idealized \(N^2\) profile with the function given as

\[
N^2(z) = 0.0002 + (0.0023) \times (1 - \text{tanh}^2[(z - 91.5)/0.55]),
\] (5.6)
Figure 5.11. Model simulation results depicting time evolution of the perturbation potential temperature field for the observed wind and temperature profiles with a small-scale perturbation added. In all panels, the vertical and horizontal data are plotted in units of km and time is in min.

where 0.0002 s$^{-2}$ and 0.0023 s$^{-2}$ are our chosen values of background and peak stability. The corresponding potential temperature is given as

$$\theta(z) = \frac{\bar{\theta}}{g}[0.0002z + (0.0023) \tanh((z-91.5)/0.55)].$$  \hfill (5.7)

The analytical form of the large perturbation added to the mean wind is given as

$$u'(z) = 0.6 \sin[2\pi(z-89.8)/3] \times \frac{1}{2} [\tanh(z-90+1.3) - \tanh(z-90-1.3)]$$

$$+1.0 \sin[2\pi(z-98.8)/2] \times \frac{1}{2} [\tanh(z-98.8+1.3) - \tanh(z-98.8-1.3)].$$  \hfill (5.8)
Figure 5.12 shows the vertical profiles of idealized $N$-squared, mean wind with the addition of a different, large amplitude perturbation [$U(z)$], and Richardson number ($Ri$) at the time $\sim$12:18 UT of the ripple occurrence. $N$-squared units are $s^{-2}$, $U(z)$ units are $ms^{-1}$, and the altitude is plotted in units of km. As shown in Figure 5.12c, the Richardson numbers at altitudes 89.6 and 98.7 km were 0.073 and 0.043 with corresponding wind shears of 56.77 and $-56.78 ms^{-1} km^{-1}$, indicating two oppositely aligned K-H shears at two layers. The upper K-H shear billow appeared at 98.7 km is less likely explanation for the observed K-H shear layer at $\sim$97.0 km. But it may be plausible to occur upper billow at 98.7 km due to the depth of the O$_2$ and OH layers ($\sim$8–10 km). The horizontal scales and time scales at both altitudes supports the interpretation that dual shear event is responsible for the observed airglow data.

This result involves two model runs with two related but different initial datasets for the profiles depicted in Figure 5.12. The first has the wind profile depicted in Figure 5.12b while the second has the same wind profile, but with the lower shear zeroed and the wind, $U(z)$, set to the constant value of 55 $ms^{-1}$ in the

![Figure 5.12](image-url)

**Figure 5.12.** Plots show the vertical profiles of (a) idealized Brunt-Väisälä frequency squared, $N^2 (s^{-2})$, (b) smoothed and extended mean wind created by the addition of the large amplitude perturbation, $U(z) (ms^{-1})$, and (c) Richardson number, $Ri$, at $\sim$12:18 UT for July 7, 2002.
lower half of the domain. Figure 5.13 shows the evolution of a potential temperature field seeded with random noise for the profiles depicted in Figure 5.12 to produce panels (a)–(c) without turning off the lower shear, and to produce panels (d)–(f) with constant mean horizontal velocity (zero shear) in the lower half of the domain. The times for both cases were chosen to align as closely as possible. The reason for turning off the lower shear was to assess its impact on the upper shear’s evolution. The panels (a) and (d) were taken at an early time in the evolution and panels (c)

Figure 5.13. Model simulation results depicting time evolution of the perturbed potential temperature field for the large amplitude perturbed, half-shear, observed wind and idealized temperature profiles. In all panels, the vertical and horizontal data are plotted in units of km and time is in min.
to (e) were taken at an intermediate time in the evolution, while the panels (c) and (f) were taken for the end of the evolution. Here the upper shear was influenced by the presence of lower shear.

5.4 Summary and Conclusions

Simultaneous measurements from the airglow imager and Na wind-temperature lidar co-located at Maui, Hawaii provide unique datasets to investigate the spatial and temporal characteristics of small-scale ripples and associated instabilities. From the successive images in the OH and O$_2$ airglow emissions, we have determined the horizontal wavelengths to be $\sim$7–14 km, and horizontal speeds to be $\sim$25–36 ms$^{-1}$, as well as the apparent period of $\sim$5–8 min. for 15 events on 10 nights of coincident lidar wind-temperature measurements. For case studies, we have reported two short-lived, small-scale K-H instability structures, simultaneously observed in OH and O$_2$ airglow emissions on the nights of July 7, 2002 and March 8, 2005. The coincident Na lidar measurements of the mesopause region temperature and horizontal winds allowed us to characterize the nature of the ripples, as well as to investigate the causal relations between ripples and dynamic instabilities. We found that the ripple events observed on both nights were induced by simultaneous occurrences of K-H instabilities due to dual shears in two airglow layers. Specifically, the dual wind shear vectors at two airglow layers for both events were aligned oppositely where small-scale ripples were observed. The wind shear vectors reveal that the speed and directional shears associated with the measured K-H instabilities were nearly perpendicular to the fronts of the observed ripples. Model results indicate two coupling mechanisms between adjacent airglow layers for the observed K-H instability structures occurring on the night of July 7, 2002. The first mechanism involves K-H billow that develops at the upper shear layer and subsequently forces a response in
the nearby region of high stability. The second mechanism involves two sets of K-H billows that develop at one altitude are able to influence the evolution of the billows at the other altitude.

We have also computed the values of static stability parameter, $N^2$, and instability parameter, $R_i$, using Na lidar temperature and horizontal wind profiles. The atmosphere is considered to be dynamically unstable when $R_i$ lies between 0 and 0.25. Dynamic instability generally occurs when there is a strong wind shear and small static stability, $N^2$. However, the results of correlation between wind shear and static stability, $N^2$, which depends on a negative value of temperature gradient ($< 9.5 \text{ K km}^{-1}$), show that strong wind shear is not always necessary to lead to dynamically unstable conditions. However, the moderate wind shear ($< 40 \text{ ms}^{-1} \text{ km}^{-1}$) associated with small static stability, $N^2$, and a large negative value of temperature gradient ($< 9.5 \text{ K km}^{-1}$) can cause the $R_i$ number to be less than 0.25, leading to dynamic instability. For both events, $R_i$ values less than 0.25 reveal that dynamically unstable regions occurred in two airglow layers due to a combination of moderate wind shears ($< 40 \text{ ms}^{-1} \text{ km}^{-1}$) and small static stability, $N^2$, associated with a large negative value of temperature gradient ($< 9.5 \text{ K km}^{-1}$), as well as large wind shears ($> 40 \text{ ms}^{-1} \text{ km}^{-1}$) accompanied by small static stability, $N^2$, associated with a small positive value of temperature gradient.
6.1 Introduction

Gravity waves in the atmosphere provide significant and important dynamic coupling between horizontally and vertically separated atmospheric regions. Even under ideal windless and inviscid conditions, upward propagation through exponentially decreasing atmospheric density will result in exponential increase in wave velocity perturbations, conserving kinetic energy with altitude [e.g., Gossard and Hooke, 1975], and leading ultimately to breaking and deposition of momentum and energy [e.g., Fritts and Alexander, 2003, and references cited therein]. Understanding the local propagation characteristics of gravity waves is very important, because under certain conditions, gravity waves are able to transport energy and momentum vertically, and under other conditions they can be trapped (ducted), confining the major flow of wave energy and momentum to a limited range of altitude, and allowing long-range horizontal propagation [e.g., Walterscheid et al., 2000].

Upward propagating gravity waves in the atmosphere are highly influenced by thermal structure and background winds. Atmospheric gravity waves may be subject to ducting in a stably stratified atmosphere that contains levels of temperature or wind maxima or minima. Gravity waves propagate in the vertical direction when the vertical wave number is real and the magnitude of intrinsic frequency is less than the Brunt-Väisälä frequency, $N$, and they become evanescent when the vertical wave number is imaginary and the intrinsic frequency exceeds the Brunt-Väisälä frequency, $N$ [Chimonas and Hines, 1986]. Ducting of gravity waves can occur in a region of

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propagation sandwiched between two regions of evanescence. The background winds Doppler-shift intrinsic frequency as they vary in the direction of wave propagation. Ducting that is facilitated by wind flow is commonly referred to as Doppler ducting [Chimonas and Hines, 1986; Isler et al., 1997; Snively et al., 2007]. Ducting due to temperature gradients (leading to variation in Brunt-Väisälä frequency) over altitude is called thermal ducting [e.g., Walterscheid et al., 2001; Snively and Pasko, 2008, and references cited therein]. In the case of thermal ducting, the region of propagation occurs in the vicinity of a local maximum in the Brunt-Väisälä frequency, surrounded by regions of relatively low static stability, or even instability.

There has been considerable theoretical research on wave ducting mechanisms in the upper atmosphere [e.g., Chimonas and Hines, 1986; Fritts and Yuan, 1989a; Walterscheid et al., 2001; Snively and Pasko, 2008, and references cited therein]. Some of these studies considered the existence of ducts caused by thermal gradients or variations in the background winds causing Doppler ducts. Observations in the airglow layers, and numerical model simulations, confirm that atmospheric gravity waves may be thermally ducted [Walterscheid et al., 1999, 2001; Hecht et al., 2001a; Yu and Hickey, 2007; Snively and Pasko, 2003, 2008] and/or Doppler-ducted [Isler et al., 1997; Hickey, 2001; Snively et al., 2007]. Airglow imaging provides an important tool for determining horizontal characteristics and the occurrence of small-scale, short-period waves at the emission altitudes [Taylor and Edwards, 1991; Taylor et al., 1995d, 1997]. It has been noted that the observed phase speed and wave period may differ significantly from the intrinsic phase speed and wave period due to the presence of background winds. When airglow image data are combined with simultaneous wind measurements, intrinsic wave parameters can be investigated over a range of heights (80–100 km), permitting a study of the propagation and ducting characteristics of the wave motions and a more precise comparison with theoretical
predictions. A study by Isler et al. [1997], using simultaneous imager observations of gravity waves and MF radar wind data obtained at Maui from 80 and 100 km altitude suggested that Doppler wave ducts commonly form due to the variation of the background mesospheric wind, such as that arising from the local tidal fields [Snively et al., 2007].

As part of Maui-MALT program, we have utilized novel airglow image data from Maui, HI to investigate short-period, small-scale, gravity wave parameters using sequential measurements in the OH and O\textsubscript{2} emission layers in the mesopause region. Coincident meteor radar wind datasets have been used to investigate the intrinsic properties of the wave events and to assess their ducting or evanescence characteristics. Here we present in particular the observed ducted and evanescent wave events. We focus on one example Case Study of Doppler ducting, and another example Case Study of evanescence, induced by the local wind structure. The ducting was associated with maxima and minima of the background wind speed as measured in the direction of wave propagation, and evanescence was associated with the maxima of background wind in the opposite direction of wave motion. We describe the data and analysis in Section 6.2, including summaries of ducted and evanescent wave observations in tabular form. Results and discussion pertaining to the analyses of two observed wave events, and numerical model case studies, are presented in Section 6.3. These are followed by summary and conclusions in Section 6.4.

6.2 Data and Analysis

Airglow imaging measurements provide unique datasets for investigating characteristics of observed small-scale ducted and evanescent waves simultaneously in OH and O\textsubscript{2} airglow layers in the mesopause region. Each image of a circular 90° field of view was flat-fielded in order to correct the non-uniform intensity in the image.
Next, the flat-fielded images were calibrated using the background star-field in order to get a geographic orientation (north on top and south on bottom). The calibrated flat-fielded images were then unwarped by projecting onto a rectangular grid of size $180 \text{ km} \times 180 \text{ km}$ [Garcia et al., 1997; Pautet and Moreels, 2002]. To measure the propagation direction and the horizontal wavelength of waves, a 3-D Fast Fourier Transform (FFT) spectral analysis was used to process unambiguous spectrum for a series of pictures [Garcia et al., 1997; Coble et al., 1998]. To determine the phase speeds of the observed waves, the displacement of a selected wave front from one image to the next was measured through a series of images using a line drawing method. The observed periods were then calculated from measured wavelengths and phase speeds. We have estimated lifetimes of the wave events by observing the sequence of OH and O$_2$ images in $\sim$5.5-minute intervals of time. The University of Illinois meteor radar provided continuous hourly wind measurements over the altitude range of 80–100 km, and was used to investigate the Doppler ducting and evanescence of the observed waves.

