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USING A BISTATIC FABRY-PEROT INTERFEROMETER SYSTEM
TO IMPROVE THE ESTIMATION OF THERMOSPHERIC WINDS

BY

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THESIS

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ABSTRACT

One significant cause of disruption in the propagation of low-power, trans-atmospheric radio waves is equatorial spread F (ESF). The Rayleigh-Taylor instability (RTI) causes the development of many types of ESF events. It has been theorized that thermospheric winds have a substantial effect on the growth rate of the RTI, but very little data has been gathered to verify or disprove this idea.

This thesis proposes a method for synchronously using two ground-based Fabry-Perot interferometers (FPI) to estimate thermospheric winds. When a single FPI is used, assumptions must be made to obtain meaningful wind data, but there are inherent limitations to the usefulness of the data caused by the use of these assumptions. It must be assumed that either the vertical wind speed is zero or the vertical and horizontal winds are uniform for the entire observation area in order to obtain meaningful wind speed estimates. In addition, data from a single FPI cannot be used to identify complex structures in the wind field. However, a bistatic FPI system (BFPIS) can be used to obtain full vector wind information in two or more locations. This observation methodology also has potential for reducing the number of assumptions that must be made. For a BFPIS, it need only be assumed that the vertical wind is uniform for the observation area.

We begin with a review of the physical principles and typical methods for using an FPI to estimate thermospheric winds and temperatures. We then propose a method for using a BFPIS to improve collected data, as compared to using a single FPI, and present simulations that validate this method.

To my parents, for their love and support

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TABLE OF CONTENTS

LIST OF ABBREVIATIONS	vi
CHAPTER 1 INTRODUCTION	1
CHAPTER 2 THE EARTH'S ATMOSPHERE	4
2.1 Basic Composition	4
2.2 Equatorial Irregularities	6
2.3 630.0-nm Redline Emission	10
CHAPTER 3 THE FABRY-PEROT INTERFEROMETER	15
3.1 History	15
3.2 Physical Principles	16
3.3 Etalon Design	18
3.4 MiniME	24
CHAPTER 4 OBSERVATION AND ANALYSIS	28
4.1 Single FPI Observation Techniques	28
4.2 Image Analysis	31
4.3 Observation Limitations	35
CHAPTER 5 BISTATIC FABRY-PEROT INTERFEROMETER OBSERVATION STRATEGY	38
5.1 Advantages of BFPIS	38
5.2 Developing a BFPIS Observation Technique	43
5.3 Extracting Full Wind Vectors	50
5.4 Simulation Details	56
5.5 Simulation Results	58
CHAPTER 6 CONCLUSION	65
6.1 Conclusions	65
6.2 Future Work	66
REFERENCES	68

LIST OF ABBREVIATIONS

ADU	analog-to-digital unit
AG	airglow
BFPI	bistatic Fabry-Perot interferometer system
CCD	charge-coupled device
CoS	center of sheer
ECEF	Earth-Centered, Earth-Fixed
EPB	equatorial plasma bubble
ESF	equatorial spread F
FOV	field of view
FPI	Fabry-Perot interferometer
FSR	free spectral range
GPS	Global Positioning System
LT	local time
LM NLS	Levenberg-Marquardt nonlinear least squares
MC	Monte Carlo
RENOIR	Remote Equatorial Nighttime Observatory of Ionospheric Regions
RTI	Rayleigh-Taylor instability
TI	thermosphere/ionosphere
UT	universal time
UV	ultraviolet

VEN	Vertical, East, North
WGS-84	World Geodetic System 1984

CHAPTER 1

INTRODUCTION

The increasingly common use of satellite radio communications for civilian, corporate, and military applications is driving a demand for more reliable trans-atmospheric communication. Irregularities in the atmosphere form frequently and can cause serious disruptions in the propagation of electromagnetic communication signals, affecting all kinds of low-powered radio transmissions. Thus, a significant threat to the reliability of satellite communications is the disruption of radio wave propagation in the atmosphere.

The Remote Equatorial Nighttime Observatory of Ionospheric Regions (RENOIR) project studies interactions between ionospheric and thermospheric dynamics that can impact satellite-based communications. One major cause of disruptions in radio wave propagation is equatorial spread F (ESF), a phenomenon characterized by plasma density irregularities that form in equatorial regions and spread polewards. It is closely connected to the formation of equatorial plasma bubbles (EPB), or plasma depletions, that develop at the magnetic equator, drift eastwards, and expand along magnetic field lines towards the poles [1, 2]. It is commonly held that ESF is caused by the Rayleigh-Taylor instability (RTI) [3], and it has been theorized that thermospheric winds can have a significant effect on the growth rate of the RTI and subsequent development of ESF events, but little data exists to confirm or dispute this idea [4, 5].

The RENOIR project uses a variety of equipment to simultaneously observe ionospheric and thermospheric activity, including Global Positioning Systems (GPS) scintillation monitors, dual-frequency GPS receivers, all-sky imaging systems, and Fabry-Perot interferometers. The Fabry-Perot Interferometer (FPI) is a passive optical sensing instrument used to estimate thermospheric winds and temperatures. It observes naturally occurring light emissions in a small region of the thermosphere and uses interferometry to precisely measure the frequency of the incoming light. When measurements

are made in multiple directions, the Doppler effect can be used to estimate the speed and direction of the atoms that emitted the light. In this way the neutral wind speeds can be estimated.

When a single FPI is used, it must be assumed that the wind field of the entire observation area is uniform in order to determine full wind vector information. For typical observation parameters, this includes a region on the order of 500 km by 500 km, making the uniformity assumption questionable. In addition to the fact that this assumption introduces error into the resulting wind estimates, it means that it is impossible to observe smaller scale fluctuations in wind patterns that could be associated with the passage of thermospheric gravity waves or similar atmospheric phenomena.

The focus of this thesis is to describe how to use a bistatic configuration of two FPIs to determine thermospheric winds more precisely and obtain full vector wind information at multiple locations. When two FPIs are used synchronously, the additional data improve the accuracy and precision of wind estimates. This improves accuracy by reducing the uniform wind field assumption to that of a uniform vertical wind over the observation area. The use of two FPIs improves the precision of each horizontal measurement by determining the complete wind vector using observations of a single volume from two directions instead of averaging the results from observations of different volumes. In addition, the full wind vector can be determined at two (or more) distinct sites, improving detection of wind fluctuations.

Chapter 2 of this thesis discusses the basic composition of the atmosphere, important irregularities, and the causes of the 630.0-nm redline emission that the FPI observes. Chapter 3 gives a brief history of the development of the FPI, explains the basic design of an FPI and some of the considerations for selecting design parameters, and describes the details of the particular FPI that the RENOIR project uses. Chapter 4 describes a typical method for applying an FPI to the estimation of thermospheric winds. It outlines a typical observation technique, the image analysis method used for the RENOIR project, and the inherent limitations of FPI data. Chapter 5 explains the advantages of a bistatic network of FPIs, describes the process of developing an observation strategy for a bistatic Fabry-Perot interferometer system (BFPIS), and derives the equations used to extract full wind vectors from the BFPIS data. It further explains the simulations we have run to verify this method and presents the conclusions from the results of this work. Chapter 6

briefly discuss the conclusions of this thesis and future research possibilities for this project.

CHAPTER 2

THE EARTH'S ATMOSPHERE

This chapter briefly describes the layers of the atmosphere, discusses equatorial irregularities that affect radio wave propagation, and explains the chemical process that produces the 630.0-nm redline emission that the RENOIR FPI observes.

2.1 Basic Composition

The first distinction that is made when looking at the upper terrestrial atmosphere is the difference between neutral atoms and ionized atoms. Because of differences in dynamics, the two are usually referred to separately as the neutral atmosphere and the ionosphere. The neutral atmosphere and the ionosphere are further divided into sections by altitude. It is most convenient to divide the neutral atmosphere by temperature profile and the ionosphere by plasma density profile. Figure 2.1 shows a typical mid-latitude temperature profile for the neutral atmosphere and plasma density profile for the ionosphere.

We will be primarily looking at the thermosphere in the neutral atmosphere and the F region in the ionosphere. As seen in Figure 2.1, the lowest part of the atmosphere is termed the troposphere. Temperature falls with increasing altitude until about 10 km above the surface of the earth, where absorption of part of the ultraviolet spectrum begins to factor in. The region of increasing temperature (from 10 km to 50 km) is termed the stratosphere. The effect of absorbing ultraviolet (UV) rays maximizes at 50 km. The volume from 50–80 km, called the mesosphere, is where radiative cooling causes a sharp temperature decrease. The thermosphere consists of the part of the neutral atmosphere above the temperature minimum (the mesopause) located at approximately 80 km. Above 90 km the temperature increases dramatically

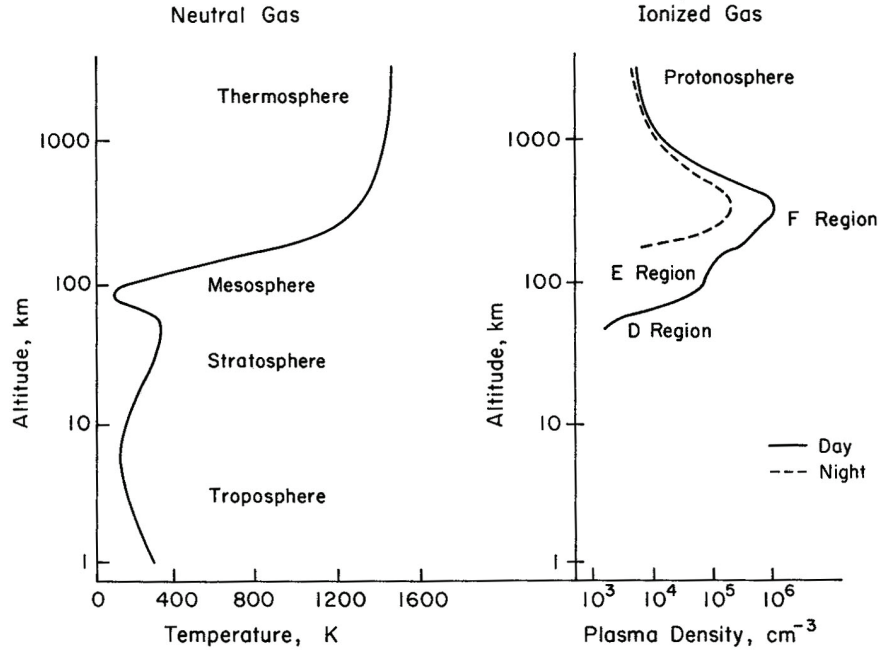


Figure 2.1: Typical profiles of neutral atmospheric temperature and ionospheric plasma density with each layer designated. After [3].

because of absorption of UV and extreme UV radiation from the sun.

In the ionosphere, the F Region denotes the altitude range from 150–500 km. The range 90–150 km is referred to as the E Region, and the section below the E Region is referred to as the D Region.

Above 100 km, particles start to separate according to their masses instead of being a fairly homogeneous mixture. In addition to causing higher temperatures in the neutral atmosphere, the absorbed extreme UV radiation also produces plasma. In the sunlit hemisphere, incoming photons reacting with neutrals causes the number density of incoming photons to decrease as altitude decreases. This, combined with the fact that neutral density increases exponentially as altitude decreases, produces the peak in plasma density seen in the right side of Figure 2.1. The daytime peak in plasma density occurs where the combination of decreasing photon density and increasing neutral density produces the most plasma. At night there are very few incoming photons, so as the sun sets the loss processes dominate. Since loss processes are faster in regions with more neutrals, the lower F layer recombines more quickly, which also causes the plasma density peak, termed the F peak, to shift upwards.

2.2 Equatorial Irregularities

This section describes how zonal neutral winds affect the F region dynamo and meridional neutral winds affect the Rayleigh-Taylor instability.

2.2.1 Equatorial F Region Dynamo

A fairly consistent, eastward, zonal wind has been observed near the equator at night [3]. This wind generates a vertical current and electric field, resulting in a phenomenon known as the equatorial F region dynamo. The effect of thermospheric winds on the motion of plasma is one of the primary reasons that thermospheric winds are of interest. In this section we explain the equatorial F region dynamo, with a particular focus on the contribution of thermospheric winds.

It can be shown [3] that the electric current in the F region can be written as

$$\mathbf{J} = \boldsymbol{\sigma} \cdot (\mathbf{E} + \mathbf{U} \times \mathbf{B}), \quad (2.1)$$

where $\boldsymbol{\sigma}$ is the conductivity tensor, $\mathbf{E}$ is the electric field, $\mathbf{U}$ is the neutral wind, and $\mathbf{B}$ is the magnetic field. We use $\{\hat{a}_x, \hat{a}_y, \hat{a}_z\}$ to denote vectors in the eastern, northern, and vertical directions, respectively. We use $\mathbf{U} = \{u, v, w\}$ to denote the winds in the eastern, northern, and vertical directions. With this notation, the conductivity tensor is [3]

$$\boldsymbol{\sigma} = \begin{pmatrix} \sigma_P & 0 & \sigma_H \\ 0 & \sigma_0 & 0 \\ -\sigma_H & 0 & \sigma_P \end{pmatrix}, \quad (2.2)$$

where σ_P is Pedersen conductivity, σ_H is Hall conductivity, and σ_0 is specific conductivity. In the nighttime F region, $\sigma_P \gg \sigma_H$ as shown in Figure 2.2, making the conductivity tensor essentially diagonal.

The meridional wind has no effect on the dynamo because it moves parallel to magnetic field lines, and the vertical wind is small enough to not have a significant effect. If we ignore the electric field for a moment, an electric

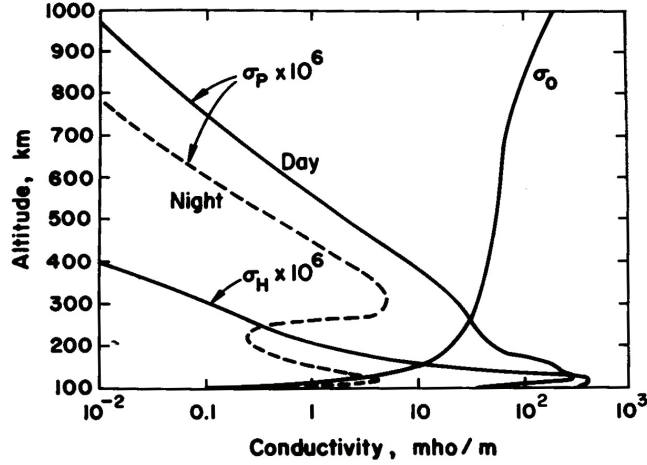


Figure 2.2: Typical conductivities for the mid-latitude daytime ionosphere. Observe that the scale for σ_H and σ_P is altered. After [6].

current will be generated by the zonal wind in the form

$$\mathbf{J} = \boldsymbol{\sigma} \cdot (u \hat{a}_x \times \mathbf{B}). \quad (2.3)$$

At the magnetic equator, the earth's magnetic field is horizontal:

$$\mathbf{B} = |\mathbf{B}| \hat{a}_y. \quad (2.4)$$

So electric current produced by the wind and magnetic field will be in the vertical direction:

$$J_z = \sigma_p u B. \quad (2.5)$$

Typically, the zonal wind will have only small fluctuations as a function of altitude. However, the nighttime Pedersen conductivity varies significantly with height, as can be seen in Figure 2.2. Therefore, the divergence of J_z does not equal zero. In order to satisfy Maxwell's equations, a vertical electric current must generate an electric field in the z direction, which in turn must produce an electric current in the y direction. However, at the equator it is a good approximation to assume that the magnetic field lines do not allow a magnetic field-aligned current to flow [3], so $J_y = 0$. This means that $E_z = 0$, and therefore J_z must also equal zero. We can further conclude that vertical current produced by the zonal neutral wind must be exactly canceled out by

a vertical current in the opposite direction generated by a vertical electric field. Therefore, we can return to Equation (2.1) and set it equal to zero. Simplifying this expression yields

$$E_z = -uB. \quad (2.6)$$

An electric field in the negative z direction, as is typical at nighttime, coupled with a magnetic field in the y direction produces a force in the x direction. This accounts for the nighttime eastward ion drift observed at the equator.

2.2.2 Equatorial Spread F

One major cause of disruption in radio signals is irregularities in the upper atmosphere. These irregularities are generated at the magnetic equator, spread polewards, and are generally termed equatorial spread F (ESF), which is strongest at the magnetic equator after local sunset. These irregularities can have an extremely detrimental effect on a wide band of radio signals that pass through the atmosphere. ESF has been studied since it was first observed by Booker and Wells in 1938 [7]. However, with the recent explosion of satellite communications, especially Global Navigation Satellite Systems such as the Global Positioning System (GPS), ESF has been studied with increased interest. Scientists have learned much about the growth mechanisms, seasonal variability, and longitudinal differences of ESF [8, 9, 10, 11, 12]. However, many things about this phenomenon remain unknown, in particular the day-to-day variability in ESF events and the underlying physical mechanisms that cause and strengthen their occurrences.

It is now commonly held that the mechanism that causes ESF events is the Rayleigh-Taylor instability (RTI) [3]. In post-sunset conditions, after the bottom side of the F layer has recombined and the entire F layer has vertically risen above its daytime altitudes because of the pre-reversal enhancement, there is a strong vertical density gradient that is made unstable by vertical perturbations in the ionosphere. While the cause of these perturbations is as of yet uncertain, the result is the formation of equatorial plasma bubbles (EPB). These bubbles are depletions in the plasma density that contain strong, small-scale electric field and electron density fluctuations within and along their edges. It is these perturbations that disrupt the propagation of

radio waves.

2.2.3 Neutral Wind Effects on RTI and ESF

Of the factors that contribute to the RTI, neutral winds are one of the least measured and may be one of the most significant contributing factors [4, 5]. We can write the growth rate of the generalized RTI as [3, 5]

$$\gamma_{RT} = \left(\frac{\Sigma_P^F}{\Sigma_P^{E,N} + \Sigma_P^F + \Sigma_P^{E,S}} \right) \left[\frac{\mathbf{E} \times \mathbf{B}}{|\mathbf{B}|^2} - \mathbf{U}_m - \frac{\mathbf{g}}{v_{in}} \right] \cdot \frac{\nabla N}{N}, \quad (2.7)$$

where Σ_P^F is the contribution of the field-line integrated Pedersen conductivity from the F region, $\Sigma_P^{E,N}$ and $\Sigma_P^{E,S}$ are the contribution from the northern and southern E regions, respectively, $\mathbf{E}$ is the electric field vector, $\mathbf{B}$ is the magnetic field vector, $\mathbf{U}_m$ is the meridional component of the neutral wind, $\mathbf{g}$ is gravity, v_{in} is the ion-neutral collision frequency, and $\frac{\nabla N}{N}$ represents the gradient in electron density. Sultan modeled the effects of neutral wind on the RTI growth rate, and found that trans-equatorial neutral wind can have three different effects on the RTI growth rate [5]. First, it contributes directly to the growth rate as $\mathbf{U}_m$ in Equation (2.7). Second, meridional wind will lift the F region in one hemisphere and lower the F region in the other hemisphere. This will alter the chemical concentrations and thereby change the effective recombination rates, which will reduce the effective flux tube growth rate. Lastly, meridional neutral wind will shift plasma from one hemisphere to the other, which produces a stabilizing affect. All three of these effects tend to be a damping force on RTI growth. Sultan concludes that neutral winds in the vertical and meridional directions are important parts of the RTI growth rate equation.

Very little data have been gathered to test Sultan's conclusions. Mendillo et al. [13] performed a study using imaging systems and an FPI in South America and concluded that meridional thermospheric wind has little affect on the day-to-day occurrence of EPBs. However, this study was based on eight nights of observation in a single month. The time period was also during a typical season of irregularity occurrences in South America (September 1998). Therefore, more data over a longer time span is needed to test Sul-

tan’s hypothesis. Also, virtually no data exist on measurements of vertical equatorial thermospheric winds.

2.3 630.0-nm Redline Emission

During the day, solar radiation is absorbed by the thermosphere causing plasma density to increase drastically. After the sun sets, the plasma recombines with free electrons. These chemical reactions cause a variety of airglow (AG) emissions; one observable example of this is the aurora borealis, also known as the northern lights. One AG that is particularly useful for studying the dynamics of the thermosphere is the 630.0-nm emission line. This emission is particularly useful for space science because the chemicals that produce it are in sufficient supply that it is bright enough to be observed by land-based optical instruments. In addition, the reactions that produce it occur slowly enough that the emission lasts the length of an entire night. The 630.0-nm AG has a variety of uses for gaining information about the thermosphere. In particular it can be observed by FPIs for estimating thermospheric winds and temperatures. We discuss the chemical process that produces the 630.0-nm emission, model the resulting intensity profile, and discuss the implications of the profile that affect the accuracy of a ground-based FPI. Rate coefficients for the relevant chemical reactions are shown in Table 2.1.

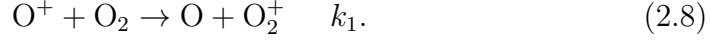
Table 2.1: Rate Coefficients Used to Determine Volume Emission Rate [14].

Coefficient	Rate ($\text{cm}^3 \cdot \text{s}^{-1}$)
α_1	$1.95 \times 10^{-7} \tau_e^{-0.7}$
k_1	$3.23 \times 10^{-12} e^{3.72/\tau_i - 1.87/\tau_i^2}$
k_3	$2.0 \times 10^{-11} e^{111.8/T_n}$
k_4	$2.9 \times 10^{-11} e^{67.5/T_n}$
k_5	$1.6 \times 10^{-12} T_e^{0.91}$
A_{1D}	6.81×10^{-3}
A_{6300}	5.15×10^{-3}

T_e is electron temperature; τ_e is $T_e/300$; τ_i is ion temperature divided by 300; T_n is neutral temperature.

2.3.1 Chemical Processes

The 630.0-nm redline emission is indirectly caused by O_2^+ in the atmosphere, which forms from charge exchange between O^+ and O_2 with rate coefficient k_1 [11, 14, 15]:



The primary cause of loss of O_2^+ is dissociative recombination, which can be written as



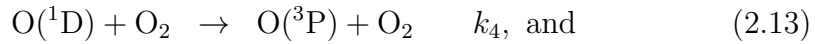
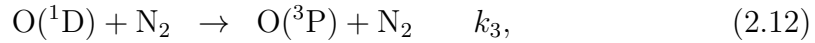
Using these two equations, the density of O_2^+ can be found by balancing the production and loss rates:

$$[O_2^+] = \frac{k_1[O_2][O^+]}{\alpha_1[e]}. \quad (2.10)$$

The (1D) atoms that can be produced by Equation (2.9) cause the 630.0-nm emission that we are interested in. The proportion of reactions that produces a $O(^1D)$ atom instead of $O(^3P)$ or $O(^1S)$ can be defined as β_{1D} . Using Equations (2.9) and (2.10) we find the production rate of $O(^1D)$ atoms in terms of O_2 and O^+ densities to be

$$\beta_{1D} \alpha_1 [O_2^+][e] = \beta_{1D} \alpha_1 \frac{k_1[O_2][O^+]}{\alpha_1[e]} [e] = \beta_{1D} k_1 [O_2][O^+]. \quad (2.11)$$

There are three significant reactions that contribute to the depletion of $O(^1D)$ [14, 16].



During the depletion of $O(^1D)$, photons can be emitted at three different wavelengths: 630.0 nm, 636.4 nm, and 639.2 nm [15, 16]. The Einstein coefficient, A_{1D} , identifies the rate at which a photon will be emitted at any one of these wavelengths. A second coefficient, A_{6300} , identifies the rate at

which an emitted photon will have a wavelength of 630.0 nm. The resulting volume emission rate in photons/cm³/s is

$$V_{6300} = \frac{A_{6300} \beta_{1D} k_1 [\text{O}_2] [\text{O}^+]}{A_{1D} + k_3 [\text{N}_2] + k_4 [\text{O}_2] + k_5 [\text{e}]} \quad (2.15)$$

2.3.2 Climatological Modeling

In order to gain a greater understanding of this emission, we simulate its volume emission rate using climatological models. Climatological models provide information about the thermosphere and ionosphere, such as number densities of a variety of species as a function of latitude, longitude, and altitude [17]. These models are constructed using data from multiple sources collected over many seasons and a variety of solar cycle conditions. Climatological models provide a good approximation of typical behavior of the thermosphere and ionosphere and therefore can be used to study the typical behavior of the 630.0-nm AG. To simulate the 630.0-nm emission we use the NRLMSISE-00 [18] and IRI-07 [19] models, which provide temperature and atom density information. The Naval Research Labs Mass Spectrometer, Incoherent Scatter (Radar) Extended-2000 (NRLMSISE-00) atmospheric model provides the densities of neutral molecules, [O₂] and [N₂]. The International Reference Ionosphere-2007 (IRI-07) provides electron densities and the temperatures of neutrals, ions, and electrons. The IRI now also has estimations of O⁺ densities, so those need not be approximated, as was done in [15]. The IRI estimates that O⁺ densities are very similar to electron densities at altitudes above 300 km and that small but important differences develop at lower altitudes. This result has good agreement with the calculations performed by Makela [15].

Figure 2.3 shows the electron density profile and 630.0-nm emission profile at 23:00 LT for a location in Paraiba, Brazil (7.126S, 37.543W) on 27 June. This day was chosen as representative of near solstice conditions. The profile was generated for the years 2002–2006 to show the variation from near solar maximum (2002), to near solar minimum (2006) conditions. The strongest concentrations of electrons and AG occurred in 2002, and concentrations drop in each progressive year, tracing the reduction of incoming solar flux.