Vertical structure of waves can be estimated under the Boussinesq approximation, using the Taylor-Goldstein equation from the linear, irrotational and inviscid equations of motion in an incompressible and stratified atmosphere [Nappo, 2002]. The Taylor-Goldstein equation for the vertical perturbation velocity $w_z$ is

$$\frac{d^2 w_z}{dz^2} + \left[ \frac{N^2}{(c - u_0)^2} + \frac{1}{(c - u_0)} \frac{d^2 u_0}{dz^2} - \frac{1}{H(c - u_0)} \frac{du_0}{dz} - k^2 - \frac{1}{4H^2} \right] w_z = 0, \quad (6.1)$$

where $N$ is the Brunt-Väisälä frequency, $c$ is observed phase speed, $u_0$ is the background wind in the direction of wave propagation; $\dot{\omega} = \omega - u_0 k$ is the intrinsic frequency of wave, $k = 2\pi/\lambda_x$ is the horizontal wave number and $H$ is the scale height. After neglecting the curvature and vertical shear terms, the local vertical wave number squared $m^2$ for the Taylor-Goldstein Equation (6.1) is given by
\[ m^2 = \frac{N^2}{(c - u_0)^2} - k^2 - \frac{1}{4H^2}. \]  
\[ (6.2) \]

Here both \( N^2 \) and \( u_0 \) vary smoothly and gradually with altitude. By computing the local profiles of \( m^2 \), the boundary conditions of ducting regions can be inferred [Isler et al., 1997; Snively et al., 2007]. A wave is vertically propagating where \( m^2 > 0 \) and evanescent where \( m^2 < 0 \), and the wave can be ducted in a region of vertical propagation \((m^2 > 0)\) bounded by regions of evanescence \((m^2 < 0)\). Alternating regions of evanescence and propagation may arise due to variation in the wind \([u_0(z)]\) or Brunt-Väisälä frequency \([N(z)]\). In our analysis here, we have computed \( m^2 \) from Equation (6.2) using airglow image data to measure ground-relative wave characteristics, and meteor radar data to provide background wind in the direction of wave propagation.

Brunt-Väisälä frequency \([N(z)]\) as function of altitude is obtained from the MSISE-90 atmospheric model [Hedin, 1991] and scale height \((H)\) is taken to be 6 km in the mesopause region. It is here assumed that the variations in wind field \((u_0)\) will dominate over variations in \( N \), leading to the formation of Doppler ducts [Isler et al., 1997]. This is not always the case in the presence of strong temperature inversion layers [e.g., Chu et al., 2005], which lead to regions of high stability that can form thermal ducts. For the cases modeled here, MTM data is used to confirm relatively weak temperature gradients, suggesting that significant inversions were not likely present at times of these events, and that MSISE-90 data will provide a good approximation of local thermal conditions. The approximate local Brunt-Väisälä frequency squared can be determined from temperature as

\[ N^2 = \frac{g}{T} \left( \frac{dT}{dz} + \frac{g}{c_p} \right), \]
\[ (6.3) \]

where \( g \) is the acceleration of gravity, \( c_p \) is the specific heat, \( T \) is temperature and \( \frac{dT}{dz} \) is the slope between temperature data points measured from OH and \( O_2 \) airglow.
layers by the MTM [Nappo, 2002]. This simple analysis is insufficient to eliminate
the possibility of thermal ducting, but it does provide insight into the conditions
present within the airglow region, confirming that steep gradients are unlikely to
exist. Although temperatures measured do not agree entirely with MSISE-90 data,
the similar $N^2$ values and temperature gradients suggest that it serves as a reasonable
approximation.

The background wind in the direction of wave propagation [$u_0(z)$], as mea-
sured clockwise from north, is calculated from meteor radar data as

$$u_0(z) = u_x(z) \sin \theta + u_y(z) \cos \theta,$$

(6.4)

where $u_x$ and $u_y$ are zonal and meridional components of observed wind profile,
respectively. The calculated wind $u_0(z)$ is then used to determine wave characteris-
tics, as shown in Figures 6.2a and 6.4a, and summarized in Table 6.1. The numerical
model requires use of a smoothed wind field [hence $U_0(z)$], which is derived from fit-
ting an analytical function to the observed wind profiles $u_0(z)$. For the analytical
modeled wind structure, we fit a function to approximate a tidal wind field as

$$U_0(z) = U_{bg} + \hat{U} \exp \left( \frac{z-z_0}{2H} \right) \cos \left[ K_z (z-z_0) \right],$$

(6.5)

where $U_{bg}$ is the unperturbed mean background wind, $\hat{U}$ is large-scale background
wave’s horizontal wind perturbation amplitude at the reference height ($z_0$), and ver-
tical wave number $K_z$ is taken to correspond to a vertical wavelength $\Lambda_z = 2\pi/K_z$.
The parameters are obtained by fitting, such that the peak of the wind agrees with
measured wind at reference height $z_0$, such that $U_0(z_0) = U_{bg} + \hat{U}$, where wavelength
is obtained by direct measurement of the wind profile. The analytical wind model
is then used to derive the Taylor-Goldstein vertical wave number squared ($m^2$) for
small-scale waves from Equation (6.2), to compare with that derived from the mea-
sured wind field, and to confirm agreement. The measurements errors for zonal and
meridional winds are calculated as the standard deviation and then measurement uncertainties are estimated as the standard deviation of the mean in the data analysis. The functions obtained are used as a basis for the numerical model background wind field, where they are enveloped within a secant-squared function to limit growth of the analytical wind profile above mesopause. This envelope function is given by $\text{sech}^2[(z - z_o)/20 \text{ km}]$, leading to approximately zero wind at upper and lower model domain boundaries.

We have examined the ducting and evanescence characteristics of small-scale waves, as determined by the meteor radar measurements of hourly background wind along the direction of wave propagation for each event. A total of 15 small-scale wave events, nine events were Doppler ducted and six events were evanescent. The waves investigated simultaneously in OH and O$_2$ emissions during 2003–2004 exhibited horizontal wavelengths $\lambda_x \approx 15–20 \pm 1 \text{ km}$, observed phase speeds $c \approx 25–50 \pm 3 \text{ ms}^{-1}$, observed periods $\tau \approx 7–12 \pm 1 \text{ min}$. We have presented one Doppler ducted event and one evanescent event, each observed simultaneously in OH and O$_2$ airglow emissions shown in Figure 6.1. The average measurement errors calculated as standard deviation for the horizontal wavelength, phase speed, propagation direction and observed period were 1 km, 3 ms$^{-1}$, 4$^\circ$ and 1 min., respectively. Here we have presented the results of the background winds in the direction of wave propagation as measured by the meteor radar, intrinsic phase speeds, intrinsic periods and Doppler-shifted intrinsic frequencies of two wave events. Table 6.1 lists the observed horizontal wavelength, wave propagation direction, observed horizontal phase speed and observed period of the wave events recorded in both OH and O$_2$ emissions.

It is important to note that none of the wave events examined here have characteristics of mesospheric bore events [e.g., Taylor et al., 1995d; Dewan and Picard, 1998], which are also likely consistent with strongly-ducted waves, although
Figure 6.1. Flat-fielded and unwarped airglow images obtained at Maui, HI, for the two case study events. Panes (a) and (b) depict OH and O$_2$ emissions, respectively, of Doppler-ducted wave observed on July 5, 2003. Panes (c) and (d) depict OH and O$_2$ emissions, respectively, of evanescent wave observed on November 30, 2003.

exhibiting distinctive dispersion characteristics and front-like structure. Many waves reported here appear with relatively gradual onset and varying magnitudes over the duration of the event. They occur under various wind conditions, although often following transitions of background wind fields that may contribute to the upward propagation of the waves from lower altitudes. We examine, in particular, two cases. First, where the wave is ducted near mesopause, although able to tunnel upward into the duct, due to relatively weak evanescence at the lower boundary. Second, where the wave is evanescent throughout the mesopause region, although able to tunnel into this region from above or below [e.g., Sutherland and Yewchuk, 2004].
Table 6.1. Data summary of horizontal wave parameters, as determined from the image data, including date of occurrence, duration, horizontal wavelength ($\lambda_x$), direction of propagation ($\theta$), phase speed ($c$), observed period ($\tau$) and ducted or evanescent status for the 80–100 km altitude range.

<table>
<thead>
<tr>
<th>Date</th>
<th>Time (UT)</th>
<th>Imager</th>
<th>$\lambda_x$ ±1 km</th>
<th>$\theta$ ±4°</th>
<th>$c$ ±3 m/s</th>
<th>$\tau$ ±1 min.</th>
<th>Ducted/ Evanescent</th>
</tr>
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<td>OH</td>
<td>20</td>
<td>46</td>
<td>50</td>
<td>6.7</td>
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<td></td>
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<td>50</td>
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<td>47</td>
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</tr>
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<td>208</td>
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<td>46</td>
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<td>232</td>
<td>41</td>
<td>7.7</td>
<td></td>
</tr>
</tbody>
</table>

The average observed horizontal wavelengths of the events reported here were similar to those of ripples, but the durations of these events were much longer (> 2 hours) than typical short-lived ripple features (< 45 min.), with larger spatial extents,
and well-defined horizontal phase velocities. For the event of July 5, 2003, the wind field was able to cause ducting of the small-scale gravity wave in the plane of wave propagation [Chimonas and Hines, 1986; Isler et al., 1997; Snively et al., 2007]. Given the strong negative background wind flow for the event of November 30, 2003, the small-scale, short-period gravity wave was evanescent in nature. Detailed data analyses and numerical modeling results for these ducted and evanescent wave events are discussed below in Section 6.3, including two numerical case studies in which comparable waves are simulated.

6.3 Results and Discussion

6.3.1 July 5, 2003: Ducted wave event

Figures 6.1a and 6.1b show extensive and distinct small-scale gravity wave structure imaged respectively in the OH emission at 11:37 UT and the O\textsubscript{2} emission at 11:40 UT on July 5. This prominent wave event was seen extending over the entire field of view with similar horizontal scale to ripple waves [Hecht, 2004]. The wave was observed for 174 min. from 10:51 to 13:45 UT in OH emission and from 10:54 to 13:48 UT in O\textsubscript{2} emission. In both OH and O\textsubscript{2} emissions, the wave was seen progressing northeast at \(\sim 50^\circ \pm 4^\circ\) as measured clockwise from north. The arrow indicates the direction of propagation of the wave. The average phase speeds of 37 and 36 ms\(^{-1}\) with equal horizontal wavelength of \(\sim 15 \pm 1\) km were measured in OH and O\textsubscript{2} emissions, suggesting an observed period of \(\sim 7 \pm 1\) min.

Figure 6.2 plots vertical profiles of the background wind \((u_0)\) in the direction of horizontal motion of the wave and the vertical wave number squared \((m^2)\). The vertical profiles in Figure 6.2a show the background wind \([u_0(z)]\) and modeled wind \([U_0(z)]\) projected along the observed direction of wave propagation, and derived from hourly meteor radar wind data at the time 12:00–13:00 UT of the wave occurrence.
The wind profile exhibited maximum positive flow of $\sim 12 \text{ ms}^{-1}$ at 92 km height, and gradually decreased above and below to minimum speeds of $\sim -33 \text{ ms}^{-1}$ at 82 km and $\sim -36 \text{ ms}^{-1}$ at 99 km. For this analysis, we estimated the measurement uncertainty as the standard deviation of the mean to be $\sim 4 \text{ ms}^{-1}$ for zonal and meridional winds between 80 and 100 km. The measurement uncertainty for background wind calculated as the standard deviation of the mean was $\sim 4 \text{ ms}^{-1}$ over height range of 80–100 km.

In order to derive the local vertical wave number squared ($m^2$) these values of background winds ($u_0$) together with observed horizontal wave numbers ($k = 2\pi/\lambda_x$), a scale height ($H$) of 6 km and static stability, $N^2$ in the mesopause region were inserted into Equation (6.2). Figure 6.3 depicts the MSISE-90 model temperature and Brunt-Väisälä frequency squared profiles for the events of July 5 and November 30, 2003. The profile of $N^2$ was derived from MSISE-90 atmospheric model temperature profile over 80–100 km altitude at 12:00 UT as shown in Figure 6.3a. To confirm that the MSISE-90 data is approximately valid for these observations, the average Brunt-Väisälä frequency is also derived from MTM temperature data. The OH and O$_2$ temperatures were found to be 210 K and 206 K, respectively, leading to an average $N^2$ of 0.000404 s$^{-2}$, consistent with the MSIS model data. The small temperature gradient suggests that significant inversion layers were not present in the airglow region during the time of the event.

Shown in Figure 6.2b, the computed values of vertical wave number squared are positive ($m^2 > 0$) at 85–95 km height range where the positive peak of $m^2 \sim 6.8 \times 10^{-7}$ occurred at $\sim 92$ km. The vertical wave number squared of the ducted wave approaches negative at altitudes of reflecting levels $\sim 84$ and $\sim 96$ km where the background wind speed becomes abruptly negative in the direction of wave motion. The negative background winds at height ranges of 80–84 and 96–100 km lead to
negative $m^2$, forming an evanescent boundary regions on both sides of the ducting region. The minimum value of intrinsic phase speed ($c - u_0$) was $\sim 25$ ms$^{-1}$, occurring at maximum $m^2$ at the center of the duct (92 km). The intrinsic frequency ($\hat{\omega}$) and period ($\hat{\tau}$) of the wave were approximately 0.01 s$^{-1}$ and 10 min at the center of duct (92 km), and the average intrinsic frequency (or intrinsic period) in the evanescent regions below and above the duct was 0.0285 s$^{-1}$ ($\sim 3.8$ min.).

The wind structure alone was able to support ducted small-scale wave propagation within a $\sim 10$ km wide ducting region. Background winds Doppler-shift the intrinsic frequencies as they vary with height from maximum value to minimum value in the direction of wave propagation. The Doppler-shifted intrinsic frequency, $(\hat{\omega} = \omega - ku_0)$ was less than Brunt-Väisälä frequency, $N$ (or intrinsic period, $\hat{\tau}$ was greater than the Brunt-Väisälä period) in the vicinity of the maximum positive background wind at 92 km for the July 5 event [Chimonas and Hines, 1986].

Figure 6.2a also depicts the model horizontal wind $U_0(z)$, calculated from Equation (6.5), using parameters based on wind observations, including amplitude.
Figure 6.3. Plots depict the MSISE-90 model temperature (red) and Brunt-Väisälä frequency squared (blue) profiles for (a) 5 July 2003 ducted wave event at 12:00 UT, and (b) 30 November 2003 evanescent event at 15:00 UT.

of $\hat{U} = 23\text{ m s}^{-1}$ at reference height $z_0 = 92\text{ km}$, average vertical wavelength of $\Lambda_z = 19\text{ km}$, and unperturbed mean background wind of $U_{bg} = -12\text{ m s}^{-1}$. The analytical horizontal wind profile agrees well with the observed background wind profile at altitude of 80–100 km, and is used in numerical simulations presented in section 6.3.3. As shown in Figure 6.2b, $m^2$ using modeled horizontal wind is approximately consistent with the $m^2$ computed using observed background wind data, indicating positive peak of $m^2$ at altitude of $\sim 92\text{ km}$, and $\sim 10\text{ km}$ wide ducting region.

Figure 6.4 show the time-altitude contours of background wind in the direction of wave propagation and $m^2$ derived from Taylor-Goldstein equation using hourly background wind in the direction of wave propagation for the wave event recorded on the night of July 5. Since observed phase speed of the wave was $37\text{ m s}^{-1}$, the background winds in Figure 6.4a measuring between $\sim 20\text{ m s}^{-1}$ (positive) and $\sim 20\text{ m s}^{-1}$ (negative) lead to the region of positive $m^2$ at the height range of $\sim 85–95\text{ km}$, while the negative background winds greater than $\sim 20\text{ m s}^{-1}$ above $\sim 96\text{ km}$ and below $\sim 84\text{ km}$ lead to the regions of negative $m^2$ during the time of wave occurrence.
Figure 6.4. Contour plots for July 5, 2003 ducted wave event depict (a) the measured background wind field over the course of the wave event (11:00–14:00 UT) and diurnal cycle, and (b) the calculated $m^2$ derived from background wind and wave parameters over the course of the wave event (11:00–14:00 UT) and diurnal cycle.

The structure of the background winds influence the nature of wave structure at the time of observation. At 12:00 UT, the positive (or negative) background winds less than 15 ms$^{-1}$ was able to make $m^2$ positive, providing a ducting region at 85–95 km altitude range, and significantly increasing negative background winds greater than 20 ms$^{-1}$ above and below this height range cause to the region of negative $m^2$.

The time-height contour of $m^2$ in Figure 6.4b shows that the propagating region ($m^2 > 0$) of the wave in the duct was formed at the height of range 85–95 km bounded by evanescent regions ($m^2 < 0$) at the top and bottom sides of the duct during the time of wave occurrence. From 10:00 to 11:00 UT, a similar wave would have been nearly evanescent for a large altitude range. At 14:00 UT and later, the wave would be able to propagate over an expanding altitude range, and potentially subject to critical level dissipation as the wind approaches and eventually exceeds its horizontal phase velocity. Thus, the ducting region was significantly modulated in width, depth, and altitude over time as the wave was observed.
6.3.2 November 30, 2003: Evanescent wave event

Figures 6.1c and 6.1d show small-scale wave structures imaged respectively in OH emission at 15:16 UT and the O$_2$ emission at 15:19 UT on November 30. This event extended over the entire field of view as coherent waves in both OH and O$_2$ emissions, progressing southwest at $\sim 250^\circ \pm 4^\circ$, as measured clockwise from north. The arrow indicates the direction of the wave propagation. The horizontal wave parameters were derived from several successive images, with average horizontal wave wavelength, phase speed and observed period determined to be $\sim 17 \pm 1$ km, $\sim 25 \pm 3$ ms$^{-1}$ and $\sim 11 \pm 1$ min., respectively. The wave event duration was $\sim 140$ min observed from 13:20 to 15:40 UT in OH emission and from 13:23 to 15:43 UT in O$_2$ emission.

Figures 6.5a and 6.5b plot vertical profiles of the background wind ($u_0$) along the observed direction of horizontal motion of the wave and the vertical wave number squared ($m^2$) at the time of 15:00–16:00 UT. The profile in Figure 6.5a shows the strong negative background winds, clearly directed opposite to that of the wave motion over the altitude range of 80–100 km. The maximum of the projected background wind in the opposite direction of wave motion was $\sim 80$ ms$^{-1}$ at 91 km height. The magnitude of background wind gradually decreased below and above this peak height. For this analysis, we have recorded the measurement uncertainty as the standard deviation of the mean to be $\sim 4$ and $\sim 5$ ms$^{-1}$ for zonal and meridional winds over 80–100 km. Likewise, the calculated uncertainty for background wind as the standard deviation of the mean over height range of 80–100 km was $\sim 3$ ms$^{-1}$.

As described above, we calculated local vertical wave number squared ($m^2$) from Equation (6.2) using measured values of background wind ($u_0$) and observed horizontal wave number ($k = 2\pi/\lambda_x$) with a scale height ($H = 6$ km) and $N^2$ obtained from MSISE-90 atmospheric model temperature profile [Hedin, 1991] as shown in
Figure 6.5. Plots for November 30, 2003 evanescent wave event depicting the vertical profiles of (a) the measured (blue-solid) and modeled (red-dashed) background wind fields with respect to wave propagation, (b) the Taylor-Goldstein vertical wave number squared \( (m^2) \) computed for measured (blue-solid) and modeled (red-dashed) background wind fields.

Figure 6.3b. Again, to confirm that the MSISE-90 data is approximately valid for these observations, the average Brunt-Väisälä frequency was derived from MTM temperature data. The OH and O\(_2\) temperatures were found to be 208 K and 213 K, respectively, leading to an average \( N^2 \) of 0.000466 s\(^{-2}\), consistent with the MSIS model data. The small temperature gradient again suggests that significant inversion layers were not present in the airglow region during the time of the event.

As shown in Figure 6.5b, the computed \( m^2 \) for this event reveals strong evanescence \( (m^2 < 0) \) over the 80–100 km range encompassing the OH and O\(_2\) emission layer peaks with minimum \( m^2 \) of \( \sim -0.9 \times 10^{-7} \) m\(^{-2}\) at 91 km altitude. This was due to the large negative background winds encountered by the wave of small phase speed \( \sim 25 \) ms\(^{-1}\) throughout the range of the wind field. The greater increase in the magnitude of \( (c - u_0)^2 \) due to large negative background winds contributes to the negative \( m^2 \) in this region. The maximum value of intrinsic phase speed \( (c - u_0) \) of \( \sim 105 \) ms\(^{-1}\) of the wave suggests a minimum intrinsic period of \( \hat{\tau} = 2.7 \) min. at 91 km, and the average period \( \langle \hat{\tau} \rangle \) over the 80–100 km height range was 3.6 min. The intrinsic phase speeds
\( c - u_0 \) gradually decrease below and above the peak height with slower negative background wind flow, but remain large enough for \( m^2 \) to be negative over a large region, suggesting the wave was likely evanescent in nature from 80–100 km. This indicates higher intrinsic frequency (and lower intrinsic period) than the Brunt-Väisälä frequency (and Brunt-Väisälä period) of the evanescent wave. Therefore, in order for the wave to have been propagating, rather than evanescent, at this altitude, the Brunt-Väisälä period would need to be < 2.7 min. over a range of several kilometers, which is highly unlikely given the measured height-averaged MTM temperatures and estimated \( N^2 \) of 0.000466 s\(^{-2}\), suggesting a Brunt-Väisälä period of 4.85 min.

Figure 6.5a also depicts the model horizontal wind \( U_0(z) \) as calculated from Equation (6.5). The parameters used for the modeled wind profile include amplitude of \( \hat{U} = -22 \text{ ms}^{-1} \) at reference height \( z_0 = 91 \text{ km} \), average vertical wavelength \( \Lambda_z = 22 \text{ km} \) and unperturbed mean background wind of \( U_{bg} = -57 \text{ ms}^{-1} \). The modeled horizontal wind profile agrees well with the observed background wind profile over the range of 80–100 km. This modeled horizontal wind was then inserted in Equation (6.2) to derive the vertical wave number squared (\( m^2 \)). As shown in Figure 6.5b, the profile of negative values of the modeled \( m^2 \) agrees well with the negative value of computed \( m^2 \), leading to the strong evanescence.

Figure 6.6 shows the time-altitude contour plots of background wind in the direction of wave propagation as derived from the zonal and meridional wind components and \( m^2 \) derived from Taylor-Goldstein equation using hourly background wind in the direction of wave propagation for November 30 event. As shown in Figure 6.6a, the strong background wind opposite to the direction of wave motion was progressed down from 100 km to 80 km at \( \sim 9:00–23:00 \text{ UT} \) time. Because of the average observed phase speed (25 ms\(^{-1}\)) of the wave, the average negative background wind of 57 ms\(^{-1}\) measured between 13:00 and 16:00 UT leads to the region of negative
Figure 6.6. Contour plots for November 30 evanescent wave event depict (a) the measured background wind field over the course of the wave event (13:00–16:00 UT) and diurnal cycle, and (b) the calculated $m^2$ derived from background wind and wave parameters over the course of the wave event (13:00–16:00 UT) and diurnal cycle.

$m^2$ where the wave was occurred. At 15:00 UT, the negative background wind of 80 ms$^{-1}$ was measured at 91 km and then negative background wind magnitudes decreased from this peak height to 31 ms$^{-1}$ at 80 km and 42 ms$^{-1}$ at 100 km.

Because of strong negative background wind the time-height contour of $m^2$ in Figure 6.6b shows the evanescent region ($m^2 < 0$) progressing down from 9:00 UT at 100 km to 23:00 UT at 80 km. The derived negative values of $m^2$ indicated that the evanescent region was occurred at the altitude range of 88–100 km at 13:00 UT and of 80–98 km at 14:00 UT, with variation of background winds. At 15:00 UT, evanescent region extended throughout the range of 80–100 km. The region of negative $m^2$ occurred due to the presence of large background wind speed in the opposite direction of the wave motion, causing the wave of observed speed 25 ms$^{-1}$ to be purely evanescent in this region. However, wave was likely able to propagate above the region of evanescence, specially at later times after 15:00 16:00 UT.
6.3.3 Numerical model simulation

Numerical simulation results are obtained with the model of *Snively and Pasko* [2008], using the CLAWPACK software package. Background atmospheric conditions use the analytic model wind profiles depicted in Figure 6.2a and 6.5a, along with MSISE-90 temperature and neutral density fields obtained for the times and locations of the events *Hedin, 1991*. Increasing molecular kinematic viscosity and thermal conduction naturally damps waves that propagate vertically towards the upper boundary, with no auxiliary sponge layer *Snively and Pasko, 2008, and references cited therein*. The domain boundaries extend from -500 to +500 km in the horizontal direction (1000 km total), and 0 to 220 km in the vertical direction, with 0.5 km uniform grid spacing.

Waves are generated using a vertical force applied by an idealized momentum source located near tropopause. The source is a traveling wave oscillator with Gaussian form that excites waves principally in the rightward horizontal direction. The source of the gravity waves is positioned at \(x_0 = 0\) km and \(z_0 = 12\) km (i.e., 500 km from the left boundary of the simulation domain and at 12 km altitude), with peak forcing occurring at \(t_0 = 1500\) seconds. It provides a vertical force at a chosen frequency \((\omega)\) and horizontal wave number \((k_x)\) of the form

\[
\sim \exp[-(x-x_0)^2/2\sigma_x^2 -(z-z_0)^2/2\sigma_z^2 -(t-t_0)^2/2\sigma_t^2] \cos[\omega t - k_x(x-x_0) - k_z(z-z_0)],
\]

where \(\sigma_x\) and \(\sigma_z\) are the Gaussian envelope’s horizontal and vertical half-widths, respectively, and \(\sigma_t\) is the temporal Gaussian half-width; the position given by \(x_0, z_0,\) and \(t_0\) corresponds to the source maximum in space and time. The source characteristics are consistent with observed ground-relative wave parameters. For the ducted wave case study, parameters are defined: \(\omega = 0.0155\) rad/sec \((T = 6.75\) min.), \(k_x = 4.18 \times 10^{-4}\) rad/m \((\lambda_x = 15\) km\), \(k_z = 0\), \(\sigma_x = 15\) km, \(\sigma_z = 3\) km, and \(\sigma_t = 25\) minutes. For the evanescent wave case study, parameters are defined: \(\omega = 0.0092\)
rad/sec ($T = 11.38$ min.), $k_x = 3.7 \times 10^{-4}$ rad/m ($\lambda_x = 17$ km), $k_z = 0$, $\sigma_x = 15$ km, $\sigma_z = 3$ km, and $\sigma_t = 25$ minutes.