The AG emission peak occurs around 30 km below the F peak. The dif-

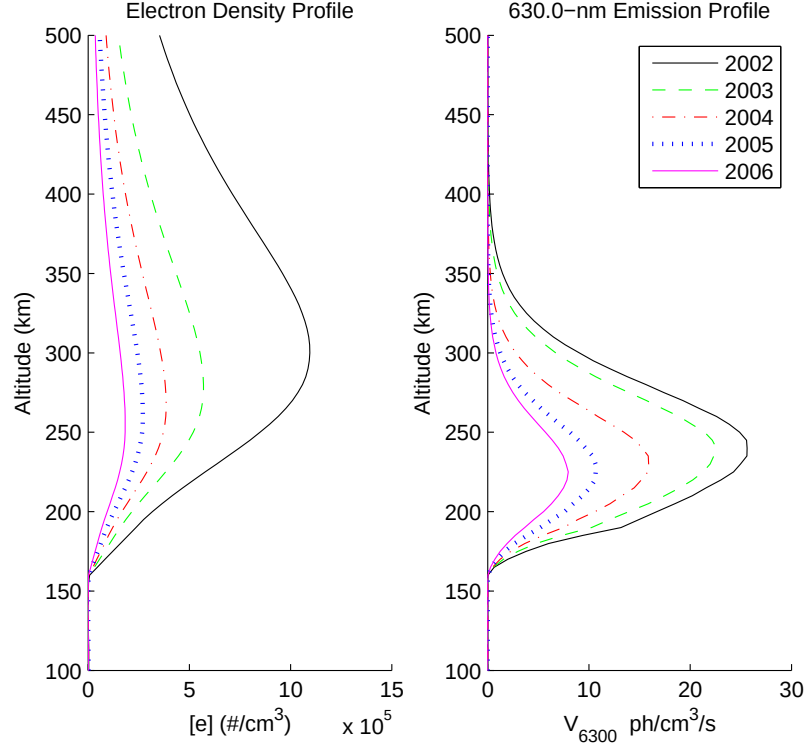


Figure 2.3: Profiles of electron density (left) and emission rate per volume of the 630-nm emission (right) are shown for 23:00 LT on 27 June 2002–2006. The highest concentrations and emission rates occur in 2002, the year near solar maximum, and each progressive year has lower concentrations and emission rates.

ference between the two is caused by the emission's dependency on neutrals. As Figure 2.4 shows, the concentration of neutrals is inversely proportional to altitude, so the AG emission peak is shifted below the F peak. Since free electrons contribute to the production of $O(^1D)$, electron density is an important factor in the intensity of the redline emission. However, since concentration of neutrals is also a factor, it follows that the altitude of the F peak matters as well. If the F peak rises, the overlap of neutrals and free electrons decreases, causing a decrease in the production of $O(^1D)$ and, therefore, of 630.0-nm photons. These factors can be observed in the year-to-year variability shown in Figure 2.3. Each year the electron density is smaller and the F peak drops. The reduction in electron density correlates to a reduction in AG intensity, but the reduction of AG intensity is proportionally less than the reduction of electron density. This is because each year the F peak also occurs at a lower altitude, where neutrals are more plentiful. Therefore, the

loss of electron density is partially offset by the lower F peak.

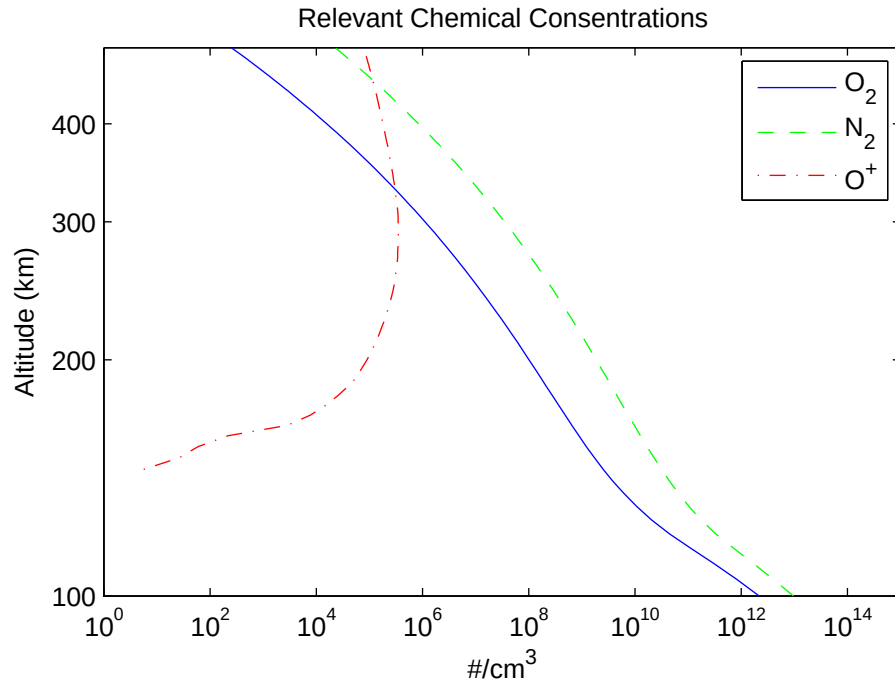


Figure 2.4: Typical densities of chemicals relevant to the 630.0-nm emission are plotted as functions of temperature as estimated by the NRLMCISE and IRI for 23:00 LT on 27 June 2004 in Pariaba, Brazil.

CHAPTER 3

THE FABRY-PEROT INTERFEROMETER

In this chapter we describe the typical function of a Fabry-Perot interferometer. We begin with a brief history of the development of the FPI. Then, we explain the basic design of an FPI, particularly the etalon, and the optical principles that it employs. Lastly, we discuss the details of the MiniME design and how it differs from a typical astronomical FPI.

3.1 History

In 1831 George Airy, a professor of astronomy at the University of Cambridge, published the second edition of *Mathematical Tracts*. To this edition he added a section proposing and defending the theory of undulations in optics [20]. Within this section, he also included the concept of optical interference and calculated several interference problems. This work led to the formula now known as the Airy function [20], which is the basis of much of the optical interferometry research done today. While many researchers studied the implications and applications of the Airy function, it was nearly seventy years later that Fabry and Perot published a series of papers on the application of the Airy function to a thin film of air between two glass plates. In a paper published in 1899, they provided a detailed description of what is now known as a scanning Fabry-Perot interferometer that could scan several orders of fringes [20]. In 1901 they published the first photograph of FPI interference lines: the interference rings of the green mercury line. Over the next several years, other scientists began experimenting with different types of FPIs. However, it was Fabry who, in 1911, first applied the use of the FPI to astronomy [20, 21]. By the 1920s, the FPI had become a well-established spectroscopic instrument. The FPI is still the best optical instrument available when observing faint emission lines [22] and has been used for decades

in atmospheric sciences for measuring atmospheric winds [23].

3.2 Physical Principles

Before we discuss the details of how the FPI works, it is necessary to explain a few physical principles that it employs. In this section, we describe wave interference, the Doppler shift, and Doppler broadening.

3.2.1 Wave Interference

Wave interference is the term used to describe the superposition of two or more waves. Interference occurs between rays coming from the same source of radiation that take different paths to reach the same destination. In this case, the rays are coherent, as opposed to rays from multiple sources, which are incoherent [21]. The intensity of incoherent waves adds directly, but the intensity of coherent waves adds or subtracts depending on the phase difference of the waves. Figure 3.1 shows constructive and destructively interfering waves. When superimposed on one another, the constructively interfering waves will illuminate their target more brightly than a single wave, while the destructively interfering waves cancel one another.

3.2.2 Doppler Shift

The Doppler shift is the observed change in wavelength of a wave when the source of the wave is moving relative to the observer. The wavelength is shorter when the source is moving towards the observer, and it is longer when the source is moving away from the observer. The observed wavelength can be written as [24, 25]

$$\lambda = \frac{\lambda_0}{1 - v/c}, \quad (3.1)$$

where λ_0 is the wavelength observed when the source is stationary (or the reference wavelength), v is the speed of the source towards the observer, and c is the velocity of the wave. Given a fixed observation point and an observed

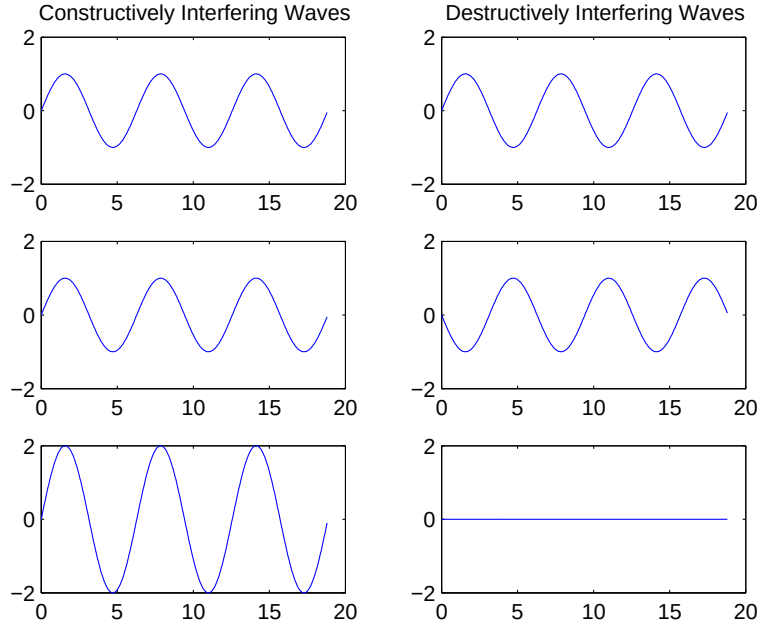


Figure 3.1: Constructive and destructive interference is shown. The figures on the left show two waves with magnitude one and a phase difference of zero summing to a wave with magnitude two. The figures on the right show two waves with a phase difference of π summing to zero.

wavelength for a known reference wavelength, Equation (3.1) can be used to find the velocity of the source:

$$v = c \left(1 - \frac{\lambda_0}{\lambda} \right). \quad (3.2)$$

This is the principle used for determining thermospheric winds with an FPI. When the Doppler shift is used for atmospheric sciences, we can accurately measure the wavelength of observed light. But difficulty arises from having no reference wavelength to compare this measurement to. Methods for overcoming this problem are discussed in later sections.

3.2.3 Doppler Broadening

Doppler broadening, also known as inhomogeneous broadening, is the broadening of spectral lines due to the Doppler effect [24]. While the average motion of an observed region of light-emitting molecules has a particular velocity, the motion of individual molecules has variance within that region.

Since light is emitted and collected from the entire area, the observed Doppler shift is the sum of all the individual Doppler shifts within the region. Therefore, an observation of the wavelength of light emitted in that region produces a broadened spectra.

In atmospheric science, the most significant form of Doppler broadening is thermal broadening, which is broadening due to the thermal motion of molecules. The higher the temperature of a region, the more motion there is in that region, and, therefore, the greater the broadening of emitted spectra. Therefore, the width of observed fringes is a function of the temperature of the observed region. The Doppler profile caused by thermal broadening is derived from the Maxwell-Boltzmann distribution function and can be written as [24, 25]

$$g(\lambda) = \left(\frac{mc^2}{2\pi kT\lambda_0^2} \right)^{1/2} \exp \left[\left(-\frac{mc^2}{2kT} \right) \left(\frac{\lambda - \lambda_0}{\lambda_0} \right)^2 \right], \quad (3.3)$$

where m is the mass of the emitting particle, k is Boltzmann's constant, T is temperature, and λ_0 is the peak wavelength. Observe that this is a Gaussian distribution with mean λ_0 and variance equal to

$$\sigma^2 = \frac{kT\lambda_0^2}{mc^2}. \quad (3.4)$$

3.3 Etalon Design

The fundamental idea of the Fabry-Perot interferometer is to form multipath ray interference to produce a pattern of interference fringes such that the position of the fringes is a function of the wavelength of the light source. These fringes form with high spectral resolution so that Doppler properties can be estimated. The interference fringes are generated with the use of two transparent plates positioned parallel to one another. These plates have plane surfaces with highly reflective coatings [20, 21]. When the distance between the two plates is fixed, they are referred to as an etalon.

3.3.1 Generation of Interference

This section explains how the etalon of an FPI produces interference fringes and the mathematical details of the fringes it produces. Referring to Figure 3.2, we call n the refractive index inside the etalon and assume the refractive index outside the etalon is 1. We call the coefficients of reflection and transmission of each plate r and t , respectively, and designate the etalon reflection and transmission as R and T , equal to r^2 and t^2 , respectively. We assume that loss is minimal at the interfaces so that $R + T = 1$.

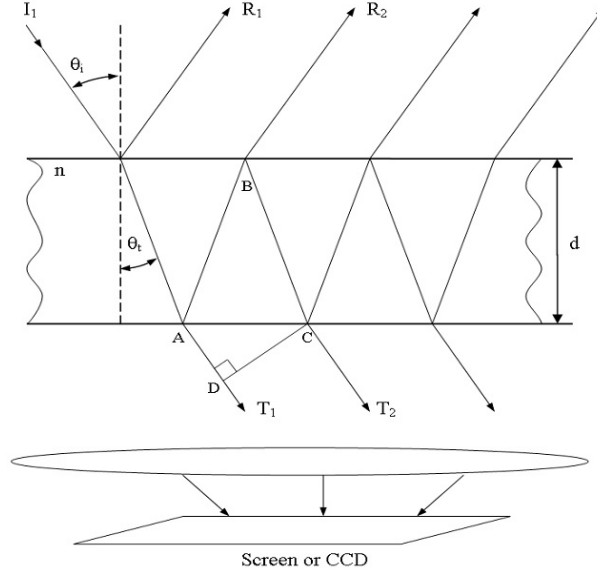


Figure 3.2: A two-dimensional cross section of the etalon of a typical FPI is shown with an incident light ray.