Figure 6.7a depicts the numerical model results for the Doppler-ducted wave simulation (July 5 Case Study), at a time 100 minutes after the initial turn-on of the source. The wave propagates within the analytical wind profile shown in Figure 6.2a. It has been assumed that the wave is forced from below by a tropospheric source, and becomes trapped via linear tunneling into the duct through the evanescent boundary near 80–85 km altitude. This simple case of ducted wave excitation is consistent with the process described by Walterscheid et al. [2001]. This forms a robust mechanism for wave excitation at mesopause altitudes, particularly for waves of 6–8 minute ground-relative periods. The wind graphic overlaying the numerical results illustrates the approximate local wind profile. It is clear that the propagation region of the ducted wave is concentrated near the wind peak region occurring at $\sim 90$ km, with evanescent standing wave structure above and below. Ducted propagation also occurs below $\sim 75$ km, as the wave is able to propagate at lower altitudes in the mesosphere. It is important to note that this wave is a non-ideally ducted packet, and it will exhibit propagation loss and periodic upward and downward momentum flux as it propagates [e.g., Yu and Hickey, 2007]. Local wave-induced vertical wind velocities are on the order of a few ms$^{-1}$ at $\sim 90$ km, with peak velocities near $\sim 10$ ms$^{-1}$, and may be expected to lead to strong perturbations of the airglow layers [Hickey, 2001; Snively et al., 2007]. At larger magnitude, breaking of the wave within or above the duct may occur.

Figure 6.7b depicts the numerical model results for the evanescent wave simulation (November 30 Case Study), excited within the analytically specified wind field depicted in Figure 6.5a. Again, the wave is forced from below by a tropospheric source, and reaches evanescence at $\sim 75$ km altitude. The gradual shift to
Figure 6.7. Model simulation results depicting upward wave propagation through the mesopause region, for (a) wind profile consistent with non-ideal Doppler ducting from 87–97 km, based on the wave event of July 5, 2003, and (b) wind profile consistent with evanescent propagation from 75–100 km, based on the wave event of November 30, 2003. In both (a) and (b), data is plotted at a time 100 minutes after start of the model simulation. The waves are seen refracted by the ambient wind field, leading to regions of propagation and evanescence, and thus weak trapping.

near-infinite vertical wavelength leads to strong standing wave structure below the evanescent layer, where the wave experiences gradual downward reflection. Simultaneously, a portion of wave energy tunnels through this layer of evanescence, to emerge above at 100 km altitude. The positive wind flow leads to relatively small vertical wavelengths, and thus enhanced dissipation by viscosity. The wave-induced vertical winds are again on the order of a few ms$^{-1}$ within the OH and O$_2$ airglow
layers, and quite sufficient for measurable perturbations of airglow emissions intensity to arise. Given the relatively strong gravity wave activity present in the model throughout mesopause, despite evanescence, it appears quite reasonable that waves propagating upward from below could explain the observed signatures. Actual wave magnitude, and airglow intensity variations, would thus be determined by original source intensity.

Model results suggest that, in both cases, the dominant airglow signatures may have arisen from the evanescent standing wave structure, which is present throughout the airglow region (note wave structures at 95–100 km and 80–85 km in Figure 6.7a, and 75–95 km in Figure 6.7b). Such waves are quite effective at perturbing the airglow layers [e.g., Hines and Tarasick, 1994; Snively et al., 2007], as a result of the long vertical wavelengths, and nearly vertical polarization, which enhance density perturbations to layered minor species.

### 6.4 Summary and Conclusions

Simultaneous airglow image measurements in the OH and O$_2$ emissions from a mesosphere temperature mapper (MTM) have been used to investigate the horizontal parameters of small-scale ducted and evanescent gravity waves, and their propagation characteristics. Average observed wave parameters of 15 events include horizontal wavelengths of $\sim$18 km, phase speeds of $\sim$37 ms$^{-1}$, and observed periods of $\sim$8 min. Coincident meteor radar wind data have been used to investigate the intrinsic properties of these waves. This analysis reveals that nine of the total wave events were Doppler-ducted and six of them were evanescent in the mesopause airglow region. All exhibit relatively similar airglow signatures.

Two wave events were selected for numerical case studies. The first of the two wave events studied was clearly ducted in the airglow region. Observations strongly
indicate that the wave event was trapped non-ideally in a well-defined \( \sim 10 \) km wide Doppler duct that persisted for \( \sim 3 \) hours, coinciding with the approximate duration of observation. The second wave event exhibited similar wave characteristics (wavelength, and observed phase velocity and period), but was purely evanescent throughout the 80–100 km region, as a consequence of strong opposing background winds. Numerical model simulations indicate that the measured ducted and evanescent wave events are consistent with theoretical Doppler-ducted and evanescent wave characteristics. In both cases the background winds strongly influenced the observability of the waves and the local wave vertical structure. Despite the obstacle of evanescence, model results indicate that it is quite reasonable for the observed waves to have been generated by tropospheric convective sources [e.g., Alexander, 1996], and to be of sufficient intensity to be detectable in airglow data.
CHAPTER 7
OBSERVATION AND ANALYSIS OF MESOSPHERIC BORES IN OH AND O$_2$ AIRGLOW EMISSIONS

7.1 Introduction

Wave-like structures have been frequently observed in the mesopause airglow emission layers. The most commonly observed band-type wave structures are considered to be due to freely propagating [Taylor et al., 1995b, 1997] and ducted or evanescent [Isler et al., 1997; Walterscheid et al., 1999] internal atmospheric gravity waves propagation through these emission layers. In addition, there exist two distinct types of frontal disturbances, which occur in airglow emissions. The first type of disturbance is the bore event that is characterized by a sharp leading front followed by either a train of propagating phase-locked wave crests known as undular bore or an extended turbulent region known as a turbulent bore. The change in airglow intensity and temperature are often accompanied with the passage of the leading front. Bores are a class of nonlinear gravity waves that exist in two distinct forms as sharp boundary bores in which a density discontinuity exists and internal bores which exist in a stably stratified fluid. The other type frontal disturbance is termed as a wall event [Swenson et al., 1998]. This event is referred to as the large-scale freely propagating gravity wave that becomes unstable and then its leading edge induces significant perturbation in the mesopause airglow emission brightness [Swenson and Espy, 1995; Swenson and Liu, 1998].

The first spectacular mesospheric bores were observed by Taylor et al. [1995d] during the ALOHA-93 campaign in four nightglow emissions from Hawaii. Dewan and Picard [1998] suggested that this wave event exhibited characteristics of an internal undular mesospheric bore. They developed a simple linear model for bore
propagation into a ducting region in the mesosphere. Dewan and Picard [2001] further suggested that mesospheric bores may occur as a result of gravity wave interaction with the mean wind flow at a critical level, providing a strong temperature inversion layer that acts as a wave duct for the bore channel. Unlike wall events, internal bores required a stable ducting region to propagate horizontally over large distance without significant attenuation. In the upper mesosphere and lower thermosphere (MLT), a temperature inversion or a vertical wind shear can provide a stable ducting region for the propagation of a bore.

There have been reported several mesospheric bores observed at mid, high and equatorial latitudes. Smith et al. [2003] and She et al. [2004] reported mesospheric internal undular bores associated with the temperature inversion layer and proposed that the thermal duct would support the wave fronts propagation. Fechine et al. [2005] and Medeiros et al. [2005] have reported the occurrence of mesospheric bores observed at equatorial latitudes, and Nielsen et al. [2006] observed an unusual mesospheric bore event at high altitudes over Antarctica. The existence of a ducting region for observed mesospheric front propagation was also supported by Doppler duct [Shiokawa et al., 2006; Fechine et al., 2009].

We have analyzed the frequency of occurrence of bore-like events and the complementary effects in OH and O$_2$ emission layers observed in the low-latitude mesopause region from Hawaii. We also report the wave parameters such as horizontal wavelength, phase speed, observed period and propagation direction. We have described the data analysis in Section 7.2. Observations and results of bore-like events are presented in Section 7.3. In Section 7.4, we discuss case studies for investigation of propagation conditions of the most prominent and unique mesospheric bores. Then these are followed by summary and conclusions in Section 7.5.
7.2 Data Analysis

As part of the Maui-MALT program, the USU mesospheric temperature mapper (MTM) datasets were utilized to determine the occurrence frequency and horizontal parameters of short-period front-like events imaged in both the OH \([P_1(2)\) and \(P_1(4)\) and \(O_2\) (866 and 868 nm) nightglow emissions at low-latitude from Haleakela, Maui HI \(20.7^\circ\ N, 156.3^\circ\ W\). The exposure time for each image was 1 min. with two observations per emission followed by a background image, resulting in a cycle time of \(\sim5.5\) min. Each image of a circular field view of \(90^\circ\) was flat-fielded in order to correct the nonuniform intensity in the image. The flat-fielded OH and \(O_2\) data were then calibrated using the background star field in order to get geographic orientation and then unwarped by projecting onto a uniformly sampled 180 km \(\times\) 180 km rectangular grid \(\text{e.g.,}\ Garcia\ et\ al.,\ 1997;\ Pautet\ and\ Moreels,\ 2002\) assuming nominal emission altitudes of \(\sim87\) km and \(\sim94\) km, respectively.

The horizontal wavelength and propagation direction of each wave event was then computed using a three-dimensional Fast Fourier Transform (FFT) spectral analysis technique to determine their unambiguous spectrum using a time series of typical four images \(\text{e.g.,}\ Garcia\ et\ al.,\ 1997;\ Coble\ et\ al.,\ 1998\). The apparent phase speeds of waves were determined separately by measuring the horizontal displacement of selected wave crests as a function of time. The apparent periods were then calculated from measured wavelengths and phase speeds. The University of Illinois (UIUC) meteor radar hourly wind measurements were used to investigate the Doppler ducting of the observed bore event, while Na lidar wind and temperature measurements of 15 minutes-100 meters temporal-spatial resolution were used to determine the thermal ducting of other observed bore events.

Vertical structure of waves can be estimated under the Boussinesq approximation, using the Taylor-Goldstein equation. The second order linear, irrotational
and inviscid Taylor-Goldstein equation in incompressible and stratified atmosphere

\[ Nappo, 2002 \]

for the vertical perturbation velocity \( w_z \) is

\[
\frac{d^2 w_z}{dz^2} + \left[ \frac{N^2}{(c - u_0)^2} + \frac{1}{(c - u_0)} \frac{d^2 u_0}{dz^2} - \frac{1}{H(c - u_0)} \frac{du_0}{dz} - k^2 - \frac{1}{4H^2} \right] w_z = 0, \tag{7.1}
\]

where \( N \) is the Brunt-Väisälä frequency, \( c \) is the observed phase speed; \( u_0 \) is the
bloodwind in the direction of wave propagation; \( \hat{\omega} = \omega - u_0 k \) is the intrinsic
frequency of wave; \( k = \left( \frac{2\pi}{\lambda_x} \right) \) is the horizontal wave number correspond to a horizontal
wavelength (\( \lambda_x \)); and \( H \) is the scale height. After neglecting the curvature and
vertical shear terms, the WKB solution to the Taylor-Goldstein equation (7.1) for
vertical wave number, \( m \), is given by

\[
m^2 = \frac{N^2}{(c - u_0^2)} - k^2 - \frac{1}{4H^2}. \tag{7.2}
\]

If both functions, \( N^2 \) and \( u_0 \), are slowly varying with altitude, a region of ducting
will exist for the propagation of waves in the horizontal direction. By computing
the local vertical wave number squared, \( m^2 \), profiles, the boundary condition of the
ducting region can be inferred [Isler et al., 1997; Snively et al., 2007; Simkhada et al.,
2009]. A wave is vertically propagating if \( m^2 > 0 \) and it is vertically evanescent if
\( m^2 < 0 \). Due to variation in the wind (\( u_0 \)) and \( N^2 \), the wave can be ducted in a
region of vertical propagation (\( m^2 > 0 \)) bounded by regions of evanescence (\( m^2 < 0 \))
in which the wave cannot be freely propagated. The region of evanescence is favored
by large \( k \) when \( N^2 \) is small or \( (c - u_0) \) is large, causing reflection of wave energy.