To understand how the FPI works, consider a plane wave incident on the top of the etalon. The path of one ray of the plane wave is represented by I_1 . The ray is incident on the first plate with angle θ_i . Part of the ray reflects with angle θ_i and the remainder of the ray passes into the etalon at angle θ_t . At the second interface, part of the ray reflects back into the etalon and part of it transmits through. The transmitted part (T_1) is focused onto a screen by a lens. The reflected part is incident on the upper plate of the etalon, where again part of it reflects and part of it transmits, and so on. The ray interference occurs as T_1 and T_2 exit the etalon. Since it is not a single ray but a plane wave that is incident on the etalon, these rays interfere with one another. In addition, the rays interfere differently at different locations on the observation screen. The path difference between the two can be written

as [24]

$$ABC - AD = \frac{2dn}{\cos \theta_t} - AC \sin \theta_i, \quad (3.5)$$

$$= \frac{2dn}{\cos \theta_i} - 2d \tan \theta_t \sin \theta_i, \quad (3.6)$$

$$= \frac{2dn - 2dn \sin^2 \theta_i}{\cos \theta_t}, \quad (3.7)$$

$$= 2dn \cos \theta_t. \quad (3.8)$$

Now we can determine the phase difference of the two rays:

$$\phi = \frac{\text{path difference} \cdot \text{angular frequency}}{\text{speed}}, \quad (3.9)$$

$$= \frac{2dn \cos \theta_t \cdot 2\pi f}{c/n}, \quad (3.10)$$

$$= \frac{4\pi dn \cos \theta_t}{\lambda}, \quad (3.11)$$

where λ is the frequency of the incoming light. In our case we can choose d and n in the etalon design process and λ is fixed, so Equation (3.11) describes the phase shift between T_1 and T_2 as a function of θ_t . The fraction of light transmitted to T_1 and T_2 is equal to t^2 and $tr^2e^{j\phi}t$, respectively. The summation of all light transmitted from the ray can be written as [22]

$$A = t^2 + t^2r^2e^{j\phi} + t^2r^4e^{j2\phi} + \dots = \frac{t^2}{1 - r^2e^{j\phi}}. \quad (3.12)$$

Using this, we write the transmitted intensity as [22]

$$T(\phi) = AA^* = \frac{t^4}{1 - r^2 \cos \phi + r^4} = \frac{(1 - R)^2}{1 - 2R \cos \phi + R^2}. \quad (3.13)$$

Since we choose R in the design process of the etalon, this equation is still a function of θ_t , the angle of the incident light. This equation can be further simplified to [22, 26]

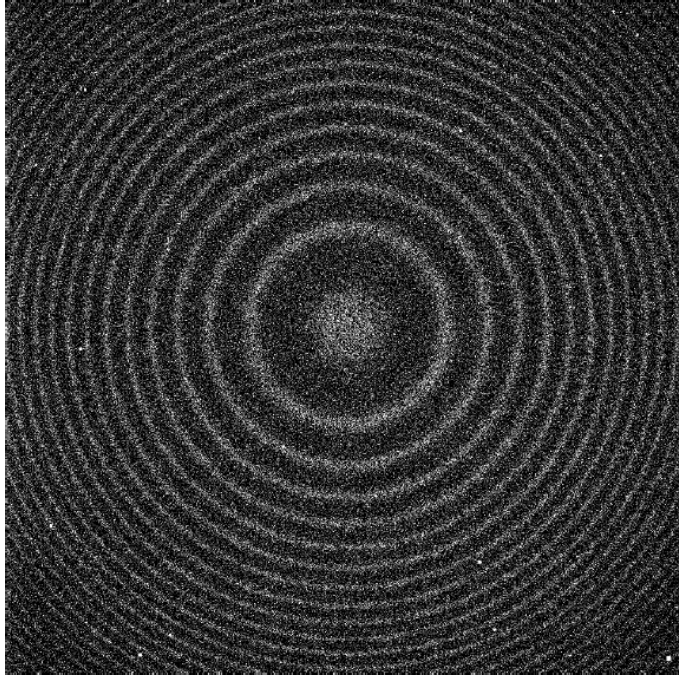


Figure 3.3: A typical image captured by a Fabry-Perot interferometer.

$$T(\phi) = \frac{(1-R)^2}{(1+R^2)} \left(\frac{1}{1 + [4R/(1-R)^2] \sin^2(\phi/2)} \right), \quad (3.14)$$

$$= \left(\frac{T}{(1-R^2)} \right)^2 (1 + F \sin^2(\phi/2))^{-1}, \quad (3.15)$$

$$= \left(\frac{T}{(1-R)} \right)^2 \mathcal{A}(\phi), \quad (3.16)$$

where $\mathcal{A}(\phi)$ is the Airy function and F is the finesse coefficient,

$$F = \frac{4R}{(1-R)^2}. \quad (3.17)$$

Finesse is defined as the FSR divided by the full-width, half-maximum of a fringe. The larger the finesse, the narrower the fringe is. Note that with our simplifying assumption of $R + T = 1$, the first term of Equation (3.16) is equal to one. The Airy function is discussed further in Section 3.3.3. As the angle of incidence of a light ray changes, the location where it is projected on the charge-coupled device (CCD) also changes. So the interference patterns form a series of concentric circles, as seen in Figure 3.3.

3.3.2 Etalon Distance and Index of Refraction

When designing an FPI etalon, a significant consideration for choosing the spacing and index of refraction of the etalon is free spectral range (FSR). For a center wavelength λ_0 , the FSR can be written as [22]

$$\Delta\lambda_0 = \frac{\lambda_0^2}{2dn}. \quad (3.18)$$

The FSR is the maximum change in wavelength that can be unambiguously determined by an etalon with distance d and index of refraction n . As an example, we simulated an etalon with $d = 0.015$ m, $n = 1$, and $R = 0.77$. The FSR of this etalon at 630.0 nm is 13.23 pm. Figure 3.4 shows transmitted intensity as a function of θ_t for wavelengths of 630.0 nm, 630.001 nm, and 630.013 nm. The shift of 0.008 nm is easily distinguishable from the original, but the shift of 0.013 nm causes a fringe pattern that is indistinguishable from the original pattern. This means that the etalon distance and index of refraction must be chosen so that the FSR is large enough for the fringe position to be a unique function of all incoming wavelengths of light for the desired application. Increasing the FSR increases the space between

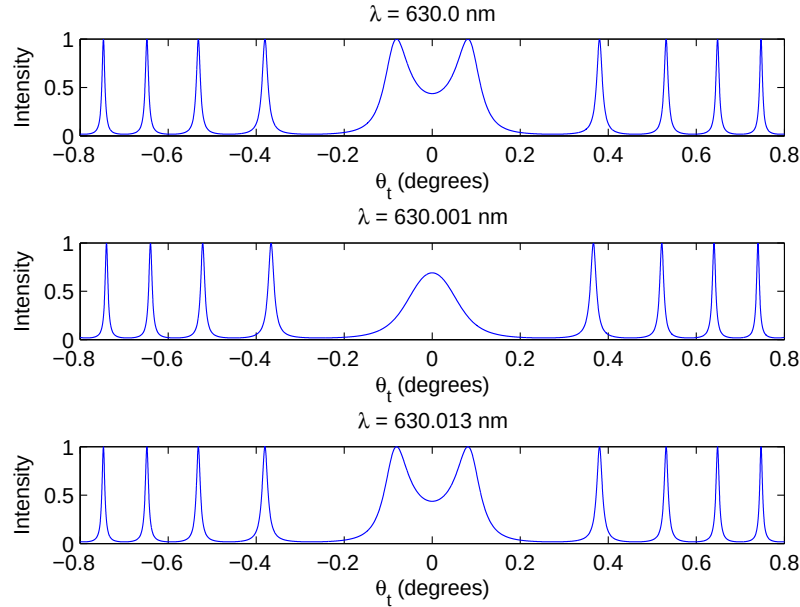


Figure 3.4: Simulated fringes generated by an etalon with $d = 1.5$ cm, $n = 1$, and $R = 0.77$.

each fringe, so that larger shifts produce unique fringes. In addition to this advantage, it also reduces the effect of fringe overlap, which is particularly important when high temperatures produce significant spreading. However, moving the fringes farther away from one another also means that fewer fringes will be produced from a fixed field of view (FOV), so it is important to balance the FSR between these priorities.

3.3.3 Etalon Reflectivity

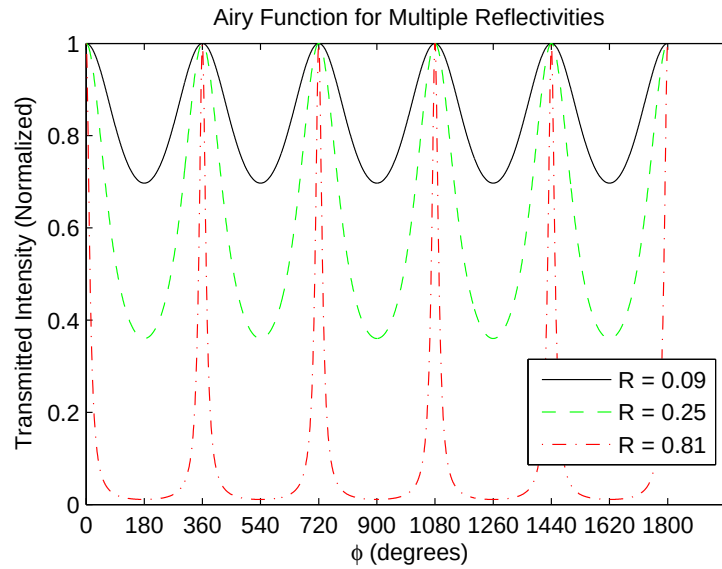


Figure 3.5: The Airy function, or the transmitted intensity of an FPI etalon, with a distance of 1.5 cm and index of refraction equal to 1.

Once the etalon distance and index of refraction have been chosen, we return to Equation (3.16), where the only remaining variable is R . Figure 3.5 shows a plot of the Airy function as a function of ϕ (defined in Equation (3.11)) for different values of reflectivity. It would seem that the higher reflectivity the better, since increasing reflectivity increases the finesse coefficient, or the “narrowness” of each fringe. The increase of spectra resolution is useful for accurate analysis of fringes; however, we must also take throughput into consideration. Throughput is the amount of light that transmits through the etalon, and the higher the reflectivity of the etalon, the lower the throughput is. We can write the throughput as $(1 - R)/(1 + R)$ [4]. The throughput of a system is roughly a linear inverse function of reflectivity, as

can be seen in Figure 3.6. Despite the significant loss of throughput, typical reflectivities range from 0.75 to 0.90 because narrow, easily resolvable fringes are a greater priority than throughput.

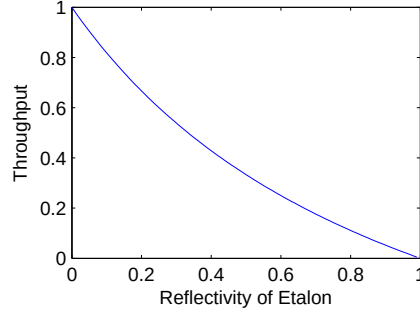


Figure 3.6: The percentage of light transmitted through an etalon as a function of its reflectivity is shown.

3.4 MiniME

The RENOIR project has a number of instruments located at two different field sites in Paraiba, Brazil. Each site is equipped with a Miniaturized Fabry-Perot interferometer (MiniME FPI) and one or more single-frequency GPS scintillation monitors. Additionally, one of the sites is equipped with an all-sky imaging system and a dual-frequency GPS receiver. This array of instruments allows the near-equatorial thermosphere/ionosphere (TI) to be studied in great detail. In particular, we study equatorial irregularities and the effects that neutral winds and gravity waves have on their development, the midnight temperature maximum, and nighttime ion-neutral coupling in the TI [4]. In this section, we list the technical details of MiniME and discuss the ramifications of its design.

The MiniME is a miniaturized, portable FPI system that uses a back-thinned CCD detector to estimate thermospheric winds and temperatures. Its name signifies that it is a miniaturized version of typical astronomical FPIs. The use of a high-performance CCD detector gives excellent sensitivity, spectral resolution, and temporal resolution. The MiniME etalon has a gap of 1.5 cm, index of refraction of 1, reflectivity of 77%, and a 42-mm aperture. It uses a 630.0-nm prefilter with a transmission of 55%. Referring back to

Section 3.3.2, we can use Equation (3.18) to calculate the FSR of the etalon and determine the maximum detectable wavelength shift:

$$\Delta\lambda_0 = \frac{\lambda_0^2}{2nd} = \frac{(630 \times 10^{-9})^2 \text{ m}^2}{2 \cdot 1 \cdot 1.5 \times 10^{-2} \text{ m}} = 13.23 \text{ pm.} \quad (3.19)$$

Using Equation (3.2), we find that this Doppler shift corresponds to a difference in wind speed of

$$\Delta v = c \cdot \left(1 - \frac{\lambda_0}{\lambda_0 + \Delta\lambda_0}\right), \quad (3.20)$$

$$= 3 \times 10^8 \text{ ms}^{-1} \cdot \left(1 - \frac{630.0 \times 10^{-9} \text{ m}}{630.1323 \times 10^{-9} \text{ m}}\right), \quad (3.21)$$

$$= 6,299 \text{ ms}^{-1}. \quad (3.22)$$

The largest expected thermospheric winds are less than 300 ms^{-1} , so this FSR is more than sufficient for unambiguous mapping of wavelength to fringe position. It would indeed be too large if the incoming light produced fringes with a high finesse. However, fringe spreading from the instrument function and thermal Doppler broadening cause the fringes to have a significant width. If the fringes are too close to one another, the overlap makes it difficult to accurately analyze the resulting images and the measurement error becomes large. Therefore, a large FSR is appropriate for this application. To demonstrate this, the fringe pattern shown in the top panel of Figure 3.4 was simulated with thermal Doppler broadening corresponding to a temperature of 1000 K, which is a typical nighttime temperature during times of middle to maximum solar activity. The broadened spectra is shown in Figure 3.7. The broadened fringes and increased minimum level show that the fringes are wider and are beginning to overlap.