However, in our analysis here, we have computed \( m^2 \) from equation (7.2)
using airglow image data to estimate ground-relative phase speed (\( c \)) and meteor
radar wind data to estimate background wind in the direction of wave propagation.
We have used Brunt-Väisälä frequency squared, \( N^2 \), as function of altitude from the
MSISE-90 atmospheric model [Hedin, 1991], and scale height, \( H \), is taken to be 6
km in the mesopause region. However, \( N^2 \) can vary with altitude in this region, but it is assumed that the variations in wind field \( (u_0) \) will dominate over variations in \( N^2 \) for the formation of Doppler ducts [Isler et al., 1997].

The background wind in the direction of wave propagation \((\theta)\), as measured clockwise from north, is calculated using meteor wind data as

\[
 u_o(z) = u_x \sin(\theta) + u_y \cos(\theta),
\]

where \( u_x \) and \( u_y \) are zonal and meridional components of observed wind profile, respectively. For the analytical modeled wind structure, we fit a function to approximate a tidal wind field by assuming the wave-like solution as

\[
 U_0(z) = U_{bg} + \hat{U} \exp \left( \frac{z - z_o}{2H} \right) \cos \left( K_z (z - z_o) \right),
\]

where \( U_{bg} \) is the unperturbed mean background wind, \( \hat{U} \) is wind amplitude at the reference height \((z_0)\) and vertical wave number, \( K_z = \frac{2\pi}{\Lambda_z} \), is taken to correspond to a vertical wavelength \((\Lambda_z)\). The parameters are obtained by fitting the peak of the wind in agreement with measured wind at reference height \((z_0)\), such that \( U_0(z_0) = U_{bg} + \hat{U} \) and vertical wavelength is obtained by direct measurement of the wind profile. The analytical wind model is then used to derive the vertical wave number squared, \( m^2 \), from equation (7.2), and compared with \( m^2 \) derived from the measured wind field to confirm agreement.

The UIUC Na lidar temperature data is used to confirm significant presence of temperature inversions at the times of wave events and to provide a good approximation of local thermal conditions. The local Brunt-Väisälä frequency squared, \( N^2 \), can be determined from the temperature gradient as

\[
 N^2 = \frac{g}{T_a} \left( \frac{dT_a}{dz} - \frac{dT_d}{dz} \right),
\]

where \( g \), \( T_a \), \( \frac{dT_a}{dz} \) and \( \frac{dT_d}{dz} \) are acceleration due to gravity, atmospheric temperature,
temperature gradient, and adiabatic lapse rate. In the mesopause region, adiabatic lapse rate and \( g \) are taken to be \(-9.5 \text{ K km}^{-1}\) and \(9.5 \times 10^{-3} \text{ km s}^{-2}\). The existence of instabilities can be quantified by considering the Brunt-Väisälä frequency squared, \( N^2 \) and the gradient Richardson number, \( Ri \) [Gossard and Hooke, 1975]. By measuring vertical profiles of temperature and wind gradients, the Richardson number, \( Ri \), is computed as

\[
Ri = \frac{N^2}{\left( \frac{du}{dz} \right)^2 + \left( \frac{dv}{dz} \right)^2},
\]

(7.6)

where \( \frac{du}{dz} \) and \( \frac{dv}{dz} \) are zonal and meridional wind gradients. If \( Ri \) is \(0 < Ri < 0.25\), dynamic instability occurs and if \( Ri \leq 0\), convective instability occurs. For the derivation of temperature and wind gradients, \( N^2 \) and \( Ri \) values, 1 km averaged temperature and wind data were applied between 80 and 100 km altitude. The measurement errors for meteor radar wind, and Na lidar wind and temperature are calculated as the standard deviation and then uncertainties in the wind measurements are estimated as the standard deviation of the mean in the data analysis.

For the analytical modeled temperature structure, we fit a function to approximate a tidal temperature field by assuming a wave-like solution as

\[
T_0(z) = T_{bg} + \hat{T} \exp \left( \frac{z - z_o}{2H} \right) \cos \left[ K_z' (z - z_o) \right],
\]

(7.7)

where \( T_{bg} \) is unperturbed mean background temperature, \( \hat{T} \) is temperature amplitude at the reference height \((z_o)\), and vertical wave number \( K_z' = \frac{2\pi}{\Lambda_z'} \) is taken to correspond to a vertical wavelength \( \Lambda_z' \). The parameters are obtained by fitting the peak of the temperature until it agrees with measured temperature at reference height \( z_0 \), such that \( T_0(z_0) = T_{bg} + \hat{T} \) and vertical wavelength is obtained by direct measurement of the temperature profile. The analytical temperature model is then used to derive Brunt-Väisälä frequency squared, \( N^2 \) from equation (7.5), and compared with \( N^2 \).
derived from the measured temperature profile to confirm agreement.

Here we presented in particular two case studies. First, we examined the Doppler ducting characteristics of a small-scale bore event, as determined by meteor radar measurements of hourly background wind along the direction of wave propagation. Second, we examined the thermal ducting characteristics of two oppositely moving small-scale bore events, as determined by Na lidar wind/temperature measurements. Detailed data analyses and results for these ducted wave events are discussed below in Section 7.4.

7.3 Observations and Results

Airglow imaging observations were carried out from Maui, Hawaii in the low latitude mesopause region during 2003–2004. From a total of 64 front-like events, 44 events were observed on 37 nights in 2003 and 20 events were observed on 17 nights in 2004. During 2003, 36 events were observed and recorded in both OH and O₂ emissions, and only four events in the OH emission and only four events in the O₂ emission. In 2004, a total of 20 events were recorded in both OH and O₂ emissions. Figure 7.1a shows the frequency of occurrence of bore-like events observed in OH and O₂ images in each month. Of the total events, 26 events in summer and 13 events in winter were recorded. It indicates that the occurrences of bore activity were higher in summer than in winter. Also the frequency of bore-like event activity was higher in autumn months (September and October), with less activity in spring. The minimum and maximum duration of bore events were 30 and 116 minutes, respectively, and the average duration of the wave was 62 minutes.

Figure 7.1b shows the statistics of patterns of the complementary behavior of OH and O₂ airglow layers for observed bore-like events. It is notable that 23 bore-like events showed the complementary effects of brightness in the OH and darkness in the
Figure 7.1. Plots showing frequency of occurrence of (a) the number of events for each month for 2003 and 2004, and (b) the patterns of complementary effects in OH and O\textsubscript{2} layers of bore-like events observed in 2003−2004.

O\textsubscript{2}, (BD), and the other 23 bore-like events showed darkness in the OH and darkness in the O\textsubscript{2}, (DD). The patterns of complementary effects of bright-bright (BB) and dark-bright (DB) in OH and O\textsubscript{2} airglow emission layers were exhibited by four and six bore events, respectively. Only four events observed in OH emission exhibited dark fronts (D). Of four events observed in the O\textsubscript{2} emission, only one event showed a dark front (D) and the other three events showed bright fronts (B). According to the Dewan and Picard [1998] wave model of complementary effects in the emission layers, the brighter or darker front of the images is used to characterize the lowering or raising of the airglow layer. The patterns BD and DD of the complementary effects in OH and O\textsubscript{2} emission layers have a higher frequency of occurrence, suggesting a possibility of duct formation between two altitudes of OH (∼87 km) and O\textsubscript{2} (∼94 km) layers or at around the altitude of the OH (∼87 km) layer.

Figure 7.2 displays the distribution of the observed horizontal wavelengths, horizontal phase speeds and observed periods of bore-like events recorded simultaneously in OH and O\textsubscript{2} emissions during 2003−2004. As shown in Figure 7.2a, the
Figure 7.2. Histogram plots showing the distribution of (a) horizontal wavelengths, (b) apparent phase speeds, and (c) apparent periods of bore-like events observed during 2003–2004.

distribution exhibits a range of horizontal wavelength extending from 12 to 42 km that is plotted in the histogram as 6 km in width. The highest frequency of occurrence of wavelengths was the ranges of 12−18 and 18−24 km with an average wavelength of 21 km. Figure 7.2b displays the distribution of wave phase speeds binned at 10 ms$^{-1}$ intervals as shown in the histogram. There is a clear tendency of the waves to exhibit the highest frequency of occurrence of phase speeds in the 38−58 ms$^{-1}$ range. The average phase speed of the bore-like events was measured as 46 ms$^{-1}$. Only two events observed in both types of emissions exhibited phase speeds in the range of 18−28 ms$^{-1}$. Likewise, one event recorded in both OH and O$_2$ emissions exhibited a phase speed of $\sim$94 ms$^{-1}$, and one event recorded in the O$_2$ emission exhibited a phase speed of $\sim$78 ms$^{-1}$. The distribution of observed periods of bore-like events binned at 2-minute intervals in the histogram is displayed in Figure 7.2c. The bore-like events exhibited a narrow range of distribution of observed periods between 4−14 min. It is notable that the bore-like events exhibited peak distribution of periods in the range of 6−8 min. with an average observed period of 7.6 min., suggesting these bore-like events were short-period waves observed in the
Mesopause airglow emissions. These results are very similar to those found in earlier study [e.g., Taylor et al., 1995d; Smith et al., 2003; She et al., 2004; Fechine et al., 2005; Medeiros et al., 2005; Nielsen et al., 2006]. The typical measurement errors calculated as standard deviations for the horizontal wavelength, phase speed, propagation direction and observed period were 1 km, 4 ms\(^{-1}\), 4° and 1 min., respectively.

The polar plots in Figure 7.3 show the distribution of propagation directions with horizontal phase velocities of the bore-like events for each season during two years period. The preferred propagation directions of bore-like events were towards

![Figure 7.3](image)

**Figure 7.3.** Plots showing the azimuthal distribution of bore-like events with phase speeds for each season during 2003—2004.
the NE during summer and generally southward in winter, though fewer wave events were observed in winter months. There was no preferential propagation direction observed in spring, but more events were observed in autumn months, which exhibited preferential propagation directions towards the S-SE. Here we present two case studies of typical mesopause bore events imaged simultaneously in both OH and O$_2$ airglow emissions on the nights of July 3 and October 21, 2003.

7.3.1 July 3, 2003 wave event

Figure 7.4 shows two pair of flat-fielded and unwarped distinct images of a short-period bore event recorded (a) in the OH emission at 11:02–11:07 UT and (b) in the O$_2$ emission at 11:05–11:10 UT on July 3, 2003. The bore event first appeared at 10:27 UT (OH) and 10:30 UT (O$_2$), and ended at 11:49 and 11:52 UT in the sequential OH and O$_2$ images, respectively, persisting for a total of 82 min. As shown in the OH and O$_2$ images, the eight to nine wave crests were clearly visible and then growing to over 15 in number. This coherent wave extended over the entire field of view in both OH and O$_2$ emissions, heading towards the northeast at 33° as measured clockwise from north. The arrow indicates the direction of the wave propagation. From the wave analysis, the horizontal wave parameters were derived using several successive images. The average horizontal wavelength and apparent phase speed were determined to be $\sim$14 km and $\sim$40 ms$^{-1}$, yielding an apparent period of $\sim$6 min. This undular mesopause bore event is characterized by a sharp bright leading front in the OH emission and a sharp dark leading front in the O$_2$ emission, followed by a growing number of phase-locked wave crests, agreeing well with the [Dewan and Picard, 1998] model of complementary effect of bright-dark in airglow layers. The complementarity brightness variations observed in both airglow emissions indicated that the bore was centered between the O$_2$ and OH layers.
Figure 7.4. Examples of flat-fielded and unwarped images of a typical undular bore observed (a) in the OH emission at 11:02–11:07 UT, and (b) in the O\textsubscript{2} emission at 11:05–11:10 UT on July 3, 2003.

7.3.2 October 21, 2003 wave events

Figure 7.5 shows four pair of flat-fielded and unwarped airglow images of small-scale bore events recorded simultaneously in OH and O\textsubscript{2} emissions on October 21, 2003. Panes (a) and (b) in Figure 7.5 show examples of two images of the first event recorded in the OH emission at 12:19–12:25 UT and in the O\textsubscript{2} emission at 12:22–12:28 UT, respectively. Panes (c) and (d) in Figure 7.5 show examples of two images of the second event recorded in OH and O\textsubscript{2} emissions at 13:12–13:17 and 13:15–13:20 UT, respectively. The sequential images of the first event were observed in OH and O\textsubscript{2} emissions at 11:50–13:00 and 11:53–13:03 UT, lasting for a total of 70 min. About 64 minutes later a sharp leading front of the second event was appeared.
Figure 7.5. Examples of flat-fielded and unwarped images of reverse bore events observed on October 21, 2003 in the OH emission at (a) 12:19–12:25 UT and (b) 13:12–13:17 UT, and in the O$_2$ emission at (c) 12:22–12:28 UT and (d) 13:15–13:20 UT.