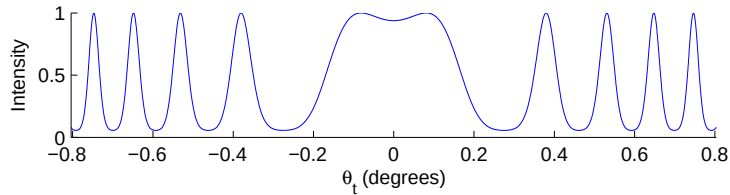


Figure 3.7: A one dimensional fringe pattern was generated using the same parameters as that of the fringes generated in Figure 3.4 with the addition of thermal broadening corresponding to a thermospheric temperature of 1000 K.

MiniME's reflectivity is lower than that of typical FPIs, which therefore yields an improvement in sensitivity in exchange for a decrease in spectral resolution and finesse. The effect of this adjustment on the overall accuracy of the instrument is negligible. Each image shows a total of 13 fringes on a CCD detector with a size of 1.33 cm. Increasing the number of fringe orders is equivalent to broadening the FOV of the instrument and increasing throughput. This partially compensates for the relatively small aperture of MiniME. The FOV angle for the outermost ring is 1.7° from center [4]. The Doppler shift and broadening can be calculated separately for each order, so increasing the number of orders provides an averaging affect that improves the accuracy of the instrument. Overall, the theoretical accuracy of this instrument is $\sim 3\text{--}5\text{ ms}^{-1}$ for typical brightnesses of the nighttime low-latitude 630.0-nm emission [4].

Using an FPI to measure Doppler shifts requires a mechanism for directing the line of sight (LOS) of the instrument. One mechanism that can be used is appropriately named the SkyScanner. The SkyScanner consists of a double-axis mirror system that can be rotated to point in any direction. It is controlled by software that is synchronized with the CCD software so that after calibration, it can point in a series of specified azimuthal and elevation angles with an accuracy of $\pm 0.1^\circ$. Figure 3.8 shows a SkyScanner and FPI installed in a cargo trailer. The FPI is affixed to a hole in the ceiling inside the trailer, and the SkyScanner is placed on the roof above the hole underneath a protective dome.



Figure 3.8: A SkyScanner (top) and MiniME FPI (bottom) are shown installed on the roof of a cargo trailer.

CHAPTER 4

OBSERVATION AND ANALYSIS

Now that we have discussed the details of the FPI itself, we move on to explain how it is applied to the field of atmospheric science to estimate thermospheric winds and temperatures. This chapter details a typical observation technique used for a single FPI. It continues with an explanation of the analysis method used for the RENOIR project and then finishes by discussing and demonstrating some of the inherent limitations of the FPI. For this discussion, we define the zonal, meridional, and vertical winds as u , v , and w , respectively. We additionally define the azimuthal angle as θ and the elevation angle as α .

4.1 Single FPI Observation Techniques

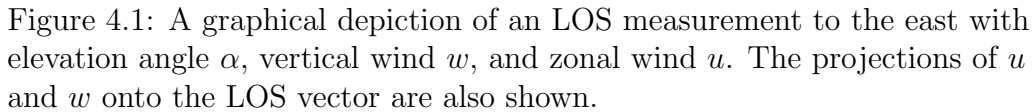
When a single FPI is used to observe thermospheric winds, it typically has an observation cycle consisting of a zenith measurement followed by an LOS observation in each of the cardinal directions with an elevation angle of 40–60 degrees.

The zenith measurement, or reference velocity, returns the vertical wind, w , and the instrument bias, β :

$$v_{ref} = w + \beta. \quad (4.1)$$

The instrument bias is primarily the result of the temperature variation of the etalon and the optics of the instrument. The etalon is temperature controlled, but slight fluctuations still have a significant effect. Failure to compensate for the instrument bias could produce a drift of hundreds of ms^{-1} in the data.

Any LOS measurement can be described as a function of the zonal, merid-


$$v = u \cos \alpha \sin \theta + v \cos \alpha \cos \theta + w \sin \alpha + \beta. \quad (4.2)$$
$$v_{east} = u \cos \alpha + w \sin \alpha + \beta. \quad (4.3)$$
$$v_{downst} = u \cos \alpha + w (\sin \alpha - 1). \quad (4.4)$$

29

measurement. Several methods have been proposed to calculate this wind velocity [27, 28, 29]. Most simplistically, it can be assumed that the vertical wind velocity is approximately zero. Under this assumption, the zenith measurement becomes a direct estimate of the instrument bias, and we can directly solve Equation (4.3) for the eastward wind:

$$u = \frac{v_{dop_e}}{\cos \alpha}. \quad (4.5)$$

Alternatively, Shiokawa [27] used the assumption that the wind field is a constant for the field of view of the instrument. Under this assumption, a measurement to the west is identical to the eastern measurement in Equation (4.4), except for a change in the sign of the Doppler shift, and can be written as

$$v_{dop_w} = -u \cos \alpha + w (\sin \alpha - 1). \quad (4.6)$$

Summing Equations (4.4) and (4.6) cancels out the zonal wind:

$$\begin{aligned} v_{dop_e} + v_{dop_w} &= u \cos \alpha + w (\sin \alpha - 1) - u \cos \alpha + w (\sin \alpha - 1), \\ &= 2w (\sin \alpha - 1). \end{aligned} \quad (4.7)$$

This equation can be rearranged to solve for w in terms of the observed measurements:

$$w = \frac{v_{dop_e} + v_{dop_w}}{2(\sin \alpha - 1)}. \quad (4.8)$$

The same method can be applied using observations in the northern and southern directions and any other observation pair whose angles differ by π radians. The results can be averaged to produce a more accurate estimate of the vertical wind velocity. However, this approach requires assuming that the vertical wind is both spatially and temporally constant for the entire range of measurement directions, an area that can cover 500×500 km or more. It also assumes that the zonal and meridional winds are constant over the entire measuring area for each measurement cycle. These are questionable assumptions, since the viewing area is a large region and a single measurement cycle takes 20 to 30 minutes.

Note that as α goes to $\pi/2$ in Equation (4.8), the denominator of Equation (4.8) goes to zero. This causes the uncertainty of the velocity estimation

to increase. To demonstrate this, vertical wind measurement uncertainty as a function of elevation angle was simulated. Two noisy LOS measurements in opposing directions were generated for a range of elevation angles. Each angle was simulated 100 times. The vertical wind was calculated, and the average error was determined as a function of elevation angle. The results are shown in Figure 4.2. It can be clearly seen that any angle greater than 65° will produce extremely high error.

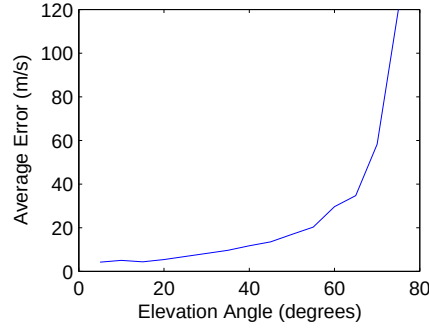


Figure 4.2: The average error of vertical wind calculation from two opposing LOS measurements as a function of elevation angle is shown.

4.2 Image Analysis

A typical observation cycle for MiniME involves taking a zenith measurement followed by an observation in each of the cardinal directions with an elevation angle of 45° . Each observation usually has a five-minute exposure time. In addition, dark images are taken periodically to approximate the effects of dark current on the CCD images, and a laser calibration image is taken every cycle. The laser image is used to determine the instrument function of the FPI. This is necessary because aberrations and imperfections in the optics of the instrument cause spreading. The laser beam serves as a bright source of a single, narrow frequency. Since the exact parameters of the source and the output are known, the effect of the optics of the instrument can be estimated. After a dark image is subtracted, each two-dimensional ring pattern is summed into a one-dimensional spectra by summing around the ring center coordinates. See [26] for a method for finding the center coordinates. This spectra is then fitted to a model using a method proposed by Killeen

and Hays [30]. This method employs the Levenberg-Marquardt nonlinear least squares (LM NLLS) algorithm to optimally fit the fringe pattern and model. This section gives a summary of the method proposed by Killeen and Hays and the modifications of it by Meriwether [31] that are now in use in the RENOIR project.

4.2.1 Analytic Description of FPI

The number of detected counts in the i th ring or channel on the CCD over an integration time of t seconds can be written as [30]

$$N_i = \frac{A_o \Omega_i t Q_i T \times 10^6}{4\pi} \int_0^\infty T_F(\lambda) \psi(\lambda, \theta_i) Y(\lambda) d\lambda + B_i, \quad (4.9)$$

where A_o is the working area of the etalon plates, Ω_i is the FOV of the i th ring detector, T is the optical transmission of the instrument, Q_i is the efficiency of the i th channel of the detector (including quantum efficiency and primary photoelectron collection efficiency), $T_F(\lambda)$ is the transmission function of the prefilter used to select the 630-nm emission line, $\psi(\lambda, \theta_i)$ is the instrument transfer function, and θ_i is the angle subtended by the i th ring detector at the imaging lens given by [30]:

$$\theta_i = \tan^{-1} \left[\frac{1}{f_o} \left(\frac{r_i^2 + r_{i+1}^2}{2} \right)^{1/2} \right], \quad (4.10)$$

where f_o is the effective focal length of the lens and r_i is the inner radius of the i th ring detector. $Y(\lambda)$ is the spectral distribution of light, which can be assumed to be two separate parts given by [30]

$$Y(\lambda) = \frac{\mathcal{R}_0 e^{-\left(\frac{\lambda - \lambda_0}{\Delta\lambda_T}\right)^2}}{\sqrt{\pi} \Delta\lambda_T} + \frac{\partial \mathcal{R}}{\partial \lambda} \Big|_0. \quad (4.11)$$

The first term in Equation (4.11) describes the normalized Gaussian corresponding to a thermal emission line with a surface brightness of $\mathcal{R}_0$ rayleighs, central wavelength λ_0 , and thermal width $\Delta\lambda_T$. The second term represents the continuum brightness underlying the emission line. Finally, B_i is the background count caused by dark current in the CCD.

Of the terms in Equation (4.9), the instrument transfer function remains

to be defined. If we express it in terms of its Fourier coefficients, it can be written as [30, 32]

$$T(\lambda, \theta_i) = a_{oi} + \sum_{n=1}^{\infty} \left[a_{ni} \cos \frac{2\pi n}{\Delta\lambda_o} (\lambda - \lambda_r + \phi_i) + b_{ni} \sin \frac{2\pi n}{\Delta\lambda_o} (\lambda - \lambda_r + \phi_i) \right], \quad (4.12)$$

where $\Delta\lambda_o$ is the FSR of the etalon, λ_r is the reference wavelength, and ϕ_i is the phase offset between the first channel and the i th channel. In this equation, the coefficients a_{ni} and b_{ni} contain all the information needed to completely describe the instrument function, including variations in sensitivity and defects in the optics. By convolving this with the spectral distribution of light defined in Equation (4.11), we can derive the instrument response to an emission line. By setting the reference wavelength, λ_r , equal to zero, we can define four unknowns as follows [31]:

$$U_1 = C_{oi} t \Delta\lambda_F \frac{\partial R}{\partial \lambda} a_o \left(\frac{1-R}{1+R} \right), \quad (4.13)$$

$$U_2 = C_{oi} a_o t T_{Fo} R_0, \quad (4.14)$$

$$U_3 = \frac{K \lambda_l}{\Delta\lambda_o}, \quad (4.15)$$

$$U_4 = G(T) = \frac{\pi}{c} \sqrt{\frac{2kT}{m}} \frac{\lambda_l}{\Delta\lambda_o}. \quad (4.16)$$

In these equations

$$C_{oi} = A_o \Omega_i Q_i T_{oi} \times 10^6 \cdot \frac{1}{4\pi}, \quad (4.17)$$

$$\Delta\lambda_F = \int_0^\infty T_F(\lambda) d\lambda, \quad (4.18)$$

The term $\partial R/\partial \lambda$ is the correction for the continuum brightness, K is the number of spectral elements in one order for the one-dimensional spectra, and m , k , and c are the atomic mass of the emitting oxygen atom, Boltzmann's constant, and the speed of light, respectively. Using the unknowns U_1 , U_2 , U_3 , and U_4 , we can write the instrument response as [30, 31]:

$$N_i = t(U_1 + U_2[a_o + \sum_{n=1}^{n_{max}} (a'_n \cos[2\pi n(\phi_i + U_3)] + b'_n \sin[2\pi n(\phi_i + U_3)]) \cdot e^{-n^2 U_4^2}] + B_i), \quad (4.19)$$

where a'_n and b'_n are defined as

$$a'_n = a_n \cos \beta + b_n \sin \beta, \quad (4.20)$$

$$b'_n = -a_n \sin \beta + b_n \cos \beta, \quad (4.21)$$

and β is the shifted phase, defined as

$$\beta = \frac{2\pi}{K} \left(1 + \frac{Peak_{emission}}{Peak_{calibration}} \right). \quad (4.22)$$

This model is now defined in such a way that the LM NLLS approximation can be used to analyze one-dimensional spectra generated from data images. Initial guesses are entered for the four unknowns, U_1 , U_2 , U_3 , and U_4 . Typically after five to nine iterations of the algorithm, estimates for the four parameters are given [31]. Then the Doppler shift is given by

$$v = -c\Delta U_3 \frac{\Delta\lambda_0}{\lambda_0 K}, \quad (4.23)$$

and the temperature is calculated by solving Equation (4.16). It should also be noted that U_2 is a measure of the intensity of the image.