The sequential images of this event were observed in the OH emission at 12:54–13:40 UT and in the O$_2$ emissions 12:57–13:43 UT, lasting for a total of 46 min. The first wave event traveled towards the northeast at $\sim 48^\circ$ as measured clockwise from north with phase speed of $\sim 18$ ms$^{-1}$, while the second wave event traveled towards the southeast at $\sim 215^\circ$ as measured clockwise from north with a phase speed of $\sim 20$ ms$^{-1}$. The horizontal wavelengths of the first and second events were measured to be $\sim 12$ and $\sim 16$ km, yielding apparent periods of $\sim 11$ and $\sim 13$ min., respectively. The fronts of both events in OH and O$_2$ emissions show a similar pattern orientation in a NW-SE direction with the propagation direction of the wave patterns indicated by the arrows. Both events illustrate a complementary effect of bright-dark in airglow layers Dewan and Picard [1998], exhibiting bright leading fronts in the OH layer and dark leading fronts in the O$_2$ layer. A summary of the measured characteristics of bore events described in Figures 7.4 and 7.5 is given in Table 7.1.
### Table 7.1.
Data summary of horizontal wavelength ($\lambda$), direction of propagation ($\theta$), phase speed ($c$) and observed period ($\tau$) of bore events.

<table>
<thead>
<tr>
<th>Date</th>
<th>Time (UT)</th>
<th>Imager</th>
<th>$\lambda$ ±1 (km)</th>
<th>$\theta$ ±4 (deg.)</th>
<th>$c$ ±4 (m/s)</th>
<th>$\tau$ ±1 (min.)</th>
</tr>
</thead>
<tbody>
<tr>
<td>07/03/03</td>
<td>10:27-11:49</td>
<td>OH</td>
<td>14.1</td>
<td>33</td>
<td>40</td>
<td>5.9</td>
</tr>
<tr>
<td></td>
<td>10:30-11:52</td>
<td>O$_2$</td>
<td>14.2</td>
<td>33</td>
<td>40.2</td>
<td>5.9</td>
</tr>
<tr>
<td>10/21/03</td>
<td>11:50-13:00</td>
<td>OH</td>
<td>12</td>
<td>48</td>
<td>18.2</td>
<td>11</td>
</tr>
<tr>
<td></td>
<td>11:53-13:03</td>
<td>O$_2$</td>
<td>12.2</td>
<td>49</td>
<td>18.4</td>
<td>11.1</td>
</tr>
<tr>
<td></td>
<td>12:54-13:40</td>
<td>OH</td>
<td>15.5</td>
<td>215</td>
<td>20</td>
<td>12.9</td>
</tr>
<tr>
<td></td>
<td>12:57-13:43</td>
<td>O$_2$</td>
<td>16</td>
<td>214</td>
<td>20.2</td>
<td>13.2</td>
</tr>
</tbody>
</table>

### 7.4 Discussion

Coincident meteor radar measurements of the hourly background wind in the direction of the wave propagation have been used to estimate the intrinsic properties of the July 3 bore event and its Doppler ducting characteristics. Coincident Na lidar wind and temperature measurements of the spatial 1 km averaged data have been utilized to estimate the thermal ducting characteristics of October 21 bore events. Here we discuss the results of the background winds, intrinsic phase speeds, vertical wave number squared, intrinsic periods, and Doppler-shifted intrinsic frequencies for the wave event recorded on the night of July 3, 2003. We also discuss the results of the background winds, vertical wave number squared, thermal inversion, Brunt-Väisälä frequencies, and Richardson numbers for the wave events recorded on the night of October 21, 2003.

#### 7.4.1 Undular mesospheric bore in a Doppler duct

Figure 7.6 shows the vertical profiles of background wind and vertical wave number squared for the bore event observed on the night of July 3, 2003 at 11:00–12:00 UT. Figure 7.6a depicts a height profile of the background wind ($u_0$) projected along the direction of wave propagation derived from equation (7.3) using hourly meteor radar wind data. The profile exhibits maximum positive background wind of
Figure 7.6. Plots for July 3, 2003 Doppler-ducted bore event depicting the vertical profiles of (a) measured (blue-solid) and modeled (red-dashed) background wind fields with respect to wave propagation, and (b) the Taylor-Goldstein vertical wave number squared \( (m^2) \) computed for measured (blue-solid) and modeled (red-dashed) background wind fields.

18.4 ms\(^{-1}\) at 92 km. The negative background winds appeared at 88 and 94 km. There was little change in the magnitude of background wind below 88 km, while the background wind increased rapidly in the opposite direction of the wave motion above 94 km. For this analysis, the maximum allowable measurement uncertainties as standard deviation of the mean between 80 and 100 km were \( \sim 3 \) ms\(^{-1}\) and \( \sim 4 \) ms\(^{-1}\) for zonal and meridional winds.

In order to derive the local vertical wave number squared \( (m^2) \) calculated values of background wind \( (u_0) \) together with the observed horizontal wave number \( (k = \frac{2\pi}{\lambda}) \), a scale height \( (H = 6 \text{ km}) \), and static stability parameter \( (N^2) \), in the mesopause region were inserted into the Taylor-Goldstein equation (7.2). The profile of Brunt-Väisälä frequency squared, \( N^2 \), was determined from the MSISE-90 atmospheric model temperature profile over a 80–100 km altitude range at 11:00 UT as plotted in Figure 7.7a. Figure 7.6b depicts the vertical profile of the computed vertical wave number squared \( (m^2) \) using measured background wind \( (u_0) \) and \( N^2 \) obtained from MSISE-90 atmospheric model data over the altitude range of 80–100
The values of $m^2$ were made positive ($m^2 > 0$) at the range of 84–96 km with large positive peak of $\sim 11.5 \times 10^{-7} \text{ m}^{-2}$ located at the center of the duct (92 km). However, the values of $m^2$ were positive at the range of 89–95 km using typical value of $N = 0.02 \text{ s}^{-1}$ (without using $N^2$ from MSISE-90 model data). The vertical wave number squared approaches zero at the altitude of 87 km and becomes negative at 97–100 km using $N^2$ from MSISE-90 model data. The altitudes at 87 and 97 km were considered to be reflecting levels. The negative background winds appeared above the 97 km supports to make negative $m^2$, leading to the evanescent regions above the ducting region. The wave was propagating within a $\sim 10$ km ($\sim 87–97$ km) wide Doppler ducting region. But, small background winds present below the 87 km caused the small positive $m^2$ at a height range of 84–86 km and the negative $m^2$ in 82–83 km. The result suggests that the bore event can exist in a stable Doppler duct arising from background wind structure.

Figure 7.6a also depicts the analytical modeled wind ($U_0$) calculated from equation (7.4), using parameters based on wind observations, which included the wave amplitude $\hat{U}$ of $\sim 20 \text{ ms}^{-1}$ at reference height 92 km, average vertical wavelength ($\Lambda_z$) of $\sim 15$ km, and an unperturbed mean background wind of $U_{bg} = \sim -3 \text{ ms}^{-1}$. The modeled wind profile agreed well with observed background wind profile at height ranges of 87–100 and 80–83 km, but showed the reverse in directions at 84–86 km. As shown in Figure 7.6b, the modeled vertical wave number squared ($m^2$) using modeled horizontal wind was consistent with the $m^2$ computed using observed background wind data. The modeled negative background winds above 97 km and below 87 km produced negative $m^2$, leading to the evanescent regions on either sides of the Doppler ducting region of $\sim 10$ km wide with a positive peak of $m^2$ at 92 km.

Figure 7.7b depicts the measured and modeled Doppler-shifted intrinsic frequency profiles. The intrinsic frequencies are Doppler-shifted by background winds.
Figure 7.7. Plots for July 3, 2003 Doppler-ducted bore event depicting the vertical profiles of (a) the MSISE-90 model temperature (red-dashed) and Brunt-Väisälä frequency squared (blue-solid), and (b) measured (blue-solid) and modeled (red-dashed) Doppler-shifted intrinsic frequencies.

as they vary with height from maximum value to minimum value in the direction of wave propagation. The plots show that the Doppler-shifted intrinsic frequencies, \( \hat{\omega} = \omega - ku_0 \) were less than the Brunt-Väisälä frequency, \( N \) (or intrinsic periods, \( \tau_i \) were greater than the Brunt-Väisälä period) in the vicinity of the maximum positive background winds at 92 km, there producing the wave ducting region. However, the Doppler-shifted intrinsic frequencies were greater than Brunt-Väisälä frequency, \( N \), well above and below the ducting regions, which produced evanescent regions on either sides of the duct, which provided reflecting levels to trap the waves [Chimonas and Hines, 1986]. The intrinsic frequency \( \hat{\omega} \) and intrinsic period \( \tau_i \) of the wave event at the center of duct (92 km) were \( \sim0.01 \, \text{s}^{-1} \) and 10.7 min., indicating the lower intrinsic frequency (or higher intrinsic period) than the Brunt-Väisälä frequency (or Brunt-Väisälä period) in the ducting region. Intrinsic frequencies (or periods) below and above the duct at 87 and 97 km were \( \sim0.02 \, \text{s}^{-1} \) (5 min.) and \( \sim0.026 \, \text{s}^{-1} \) (4.1 min.). The Doppler-shifted intrinsic frequencies as measured from meteor radar winds were quite consistent with the intrinsic frequencies calculated from the modeled winds for the wave event at an altitude range of 87–99 km.
Further, Figure 7.8a plots time-altitude contour of background wind in the direction of wave propagation for the July 3 event. Since the horizontal phase velocity (40 ms$^{-1}$) of the wave, the magnitude of background winds measuring between 20 ms$^{-1}$ (positive) and 15 ms$^{-1}$ (negative) support the positive $m^2$ region at 87–97 km altitude, while the negative background winds greater than 15 ms$^{-1}$ lead to the regions of negative $m^2$ above and below the duct at 11:00 UT. The structure of the background winds influence the nature of wave structure at the time of wave occurrence. The positive background winds less than 30 ms$^{-1}$ or negative background winds less than 15 ms$^{-1}$ were able to make the region of positive $m^2$, providing a strong ducting region at 85–95 km altitude range. Negative background winds greater than a magnitude of 20 ms$^{-1}$ were able to produce a region of negative $m^2$ between 10:00 and 13:00 UT.

Figure 7.8b plots the time-height contour of vertical wave number squared, $m^2$, derived from the Taylor-Goldstein equation using hourly background wind in the direction of wave propagation. It is noted that the ducting region ($m^2 > 0$) for the propagation of wave was formed at an 87–97 km altitude range bounded

![Figure 7.8](image-url)

**Figure 7.8.** Contour plots for the July 3, 2003 Doppler-ducted bore event depict (a) the measured background wind field, and (b) the calculated $m^2$ derived from background wind and wave parameters.
by evanescent regions ($m^2 < 0$) above and below the duct during the time of wave occurrence. The maximum magnitude of positive $m^2$ at the center of duct decreased with increasing vertical wavelength and then produced negative $m^2$ with infinite vertical wavelength in the evanescent regions. The results suggest that the strong Doppler duct was formed for the observed wave of speed $\sim 40 \text{ ms}^{-1}$ when the mean background wind field became lower than positive $\sim 20 \text{ ms}^{-1}$ inside the duct, and greater than negative $\sim 15 \text{ ms}^{-1}$ outside the duct.

### 7.4.2 Mesospheric reverse bores in a thermal duct

Figures 7.9a and 7.9b depict the vertical profiles of 1 km average observed temperatures and modeled temperatures for the first event phase front passed the region at 12:00 UT and for the second event phase front passed the region at 13:00 UT, respectively on the night of October 21, 2003. Both profiles show temperature inversions of $\sim 41 \text{ K}$ and $\sim 39 \text{ K}$ with a maximum of 93 and 92 km, respectively. The analytical modeled temperature ($T_0$) was derived from equation (7.7), using parameters based on observation including the wave amplitude ($\hat{T}$) of $\sim 20 \text{ K}$, unperturbed mean background temperature ($T_{bg}$) of $\sim 200 \text{ K}$, and average vertical wavelength ($\Lambda_z'$) of $\sim 12 \text{ km}$. The modeled temperature profiles were well-fitted with observed temperature profiles.