4.2.2 Fringe Factor

Each fringe in the MiniME interferogram can be modeled as a Gaussian distribution. The difficulty with this approach is that Gaussian distributions extend to $\pm\infty$, so the fringes overlap into one another. Makela et. al. developed a forward model of the FPI and found that cross contamination between fringes causes the appearance of greater broadening and the LM NLLMS algorithm overestimates thermospheric temperatures [33]. This effect can be reduced by restricting the area of analysis to a fraction of each fringe. The fraction of the fringe analyzed is known as the fringe factor. Makela et. al.

found that decreasing the amount of the order used in the fitting reduces the effect of cross contamination. However, one cannot reduce the fringe factor indefinitely for two reasons. First, one must use some data, or the analysis routine simply won't work. When the fringe factor drops below 60%, the algorithm begins to have difficulty fitting the fringes. Second, the locations of the orders are fixed in this analysis routine. Therefore we must use enough of the order to ensure that the peak of the fringe will be included in the analyzed data. In practice it has been found that a fringe factor of 60% is generally a good balance between accuracy of the estimates and robustness of the LM NLLS algorithm.

4.3 Observation Limitations

Since the FPI is a passive optical instrument, it is not possible to use methods such as shuttering or time delay to determine the origination altitude of received signals. This means that each image of the FPI is an integration of all the light that is emitted along the LOS of the instrument. Typically, it is assumed that the wind speed estimated from the FPI data is the actual wind speed at the altitude of the emission peak. However, if there is a significant shear in the wind along the LOS, it is completely undetectable by a single, ground-based FPI. In order to study the validity of this assumption and demonstrate the inability to detect a complex wind profile using an FPI, we calculated the return of an FPI observing a wind profile with a shear wind that shifts from 10 ms^{-1} to 70 ms^{-1} over a 40-km height.

The top three panels in Figure 4.3 show the wind profile with three different positions of the center of the shear (CoS) superimposed on the AG emission profile. The bottom panel shows the correlating wind measurement that an FPI would return superimposed on the actual wind at the CoS as a function of the position of the CoS. In some cases, the measurement of the FPI yields a wind estimate that is far from the true wind speed at the emission peak. However, in most cases the measured wind speed strongly reflects the actual wind speed. Also note that the complexity of the vertical wind profile is not reflected in any way in the wind estimation of the FPI.

Figure 4.4 shows the percentage of the total AG emission contained in a vertical section as a function of the height of the section. Over one-half of the

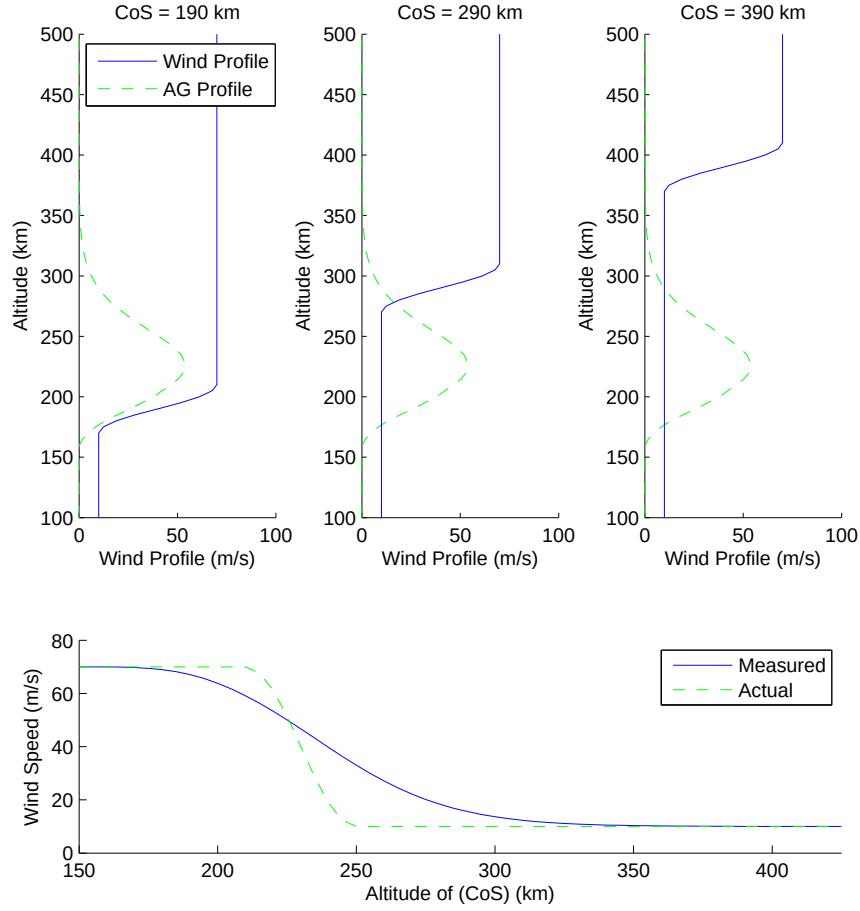


Figure 4.3: The top three panels show simulated vertical wind profiles superimposed on the AG emission profile (not to scale). The bottom panel shows a plot of the wind speed an FPI would estimate from the wind profile superimposed on the actual wind speed at the CoS.

total intensity is emitted at an altitude within 25 km of the peak emission altitude and 80% of the total intensity is within 45 km of the peak emission altitude. So, unless there is a shear in the wind profile within 45 km of the emission peak, the wind speed calculated from FPI data will be a good approximation of the neutral winds at the peak of the emission profile.

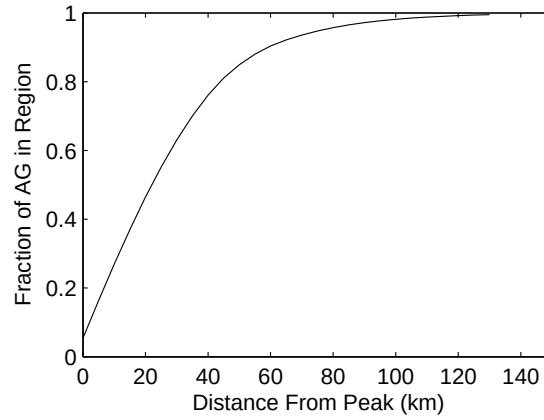


Figure 4.4: The fraction of AG emission contained in a region centered around the emission peak as a function of the distance from the peak in both directions. The x -axis is the distance from the peak upwards and downwards. The y -axis is the percentage of the total AG emission contained within the region. It can be seen that over 50% of the total AG is contained within ± 25 km of the emission peak, and 80% of the AG is contained within ± 45 km of the peak.

CHAPTER 5

BISTATIC FABRY-PEROT INTERFEROMETER OBSERVATION STRATEGY

In this chapter we describe a bistatic FPI system (BFPIS) that is optimized to minimize wind estimation error and maximize the usefulness of data. We begin by explaining the advantages of a BFPIS over the use of a single FPI. We then explain the process of developing an observation strategy that optimizes these advantages. We describe a method for extracting full wind vector information from data from the BFPIS. Finally, we show the parameters and results of simulations of this system that verify our claims of the advantages of the BFPIS over a single FPI.

5.1 Advantages of BFPIS

This section explains the details of the advantages of the use of a BFPIS over a single FPI and demonstrates the significance of these advantages.

There are two primary advantages to the BFPIS. First, it improves the accuracy of wind estimation by reducing the assumptions needed to obtain a zero wind reference wavelength. Second, it can be used to obtain full vector wind information at multiple locations.

5.1.1 Reduction of Assumptions

In Section 4.1 we explained that several assumptions must be made in order to estimate wind speeds from data from a single FPI. The most significant challenge is obtaining a zero-wind reference velocity. The most common method to do this is to assume that vertical wind is zero or assume that the wind field is uniform for the entire viewing area. Assuming that the vertical wind is zero is a straightforward approach in which the zenith measurements are used to find the reference wavelength. However, this is not a safe assump-

tion for some geographical regions. For these regions, the second approach must be used in which it is assumed that the observed wind field is uniform. The problem with this approach is that the viewing area of an FPI is quite large, making the assumption suspect. For an FPI observing the cardinal directions at a 45° elevation angle, the observation area is defined by the four observation points that are away from the instrument in each direction (see Figure 5.1). The distance from the instrument to these sites is equal to

$$d = \frac{\text{observation height}}{\tan 45^\circ} = \frac{250 \text{ km}}{1} = 250 \text{ km}, \quad (5.1)$$

where observation height is set to 250 km since the majority of the AG emission is concentrated at that height. Therefore, the observation region is approximately a $500 \text{ km} \times 500 \text{ km}$ region. If the elevation angle is set to 60° , the observation region shrinks to 250 km by 250 km, but increasing the observation angle reduces the accuracy of horizontal wind estimation, as shown in Section 4.1.

When a BFPIS is used, it is necessary to assume that only the vertical wind is uniform for the viewing area of the two FPIs. This area is approximately the same size as that of a single FPI if the observation directions of the two instruments are limited to zenith measurements and measurements towards the other instrument, as is done with RENOIR.

5.1.2 Full Vector Wind Information

The BFPIS can be used to obtain full vector wind information at multiple locations. By this, we mean that the zonal and meridional thermospheric wind speeds can be estimated for distinct locations rather than having to assume a uniform horizontal wind field. To demonstrate the advantage of having two distinct observation locations, an oscillating horizontal wind with a zero vertical wind moving through an observation area was simulated for a single FPI and a BFPIS. A wavefront with an amplitude of 10 ms^{-1} , a wavelength of 300 km, and propagation velocity of 27.8 ms^{-1} (100 km/hr) was simulated. The propagation direction had an azimuthal angle of $\frac{7\pi}{6}$ rad and the direction of the wind had an angle of $\frac{5\pi}{3}$ rad. This is shown graphically in Figure 5.2. It was assumed that the wave properties remained consistent for the entire observation time.

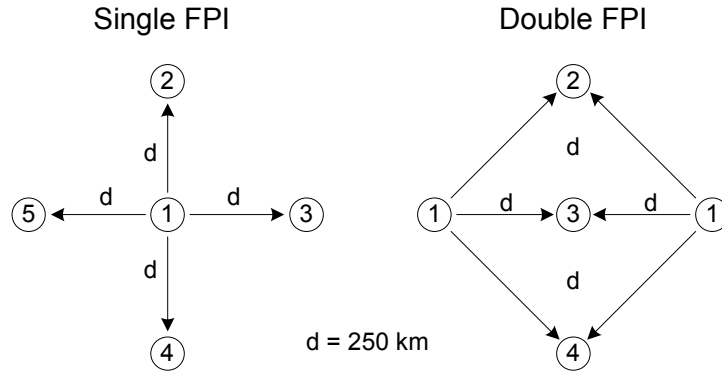


Figure 5.1: The simulated observation sequences are shown for the single FPI and the double FPI setups. Instruments are located at the “1” shown on the figure, which represents a zenith measurement.

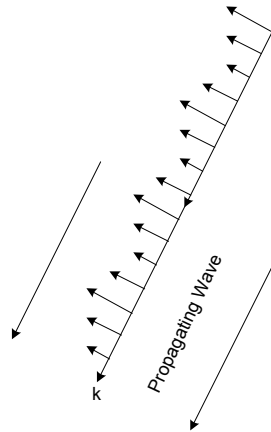


Figure 5.2: A graphical representation of the simulated wave field is shown.

The single FPI analysis was done assuming that the vertical wind was zero and no noise was added to the measurements. Measurements in the cardinal directions were simulated using a 45° elevation angle and 5-minute exposure times. The observation sequence is depicted in the left panel of Figure 5.1, where “1” designates the location of the instrument and each consecutive number designates the order of observation. In a similar manner, the BFPIS observation sequence is shown in the right panel of Figure 5.1. The two FPIs simultaneously take zenith measurements and then observe northern, inline, and southern regions. Because of the 5-minute exposure time, the single FPI obtains data at each observation point every 25 minutes, while the BFPIS obtains data at each observation point every 20 minutes.