Figures 7.9c and 7.9d depict the vertical profiles of measured and modeled Brunt-Väisälä frequency squared, $N^2$, calculated from equation (7.5), using temperature gradients of measured and modeled temperatures between 80 and 100 km at 12:00 and 13:00 UT. The $N^2$ profiles show the stable and unstable regions, as well as the ducting regions for the propagation of front-like waves. Figure 7.9c shows the negative $N^2$ of $0.06 \times 10^{-4} \text{ s}^{-2}$ ($N^2 < 0$) at 85 km and nearly a zero value of $N^2$ at 86 km, and small positive $N^2$ of $3.45 \times 10^{-4} \text{ s}^{-2}$ ($N^2 > 0$) at 93 km, indicating the
Figure 7.9. Plots for October 21, 2003 thermal-ducted bore event depicting the vertical profiles of measured (blue-solid) and modeled (red-dashed) temperature fields at (a) 12:00 UT and (b) 13:00 UT in the top panel, and the vertical profiles of measured (blue-solid) and modeled (red-dashed) Brunt-Väisälä frequency squared at (c) 12:00 UT and (d) 13:00 UT in the bottom panel.

convectively unstable region at 85–86 km and the dynamically unstable region at 93 km. The stable region between 86 and 93 km in the \( N^2 \) profile with the maximum of \( 10.37 \times 10^{-4} \, \text{s}^{-2} \) at 89 km shows the thermal ducting region of \( \sim 7 \) km width for propagation of the first wave event. As shown in Figure 7.9d, the negative \( N^2 \) of \( 0.61 \times 10^{-4} \, \text{s}^{-2} \) (\( N^2 < 0 \)) at 84 km and the small positive \( N^2 \) of \( 1.49 \times 10^{-4} \, \text{s}^{-2} \) (\( N^2 > 0 \)) at 93 km, indicate the convectively unstable region at 84 km, and the dynamically unstable region at 93 km. The stable region between 84 and 93 km of \( \sim 9 \) km width in \( N^2 \) profile with a maximum positive \( N^2 \) of \( 9.68 \times 10^{-4} \, \text{s}^{-2} \) at 89 km.
provides the thermal ducting region for propagation of the second wave event.

Figure 7.10 depicts the profiles of the Richardson number, $R_i$, derived from $N^2$ and total vertical wind shears using equation (7.6). As shown in Figure 7.10a, the $R_i$ values of $-0.03$ at 85 km and $\sim 0$ at 86 km and 0.23 at 93 km clearly indicate the convective instability region at 85–86 km and dynamic instability at 93 km, leading to a dynamically stable region between 86 and 93 km. The $R_i$ values of $-5.4$ and 0.24 also clearly indicate the convective and dynamic instabilities at 84 and 93 km as shown in Figure 7.10b, leading to a dynamically stable region at 85–92 km.

Figure 7.11 shows height profiles of the background winds ($u_0$) in the direction of wave propagation derived from the equation (7.3) using lidar wind data for the bore events observed at 12:00 and 13:00 UT for the night of October 21, 2003. The profile as shown in Figure 7.11a exhibits maximum background wind of 96 ms$^{-1}$ at 91 km in the direction of the first wave propagation at 12:00 UT. The maximum background wind of 92 ms$^{-1}$ at 90 km as shown in Figure 7.11b appeared in the opposite direction of the second wave propagation at 13:00 UT. It has been pointed out that the duct may be generated by some combination of wind shear and a temperature inversion layer [e.g., Dewan and Picard, 2001]. There is no doubt that the wind could play

Figure 7.10. Plots showing the profiles of the Richardson number, $R_i$, (a) at 12:00 UT for the first event, and (b) at 13:00 UT for the second event of October 21, 2003.
an important role in the formation of duct for wave front propagation. However, strong background winds used in the dispersion relation without curvature and shear terms given in equation (7.2) were not valid in forming the ducting conditions for the wave events observed on October 21, 2003. Since phase speeds of wave events were $\sim 18$ and $\sim 20$ ms$^{-1}$, the curvature and shear terms of strong background winds applied in the dispersion relation proved again not valid in the formation of ducts. This shows that the dispersion relation was not conducive to considering both wind and temperature gradients to infer the conditions of ducting. It is a fact that the thermal duct would support the occurrence of mesospheric fronts with the existence of temperature inversion layers [e.g., Smith et al., 2003; She et al., 2004]. Gravity wave interaction with the mean wind at critical level may also provide an inversion layer which supports a thermal duct for mesospheric bore [Dewan and Picard, 1998].

Figure 7.12 shows time-altitude contour plots of the temperature and Brunt-Väisälä frequency squared on the night of October 21, 2003. Figure 7.12a shows the downward phase propagation from about $\sim 100$ to $\sim 87$ km over $\sim 8$ hours, indicating the region of inversion layer of enhanced temperatures. This suggests that the
Figure 7.12. Contour plots for October 21, 2003 thermal-ducted bore events depict (a) the measured temperature field, and (b) the calculated $N^2$ derived from the measured temperature gradient.

upward propagating waves and the possibility of ter-diurnal tide (or a link between gravity wave and the tidal structure), causing the downward propagation trend of the temperature inversion layer. It has been noted that the gravity wave-tidal interaction is believed to be the cause for temperature inversions in the mesosphere [Whiteway and Carswell, 1995; Meriwether et al., 1998; Liu and Hagan, 1998; Liu et al., 2000; Williams et al., 2002]. As shown in the contour plot of $N^2$ in Figure 7.12b, the maximum positive $N^2$ value was also progressing down to 85 km from 98 km over 8 hours with the inversion layer, suggesting the dynamically stable region in order to form stable ducts for bore propagation in horizontal distance.

Figure 7.13 shows vertical profiles of zonal and meridional winds for the night of October 21, 2003. The profiles are spaced at 15-minute intervals, which corresponds to 50 ms$^{-1}$ while the vertical lines represent 0 ms$^{-1}$ for wind. The wind profiles, mainly the meridional winds in Figure 7.13b, show the downward phase propagation from about 93 to 88 km over 8 hours, suggesting the upward propagating waves and the possibility of ter-diurnal tide. The mean wave amplitude of meridional wind is larger than the mean amplitude of zonal wind (Figure 7.13a).
Figure 7.13. Plots showing the vertical profiles of (a) zonal wind, and (b) meridional wind for the night of October 21, 2003.

7.5 Summary and Conclusions

Simultaneous airglow image measurements in the OH and O₂ emissions from a mesosphere temperature mapper (MTM) have been used to investigate the horizontal characteristics of 64 short-period frontal events. Bore-like events observed over a two-year period (2003–2004) exhibit an average horizontal wavelength of ∼21 km, phase speed of ∼46 ms⁻¹, and observed periods of ∼7.6 min. Most of the frontal events (84%) exhibit the complementary effects in airglow layers as bright front in OH and dark front in O₂, and dark front in OH and dark front in O₂ sequences. This indicates that the ducting condition for wave propagation was to be formed between the OH and O₂ layers or around the OH layer.

Coincident meteor radar wind data have been used to investigate the Doppler ducting condition of undular bore. The July 3 bore event was clearly ducted in the airglow region. Observations strongly indicate that the wave event was trapped in a well defined ∼10 km-wide stable Doppler duct arising from wind structure. The event observed in the low-latitude mesopause region illustrates typical internal undular bore characteristics such as sharp bright (OH) and dark (O₂) leading fronts with a growing number of trailing wave crests without phase lag.
Two bores with different scale sizes observed on the night of October 21 were trapped within thermal inversion layers and propagating oppositely in ducting regions. The thermal ducting conditions of reverse bores were investigated using Na lidar wind-temperature data. The lidar data confirmed the existence of thermal inversion layers, which provided stable ducts for bores propagation along horizontal distance. The temperature inversion as well as maximum wind speed likely correlated with a long-period wave such as a ter-diurnal tide. These events illustrate sharp bright leading fronts in the OH layer and dark leading fronts in the O$_2$ layer.
CHAPTER 8
CONCLUSIONS AND DIRECTION OF FUTURE WORK

8.1 Conclusions

The Earth’s atmosphere is full of waves, not just sound waves and light waves but much larger-scale acoustic-gravity waves. These waves are generated by strong weather disturbances in the lower atmosphere, such as thunderstorms, convective or orographic force, and these waves propagate energy upwards into the upper mesosphere and lower thermosphere where they break, like waves on beach, depositing their momentum and causing large changes in upper atmospheric winds and temperatures. My dissertation research has involved a detailed investigation of short-period gravity wave events, and much smaller-scale dynamical features, called ripples, and has utilized both observational and theoretical data analyses.

Recent coordinated measurements from Haleakala Observatory, Maui, Hawaii (20.7° N, 156.3° W) by the USU Mesospheric Temperature Mapper (MTM), and UIUC Na wind-temperature lidar and meteor radar have provided a unique dataset for the study of dynamical features in the mesopause region. Airglow imaging observations have been analyzed to investigate characteristics of small-scale gravity waves and spatially localized short-lived ripple structures simultaneously in OH and O₂ emission layers centered at \( \sim 87 \) and \( \sim 94 \) km, respectively. The nature of ripple structures and associated instabilities in the mesopause regions have been analyzed using the Na lidar wind-temperature data of 100 meters-15 minutes spatial-temporal resolution on the selected nights. The meteor radar datasets of continuous hourly wind over the altitude range of 80–100 km were utilized for case studies of short-period gravity waves and their propagation characteristics.

The primary goals were to investigate the frequency of occurrence and cli-
matology of the ripples, and their seasonal variation in propagation direction in the mesopause region at low-latitude. A total of the 488 events on 223 clear image nights of ripple manifestation were recorded in OH and O$_2$ airglow images during 2003–2004. These short-lived small-scale events have been analyzed including their observed spatial and temporal characteristics such as horizontal wavelengths of $\sim 6$–18 km, horizontal speeds of $\sim 20$–50 ms$^{-1}$, and observed periods of $\sim 5$–10 min., but these parameters were not found to exhibit seasonal variations. However, the seasonal distribution of ripples was highly anisotropic in propagation direction, heading towards the northeast in summer and towards the northwest-southwest in winter. In addition, the directions of mono-chromatic gravity wave events were observed to propagate preferentially towards the northeast in summer and the northwest-southwest in winter. The measurements of anisotropy in the propagation direction of ripples associated with gravity wave propagation suggests the possible effect of variations in the source distribution for generating the ripples due to the passage of gravity wave in the mesopause region.

Simultaneous measurements of the airglow imager and Na wind-temperature lidar were made to identify individual Kelvin-Helmholtz (K-H) instability as a source of a ripple event observed in OH and O$_2$ emission layers, which compared well with theoretical computations. The derivations of static stability parameter, $N^2$, and instability parameter, $Ri$, from temperature and horizontal wind gradients have been used to characterize the nature of the ripples, and hence to study their alignment and association with dynamic or K-H instabilities. The ripple events were found to be induced by K-H instabilities that were caused by oppositely aligned wind shears in two airglow layers, and the wind shear vectors were nearly orthogonal to the phase fronts of observed ripples, suggesting the K-H instability signatures. On the other hand, the measurements of static stability and instability parameters significantly
showed that the strong wind shear was not always necessary to lead to the dynamically unstable region, but that the moderate wind shear ($< 40 \text{ ms}^{-1} \text{ km}^{-1}$) associated with small static stability, $N^2$, depends on temperature gradient, which could lower the $Ri$ number to less than 0.25, leading to dynamic or K-H instability. For case studies, the observation data were then utilized for numerical model simulation to examine apparent simultaneous occurrences of Kelvin-Helmholtz (K-H) instability due to dual shear in two airglow layers. Model results indicated that two sets of K-H billows that develop at one altitude were able to influence the evolution of the billows at the other altitude.

In this dissertation, horizontal parameters of small-scale ducted and evanescent gravity waves and their propagation characteristics have been investigated simultaneously in OH and O$_2$ airglow layers during 2003–2004. Typical observed wave parameters of the events were horizontal wavelengths of $\sim 15$–$20$ km, horizontal phase speeds of $\sim 25$–50 m s$^{-1}$, and observed periods of $\sim 7$–$12$ min. The meteor radar measurements of hourly winds have been used to investigate the intrinsic properties of these waves. Two wave events exhibiting similar airglow signatures were selected for analytical and numerical case studies. The wind effects present in our analysis of two events were significant. The observations clearly indicated that one of the events was a strongly Doppler-ducted wave due to variation in the background wind, and the other event of similar wave characteristics was a purely evanescent wave in nature due to the presence of large negative background wind in the mesopause airglow region. An analytical model based on the observation parameters was applied to estimate the ducting and evanescence characteristics of the two wave events. Then these novel results were compared with the numerical model simulations for the Doppler-ducted and evanescent waves. The model analysis confirmed that the measured ducted and evanescent wave events in OH and O$_2$ airglow layers were consistent with theoretical
Doppler-ducted and evanescent wave characteristics.

Simultaneous airglow image measurements in OH and O$_2$ emissions were used to investigate the frequency of occurrence of mesospheric bore-like events, called spectacular gravity wave events, and their horizontal characteristics. The mean horizontal parameters of bore-like events observed during 2003–2004 included wavelength of $\sim 21$ km, apparent phase speed of $\sim 46$ ms$^{-1}$, and observed period of $\sim 7.6$ min. The most prominent characteristics of the mesospheric bores were the sharp leading edge and coherent wave packet that grows with time. The leading edge often induces significant perturbation in airglow intensity and temperature. One characteristic of the front is the contrast reversal between the OH emission and the O$_2$ emission, which is supposed to be the response of an adiabatic temperature change due to the vertical displacement of the airglow layers by the passage of a disturbance in a horizontal wave guide.

Further, a stable ducting region is required for the internal mesospheric bore to propagate horizontally without significant attenuation. We have reported an internal undular mesospheric bore having a sharp bright leading front in the OH emission and a dark leading front in the O$_2$ emission, which was trapped in a well-defined stable Doppler duct arising from wind structure. Coincident meteor radar hourly wind data were used to investigate the Doppler ducting condition of the undular bore. We have also reported very typical mesospheric bores propagated horizontally in opposite directions and trapped in stable thermal ducts. These events exhibited a sharp bright leading front in the OH emission and a sharp dark leading front in the O$_2$ emission. The Na lidar wind and temperature data have been used to analyze the thermal ducting condition of the reverse bores as results of thermal inversion layers that provided stable ducts for waves propagation. In addition, the analytical models were separately consistent with the measured Doppler ducting and
thermal ducting characteristics of the observed bore events.

In summary, small-scale gravity waves and instability events were significantly present perturbations in the upper mesosphere and lower thermosphere. The measurements of their seasonal properties and propagation characteristics in this dissertation provided key data for modeling their global-scale impact on the upper atmospheric dynamics, leading to a better knowledge of the whole atmosphere meteorology. The airglow imaging observations have highlighted a novel investigation of the climatology of ripples and identified strong seasonal anisotropy in their directions of motion, suggesting strong ties with gravity waves. Simultaneous measurements from the airglow imager and Na wind-temperature lidar, and a comparison of these results with numerical models provided apparent simultaneous occurrences of K-H instabilities caused by dual shear in two airglow layers. Analysis of small-scale ducted gravity waves and mesospheric bores using comprehensive measurements of airglow image, wind, and temperature quantified conditions for Doppler and thermal ducting in the mesopause region. These important results, compared with analytical and numerical models, validated the linear gravity wave ducting theory. Furthermore, the airglow imaging study in this dissertation suggested the difference between characteristics of frontal-type and band-type gravity waves. The analysis of dispersion characteristics and intrinsic properties of these waves contributes to better understanding their importance for momentum and energy transport in the upper atmosphere.

8.2 Future Work

The results reported in this dissertation have underlined some promising areas and raised the numerous issues for further research. First, an airglow imaging analysis could be done to better quantify the gravity wave signatures including mountain waves and their seasonal wave anisotropy and effects on mesosphere and lower
thermosphere (MLT) regions. A possible cause of this anisotropy by critical level filtering of the gravity waves due to the intervening background wind fields could be analyzed using radar and lidar wind data including model winds.

Second, utilizing the measurements of local background wind fields, momentum and energy fluxes carried by vertically propagating gravity waves, particularly to upper mesospheric altitudes and hence their propagation nature could be derived. The dissipation of gravity waves can occur as they reach a critical level of the background wind, which will cause the variation of momentum flux that leads to acceleration of the local mean flow. In order to study the gravity wave dissipation, the magnitudes of acceleration of mean winds induced by small-scale gravity waves associated with variations of local wind and momentum flux could also be analyzed.

Another area of research would be to determine time-spatial gravity wave parameters such as intrinsic frequency, horizontal and vertical wave numbers, and phase and group velocities using wavelet transform analysis. The derivation of gravity wave parameters is based on wind and temperature perturbations, which could be estimated from the wind and temperature measurements by radar and Na lidar in the mesosphere. These computed perturbation parameters would then be utilized to calculate intrinsic parameters of waves. In addition, further work involving mathematical models could be a detailed analysis of ducting conditions of small-scale gravity waves and mesospheric bores using airglow imager and radar wind datasets.
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APPENDIX A
LINEAR GRAVITY WAVE EQUATIONS

The Taylor–Goldstein equation is a wave equation for linear gravity waves. Let us consider the two-dimensional Euler equations for irrotational and inviscid flow. Under the Boussinesq approximation the basic equations for two-dimensional motion of an incompressible and stratified atmosphere can be written as [Holton, 1992; Nappo, 2002]:

The momentum equations in the x-z plane:

\[
\frac{\partial u}{\partial t} + u \frac{\partial u}{\partial x} + w \frac{\partial u}{\partial z} = -\frac{1}{\rho} \frac{\partial p}{\partial x}, \tag{1}
\]

\[
\frac{\partial w}{\partial t} + u \frac{\partial w}{\partial x} + w \frac{\partial w}{\partial z} = -\frac{1}{\rho} \frac{\partial p}{\partial z} - g. \tag{2}
\]

The mass continuity equation:

\[
\frac{\partial u}{\partial x} + \frac{\partial w}{\partial z} = 0. \tag{3}
\]

The thermodynamic energy equation:

\[
\frac{\partial \rho}{\partial t} + u \frac{\partial \rho}{\partial x} + w \frac{\partial \rho}{\partial z} = 0. \tag{4}
\]

We linearize the above equations according to

\[
(u, w, \rho, p) = (u_0, w_0, \rho_0, p_0) + (u', w', \rho', p'), \tag{5}
\]

where \(u_0, w_0\) and \(\rho_0\) are steady horizontally uniform background values and \(u', w'\) and \(\rho'\) are first-order perturbations. Assuming the background flow is in hydrostatic balance, the hydrostatic approximation is

\[
\frac{\partial p}{\partial z} = -\rho g. \tag{6}
\]
Then momentum, continuity and energy equations (1-4) become

\[
\frac{\partial u'}{\partial t} + u_0 \frac{\partial u'}{\partial x} + w' \frac{\partial u_0}{\partial z} = -\frac{1}{\rho_0} \frac{\partial p'}{\partial x},
\]

(7)

\[
\frac{\partial w'}{\partial t} + u_0 \frac{\partial w'}{\partial x} = -\frac{1}{\rho_0} \frac{\partial p'}{\partial x} - \frac{\rho'}{\rho_0} g,
\]

(8)

\[
\frac{\partial u'}{\partial x} + \frac{\partial w'}{\partial z} = 0,
\]

(9)

\[
\frac{\partial \rho'}{\partial t} + u_0 \frac{\partial \rho'}{\partial x} + w' \frac{\partial \rho_0}{\partial z} = 0.
\]

(10)

We can assume simple time harmonic plane-wave solutions of the form:

\[
(u', w', \rho', p') = (\hat{u}, \hat{w}, \hat{\rho}, \hat{p}) \exp \left(i(kx - \omega t)\right),
\]

(11)

where \(k\) is the wave number and \(\omega\) is the angular frequency. Equations (7-10) become

\[
-i\omega \hat{u} + iu_0 k \hat{u} + \hat{w} \frac{du_0}{dz} = -\frac{ik}{\rho_0} \hat{p},
\]

(12)

\[
-i\omega \hat{w} + iu_0 k \hat{w} = -1 \frac{d\hat{p}}{\rho_0 dz} - \frac{\hat{\rho}}{\rho_0} g,
\]

(13)

\[
i \hat{k} \hat{u} + \frac{d\hat{w}}{dz} = 0,
\]

(14)

\[
-i\omega \hat{\rho} + iu_0 k \hat{\rho} + \hat{w} \frac{d\rho_0}{dz} = 0.
\]

(15)

The intrinsic frequency, \(\Omega\), is defined as the frequency of a wave relative to the flow, i.e., the frequency of a wave measured by an observer drifting with the fluid at speed \(u_0\). Therefore, the Doppler-shifted intrinsic wave frequency [Chimonas and Hines, 1986] is

\[
\Omega = \omega - u_0 k;
\]

(16)

where \(\omega\) is the apparent wave frequency observed in a fixed coordinate system and \(u_0\) is the component of the background wind in the direction of wave propagation.

Using \(N^2 = -\frac{g}{\rho_0} \frac{d\rho_0}{dz}\) and equation (16), the equations (12-15) become
\( i\Omega \hat{u} - \hat{w} \frac{du_0}{dz} = \frac{i}{\rho_0} k \hat{p}, \quad (17) \)

\( i\Omega \hat{w} = \frac{1}{\rho_0} \frac{d\hat{p}}{dz} + \frac{\hat{p}}{\rho_0} g, \quad (18) \)

\( ik \hat{u} + \frac{d\hat{w}}{dz} = 0, \quad (19) \)

\( i\Omega \hat{p} + \frac{\hat{w} \rho_0}{g} N^2 = 0. \quad (20) \)

Solving the polarization equations (17–20) for \( \hat{w} \) gives

\[
\frac{d^2 \hat{w}}{dz^2} + \frac{1}{\rho_0} \frac{d\hat{w}}{dz} + \left[ \frac{k^2 N^2}{\Omega^2} + \frac{k}{\Omega} \frac{d^2 u_0}{dz^2} - \frac{k}{\Omega} \frac{1}{H} \frac{du_0}{dz} - k^2 \right] \hat{w} = 0. \quad (21)
\]

The vertical variation of density in an isothermal atmosphere is \( \rho = \rho_0 e^{-\frac{z}{H}} \) or

\[
\frac{1}{\rho} \frac{d\rho}{dz} = -\frac{1}{H}, \quad (22)
\]

where \( H \) is the scale height. Equation (21) then becomes

\[
\frac{d^2 \hat{w}}{dz^2} - \frac{1}{H} \frac{d\hat{w}}{dz} + \left[ \frac{k^2 N^2}{\Omega^2} + \frac{k}{\Omega} \frac{d^2 u_0}{dz^2} - \frac{k}{\Omega} \frac{1}{H} \frac{du_0}{dz} - k^2 \right] \hat{w} = 0. \quad (23)
\]

Simplifying this equation by defining a new variable, \( w_z \), by

\( \hat{w} = e^{\frac{\pi}{4} N^2} w_z. \quad (24) \)

Substituting of \( \hat{w} \) into equation (23) leads to the Taylor-Goldstein equation as:

\[
\frac{d^2 w_z}{dz^2} + \left[ \frac{k^2 N^2}{(c - u_0)^2} + \frac{1}{(c - u_0)^2} \frac{d^2 u_0}{dz^2} - \frac{1}{H(c - u_0)} \frac{du_0}{dz} - k^2 - \frac{1}{4H^2} \right] w_z = 0. \quad (25)
\]

Using \( \Omega = \omega - u_0 k = (c - u_0) k \), the Taylor-Goldstein equation then takes the form

\[
\frac{d^2 w_z}{dz^2} + \left[ \frac{N^2}{(c - u_0)^2} + \frac{1}{(c - u_0)^2} \frac{d^2 u_0}{dz^2} - \frac{1}{H(c - u_0)} \frac{du_0}{dz} - k^2 - \frac{1}{4H^2} \right] w_z = 0, \quad (26)
\]

where \( w_z \) is vertical velocity perturbation component, \( N \) is Brunt-Väisälä frequency, \( c \) is apparent phase speed, \( u_0 \) is background wind in the direction of wave motion.
APPENDIX B

NAVIER-STOKES EQUATIONS

The equations for direct numerical simulations (DNS) of instability growth and evolution in three dimensions [Werne and Fritts, 2001; Fritts et al., 2009a,b; Laughman et al., 2009] are the non-dimensional incompressible Navier-Stokes equations subject to the Boussinesq approximation, which are given as

\[
\frac{\partial}{\partial t} \vec{u} + (\vec{u} \cdot \vec{\nabla}) \vec{u} = \nu \vec{\nabla}^2 \vec{u} - \vec{\nabla} \left( \frac{P}{\rho_0} \right) + \left( \frac{\rho}{\rho_0} \right) \vec{g}
\] (27)

\[
\frac{\partial}{\partial t} \theta + (\vec{u} \cdot \vec{\nabla}) \theta = \kappa \vec{\nabla}^2 \theta
\] (28)

\[
\vec{\nabla} \cdot \vec{u} = 0
\] (29)

\[
\rho = \rho_0 [1 - \alpha (\theta - \theta_0)]
\] (30)

where \(\vec{u} = (u, v, w)\) is the full 3-D velocity field, \(P\) is the pressure field, \(\rho\) is the density, \(\theta\) is the potential temperature, and \(g\) is gravity. Subscripts denote mean values except in the time derivative terms. The terms \(\nu\), \(\kappa\), and \(\alpha\) are kinematic viscosity, thermal diffusivity, and the coefficient of thermal expansion equal to \(1/\bar{\theta}_0\). The potential temperature scale with a length scale \((h)\) and peak static stability \((N_0^2)\) is \(\theta_0 = h N_0^2/g \alpha\).
CURRICULUM VITAE

Deepak B. Simkhada
Dhawa-1, Gorkha, Nepal
dbsimkhada@gmail.com

Professional Preparation:
Utah State University, Department of Physics, Physics, Ph.D., 2010

Professional Experience:
Graduate Research Assistant, Center for Atmospheric and Space Sciences, Utah State University, Logan, Utah, 2004–2010
Graduate Teaching Assistant, Department of Physics, Utah State University, Logan, Utah, 2003–2010

Research Collaborations:
Michael J. Taylor and Jonathan B. Snively (Utah State University); Dave C. Fritts and Brian Laughman (NorthWest Research Associates, CoRA Division); Steve J. Franke, Alan Z. Liu and Gary R. Swenson (University of Illinois at Urbana-Champaign)

Professional Affiliations:
Member of American Geophysical Union (AGU)
Member of American Physical Society (APS)
Life Member of Nepal Physical Society (NPS)