It was attempted to determine the wave structure from the data from each observation method. The meridional wind speed at the northern location of the single FPI is directly measured. To determine the zonal wind, the eastern observation data were interpolated to determine the observed wind speeds at the same time that the northern observation was taken. A comparison of the determined zonal and meridional wind at the northern site to the true wind velocities is shown in Figure 5.3. The meridional wind is accurately determined, but the zonal wind contains a phase shift because of the phase difference between the northern observation site and the eastern observation. Theoretically, the angle of propagation of the wave can be determined from this phase shift. However, in practice this process is very difficult to perform because of imprecise measurements and non-uniform wave structures.

In the case of the bistatic FPI system, the meridional and zonal winds at the northern and southern observation points can be directly determined from the two simultaneous measurements of each point. The results of these calculations are shown with the true wind velocities in Figure 5.4. It can be clearly seen that the BFPIS observations provide sufficient data to accurately reconstruct the meridional and zonal wind velocities. This method is robust even when used with noisy measurements and non-uniform wave structures. Additionally, since the wind speeds can be determined at two separate locations, the wavelength and propagation velocity can be estimated using data from the BFPIS, although these parameters cannot necessarily be unambiguously determined. Figure 5.5 shows the meridional and zonal winds from the northern volume and the southern volume plotted together. From the observation of either volume, it can be seen that the wave has a period

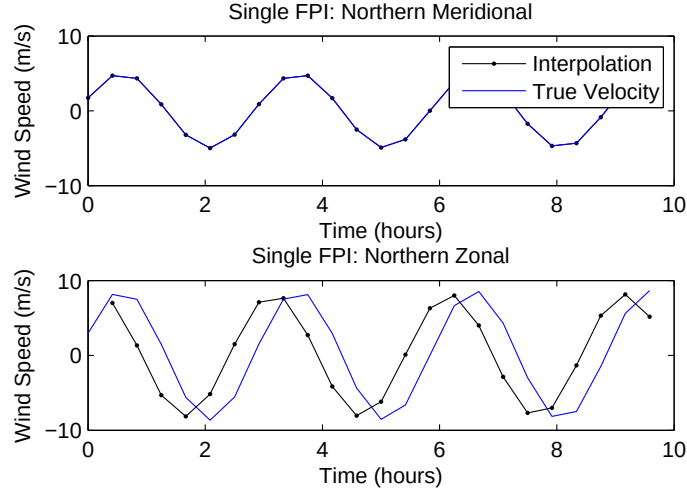


Figure 5.3: The result of a single FPI observing an oscillating wind field is shown. The differences between the north-south and east-west measurements result from the structure of the wave.

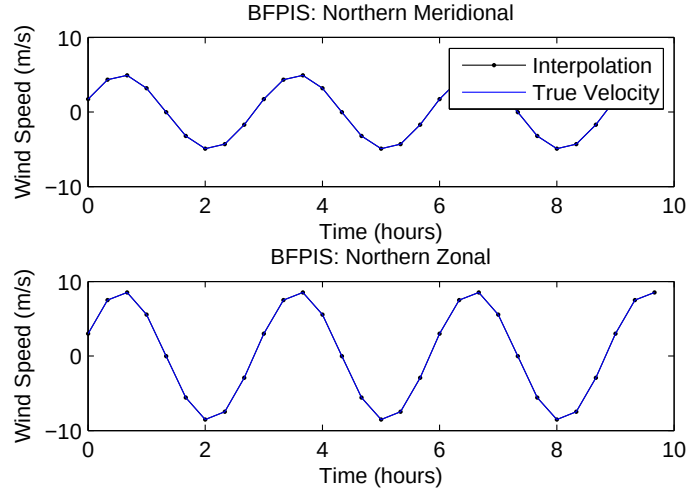


Figure 5.4: The result of a BFPIS observing an oscillating wind field is shown. Since the zonal and meridional winds are known at two separate regions, the form of the original wave can be determined.

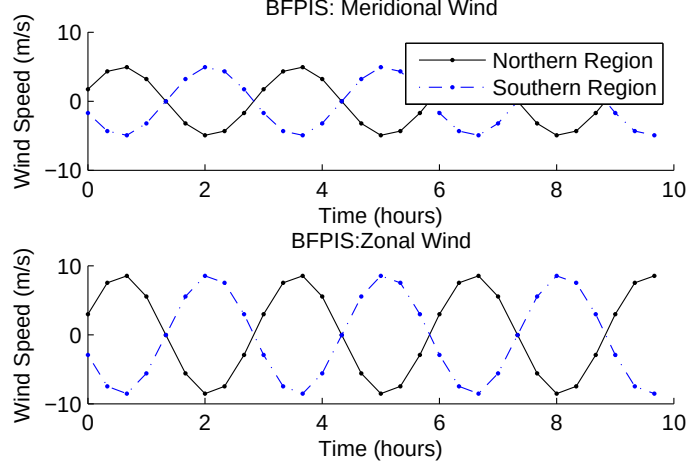


Figure 5.5: Wind vector calculations based on the simulated BFPIS measurements are shown. Meridional and zonal winds at the northern volume and the southern volume are shown together.

of three hours. But the phase difference between the waves observed at each volume cannot be known with certainty without additional information. The travel time from volume to volume is roughly equal to $1.5 + 3.0n$ hr where $n = 0, 1, 2$, etc. If an endpoint of the wave structure or another singularity is observed at both volumes, the propagation velocity and wavelength of the structure can be definitively determined. Without real data from a BFPIS, it is difficult to comment further on this topic.

5.2 Developing a BFPIS Observation Technique

We had limited flexibility in choosing the locations of the two instrument sites used in RENOIR. However, since the SkyScanner used to direct the FOV of the FPIs can be pointed in any direction, we have almost limitless options in the choice of an observation cycle. It was necessary to carefully choose the observation pattern that would allow us to estimate the cardinal directions of the thermospheric wind with minimal error. Some choices are clear: the zenith measurement is always necessary to determine the instrument bias. Also, an inline measurement, where the instruments angle directly towards one another such that they observe a common volume at the estimated maximum AG emission is essential to maximize accuracy in determining the vertical wind. This observation is the only way to directly

measure the vertical wind of a single region using a BFPIS. Finally, it was desired that we be able to observe the wind components in a northern location and a southern location to search for waves propagating in the meridional direction. In order to simplify the comparison between the two instruments, it was decided to observe along the longitudinal line directly between the instruments. In addition, it was desired that the LOS of the instruments meet at an altitude of 250 km. Even after these restrictions, there are a theoretically limitless number of points that could be observed, so we chose the two points where the errors caused by noise on the observations would be minimized. This is shown graphically in Figure 5.10 (see page 52). This chapter shows that the optimal observation directions are those in which the azimuthal angle of the observations differ by $\pi/2$ radians. It then describes how those locations were chosen and how the proper observation angles to observe them were determined.

5.2.1 Choosing Observation Directions to Minimize Error

The following section shows that error from determining the full wind vector from two noisy observations of a common volume is minimized when the azimuthal angles of the observation vectors of a common volume differ by $\pi/2$. We will begin by showing that this is true for the meridional wind vector and then generalize the result.

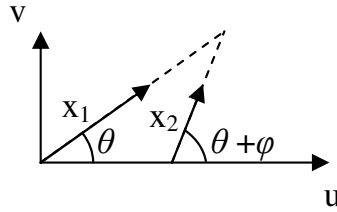


Figure 5.6: Two observations, x_1 and x_2 , are made of a horizontal wind field from two arbitrary angles, θ and $\theta + \varphi$.

Given a horizontal wind field with zonal wind u and meridional wind v , we make two observations, x_1 and x_2 , from two arbitrary angles, θ and $\theta + \varphi$ (see Figure 5.6). We can see that

$$x_1 = u \cos \theta + v \sin \theta, \quad (5.2)$$

$$x_2 = u \cos(\theta + \varphi) + v \sin(\theta + \varphi). \quad (5.3)$$

If we solve Equation (5.2) for u and substitute the resulting expression into Equation (5.3), we obtain

$$x_2 = (x_1 - v \sin \theta) \cdot \frac{\cos(\theta + \varphi)}{\cos \theta} + v \sin(\theta + \varphi). \quad (5.4)$$

Rearranging Equation (5.4) to solve for v gives us

$$v = \frac{x_2 - x_1 \cdot \frac{\cos(\theta + \varphi)}{\cos \theta}}{\sin(\theta + \varphi) - \tan \theta \cos(\theta + \varphi)}. \quad (5.5)$$

In order to determine the minimal error in the presence of noise, we must add noise to x_1 and x_2 :

$$v_{noise} = \frac{x_2 + n_2 - (x_1 + n_1) \cdot \frac{\cos(\theta + \varphi)}{\cos \theta}}{\sin(\theta + \varphi) - \tan \theta \cos(\theta + \varphi)}. \quad (5.6)$$

Because of the complexity of Equations (5.5) and (5.6), it is simplest to use Matlab to calculate v and v_{noise} for θ and φ varying from 0 to 2π . It can be simulated with x_1 and x_2 as arbitrary constants and n_1 and n_2 as random, mean zero, Gaussian noise. We generated v_{noise} for the full ranges of θ and φ fifty times and took the difference between v_{noise} and v . Table 5.1 shows the values we used for our simulation.

Table 5.1: Values Used for Simulation of Horizontal Wind Calculation.

Variable	Value(s)
x_1	50 ms ⁻¹
x_2	50 ms ⁻¹
Var(n_1)	2 ms ⁻¹
Var(n_2)	2 ms ⁻¹
θ	0 – 360 degrees
φ	0 – 360 degrees

Figure 5.7 shows the average of the absolute value of $v - v_{noise}$ as a function of φ for fifty iterations. We also averaged the varying θ without loss of generalization, since a result for an arbitrary θ is desired. We can observe that the minimum error occurs at $\varphi = \pi/2$ and $3\pi/2$. Using smaller or larger values for x_1 and x_2 had little effect on the results.

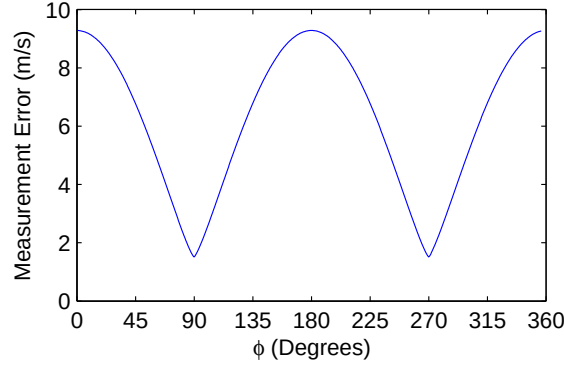


Figure 5.7: The average error from calculating the meridional wind from two noisy measurements with arbitrary angles in the horizontal plane are shown as a function of the difference between the angles. The error was simulated 100 times for all possibilities of θ and φ with random Gaussian noise. Errors for each run and for varying θ were averaged together to show average error as a function of φ .

Similar steps can be applied to find the same result for minimizing error measuring the zonal wind. So, we can conclude that when observing the wind speed in a common volume along two different horizontal vectors, error is minimized when the observation vectors differ by an angle of $\pi/2$. Alternatively, this can be shown by finding the derivative of Equation (5.5) with respect to φ , setting it equal to zero, and solving for φ ; but the tediousness of the math involved in this exercise caused us to choose the above method.

5.2.2 Determining Observation Angles

Once the appropriate observation pattern had been decided on, it was necessary to determine the proper azimuthal and elevation angles of each instrument. This involves converting GPS coordinates into local coordinate systems. Before this process is explained, it is necessary to briefly discuss several geographic coordinate systems.

Coordinate Systems

In this section we will briefly explain Earth-Centered, Earth-Fixed (ECEF) coordinates, the World Geodetic System 1984 (WGS-84), and local coordinate systems.

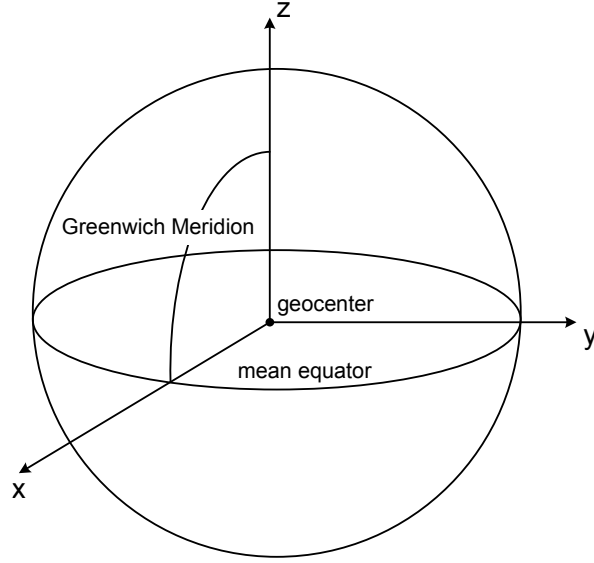


Figure 5.8: Earth-Centered, Earth-Fixed coordinates are shown.

The ECEF coordinate system is a right-handed Cartesian coordinate system with the center of the earth as its origin. The x -axis points through the mean Greenwich meridian, the z -axis coincides with the rotational axis of the earth, and the y -axis is set to complete a right-handed coordinate system. This is shown in Figure 5.8 [34].

The World Geodetic System 1984 (WGS-84) is more commonly known as GPS. According to this system, the earth is modeled as an oblate spheroid. The origin is the center of mass of the earth. Position relative to the origin is measured in latitude, longitude, and height. In this section, we will represent these parameters with φ , λ , and h . In order to define the earth as an oblate spheroid, we must define not a single radius, but a semi-major radius (which we will denote a) and a semi-minor radius (denoted as b) of the rotating ellipse. Using these parameters, we can define the relationship between ECEF coordinates and WGS-84 coordinates [34]:

$$x = (N + h) \cdot \cos(\varphi) \cdot \cos(\lambda), \quad (5.7)$$

$$y = (N + h) \cdot \sin(\lambda) \cdot \cos(\varphi), \quad (5.8)$$

$$z = (N \cdot (1 - e^2) + h) \cdot \sin(\varphi), \quad (5.9)$$

where N is the curvature of the radius of the earth, defined as [34]

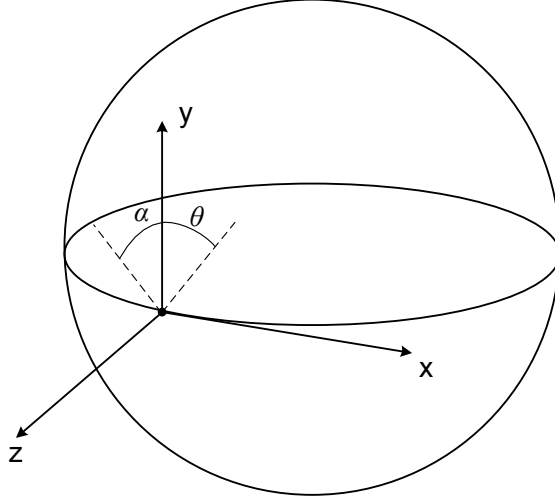


Figure 5.9: A local reference frame is shown at an arbitrary point on the surface of the earth.

$$N = \frac{a}{(1 - e^2 \sin \varphi)^{1/2}} \quad (5.10)$$

and e is known as the first eccentricities, defined by [34]

$$e = \frac{\sqrt{a^2 - b^2}}{a}. \quad (5.11)$$

A local coordinates system is defined by placing the origin at an arbitrary point. For the purposes of this thesis the origin will always be a point on the surface of the earth. A local coordinate system can be defined several ways. We define a local coordinate system as follows: from the origin the x -axis is to the east, the y -axis is to the north, and the z -axis is directly away from the center of the earth, as shown in Figure 5.9. This local coordinate system is also known as Vertical, East, North (VEN). Note that each point on the earth has a different local coordinate system with different axis. It is also helpful to define the local coordinate system in spherical coordinates using the azimuthal angle, the elevation angle, and a magnitude. The azimuthal angle is the angle between a vector in the xy -plane and the y -axis, measured rotating clockwise from the y -axis. The elevation angle is the angle between a vector and the xy -plane. This is also shown in Figure 5.9, with azimuthal and elevation angles represented by θ and α , respectively. Therefore, any

three-dimensional vector can be defined in the local reference frame in terms of its xyz coordinates or in terms of azimuthal and elevation angles and a magnitude. The matrix that converts a vector with points defined in ECEF to VEN coordinates can be written as [34]

$$VEN = \begin{pmatrix} \cos \varphi \cdot \cos \lambda & \cos \varphi \cdot \sin \lambda & \sin \varphi \\ -\sin \lambda & \cos \lambda & 0 \\ -\sin \varphi \cdot \cos \lambda & -\sin \varphi \sin \lambda & \cos \varphi \end{pmatrix} \begin{pmatrix} X_f - X_i \\ Y_f - Y_i \\ Z_f - Z_i \end{pmatrix}, \quad (5.12)$$

where $VEN = (z; x; y)$, the subscript i denotes the initial location, and the subscript f denotes the final location. The origin is set to the initial location, and the vector pointing to the final location is determined. VEN coordinates can further be put in terms of azimuthal and elevation angles using the following relationships:

$$\alpha = \frac{\pi}{2} - \cos^{-1}(z), \quad (5.13)$$

$$\theta = \tan^{-1} \left(\frac{x}{y} \right). \quad (5.14)$$

Care must be taken with the use of Equation (5.14). Depending on the implementation of the tangent inverse function, there may be a need to shift the result by π to put the vector in the correct tangent.

Converting Coordinate Systems

Now that we have defined the needed coordinate systems, we can explain the process of determining the proper azimuthal and elevation angles for each look direction of each instrument. The WGS-84 coordinates of the center region were determined by averaging the coordinates of the two instrument sites. Finding the coordinates of the northern and southern observation regions was a more intensive process. The longitude was fixed to be the same as the longitude of the center region. An array of points every 0.005° north and south of this position was then considered. Each point was converted into ECEF coordinates and then the elevation and azimuthal angles that point from the instrument locations to that observation point were calculated.

The northern and southern observation sites were chosen using the following functions:

$$site_n = \min(\theta_A - \theta_B + 270), \quad (5.15)$$

$$site_s = \min(\theta_A - \theta_B + 90), \quad (5.16)$$

where θ_A and θ_B denote the azimuthal angles from Sites A and B, respectively. The results of this calculation for the RENOIR installation in Brazil are listed in Tables 5.2 and 5.3 and shown graphically in Figure 5.10. Values are rounded to the nearest 0.1° since that is the greatest precision the SkyScanner is capable of. An error of 0.05° is acceptable since the FOV of the FPI is 3.4° .

Table 5.2: Instrument and Observation Locations of the RENOIR Brazil BFPIS.

Location	GPS Coordinates
Site A	-6.87N, -38.56E
Site B	-7.38N, -36.53E
Center Observation	-7.13N, -37.54E
Northern Observation	-6.08N, -37.54E
Southern Observation	-8.17N, -37.54E

Table 5.3: Azimuthal and Elevation Angles for Each FPI and Each Look Direction.

Look Direction	Site A		Site B	
	θ	α	θ	α
Center	104.2	64.2	284.1	64.2
Northern	52.1	59.2	322.1	52.5
Southern	142.1	52.6	232.1	59.3

5.3 Extracting Full Wind Vectors

This section shows that full vector wind information for multiple locations can be extracted from data from a BFPIS using fewer assumptions than

would be used for a single FPI. Therefore, the variability between near-equatorial and very-near-equatorial winds can be observed, and the propagation of changes in wind speed can be tracked.

5.3.1 Vertical Wind Calculation

The following describes a method for calculating the vertical wind between two FPIs. We must assume that the vertical wind is a constant for the shared FOV of the two instruments, which consists of three points between the instruments and the space directly above the instruments themselves. According to the observation technique described above, each instrument takes a measurement in the zenith direction to measure the vertical wind and the bias. Next, the instruments take a series of three measurements of common volumes along the longitudinal line directly between them, as shown in Figure 5.10. The elevation angles are adjusted for each common volume measurement, so that the observation paths of the instruments meet at the maximum AG emission altitude, 250 km above the earth. For the RENOIR project the instruments are positioned as shown in Figure 5.10 because of site availability and our desire to compare near-equatorial winds to very-near-equatorial winds. The derivation that follows applies specifically to this configuration; however, it can easily be adapted for any two FPIs in a bistatic formation.

For the inline measurements, the azimuthal angles have the following relationship:

$$\theta_2 = \theta_1 + \pi, \quad (5.17)$$

where θ_1 and θ_2 denote the azimuthal angle of the instruments at Sites A and B, respectively. The inline measurement of each instrument can be written as:

$$v_{in1} = [u \sin \theta + v \cos \theta] \cos \alpha + w \sin \alpha + \beta, \quad (5.18)$$

$$\begin{aligned} v_{in2} &= [u \sin(\theta + \pi) + v \cos(\theta + \pi)] \cos \alpha + w \sin \alpha + \beta, \\ &= -[u \sin \theta + v \cos(\theta + \pi)] \cos \alpha + w \sin \alpha + \beta, \end{aligned} \quad (5.19)$$

where α is the angle of elevation, θ is the azimuthal angle, u is zonal wind, and v is meridional wind. The observations from Site A and Site B are

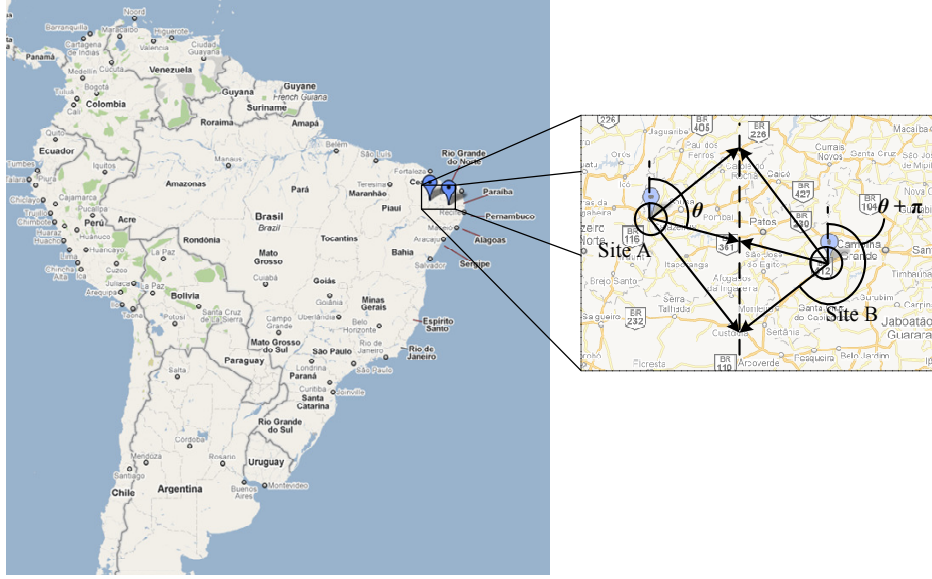


Figure 5.10: Two FPIs are located approximately 250 km away from each other. Each instrument takes a zenith measurement and then they simultaneously observe three common volumes along the longitudinal line that is equidistant from them.

represented by v_{los1} and v_{los2} , respectively. We define g to be the horizontal wind vector pointing from Site A to Site B:

$$g = [u \sin \theta + v \cos \theta]. \quad (5.20)$$

Substituting g into Equations (5.18) and (5.19) and subtracting the zenith measurement described in Equation (4.1) from each equation eliminates the instrument bias. This produces an estimate of the LOS Doppler velocity:

$$v_{dop1} = g \cos \alpha + w \sin \alpha - w, \quad (5.21)$$

$$v_{dop2} = -g \cos \alpha + w \sin \alpha - w. \quad (5.22)$$

Summing Equations (5.21) and (5.22) eliminates the horizontal wind component and allow us to solve for the vertical wind:

$$w = \frac{v_{dop1} + v_{dop2}}{2 (\sin \alpha - 1)}. \quad (5.23)$$

The horizontal wind vector can be determined from the difference of Equations (5.21) and (5.22):

tions (5.21) and (5.22), but little is gained by this information because a single wind vector cannot be decomposed to determine the zonal and meridional winds. Note that Equation (5.23) is nearly identical to Equation (4.8). The key difference is that in Equation (5.23) v_{dop1} and v_{dop2} are measurements of the same area of the thermosphere, instead of two regions that are separated by hundreds of kilometers.

5.3.2 Horizontal Wind Calculation

Having solved for the vertical wind, we can find the horizontal wind component of each measurement at the northern and southern observation points by assuming that the vertical wind is uniform over the observation region. We begin by deriving equations that apply to both observation regions and will separate the two when distinction becomes necessary.

The LOS measurements of either location from each instrument can be written as

$$v_1 = [u \sin \theta_1 + v \cos \theta_1] \cos \alpha_1 + w \sin \alpha_1 + \beta, \quad (5.24)$$

$$v_2 = [u \sin \theta_2 + v \cos \theta_2] \cos \alpha_2 + w \sin \alpha_2 + \beta, \quad (5.25)$$

where the subscripts 1 and 2 denote the observation parameters of the instruments from Site A and B, respectively. Note that the azimuthal and elevation angles for each measurement are unique for each measurement. We now subtract the zenith measurement to remove the instrument bias and use the vertical wind determined above to extract the horizontal wind measurement, which can be written as

$$v_{h1} = \frac{v_1 - (w + \beta) - w(\sin \alpha_1 - 1)}{\cos \alpha_1}, \quad (5.26)$$

$$v_{h2} = \frac{v_2 - (w + \beta) - w(\sin \alpha_2 - 1)}{\cos \alpha_2}. \quad (5.27)$$

Now that we have defined the horizontal wind in terms of the LOS measurements, we need to define the relationship between the horizontal wind vectors and the meridional and zonal wind. We obtain this by substituting Equation (5.24) into Equation (5.26) and Equation (5.25) into Equation (5.27):

$$v_{h1} = u \sin \theta_1 + v \cos \theta_1, \quad (5.28)$$

$$v_{h2} = u \sin \theta_2 + v \cos \theta_2, \quad (5.29)$$

We can define these relationships using a transform matrix, M :

$$M = \begin{pmatrix} \sin \theta_1 & \cos \theta_1 \\ \sin \theta_2 & \cos \theta_2 \end{pmatrix}, \quad (5.30)$$

$$\begin{pmatrix} v_{h1} \\ v_{h2} \end{pmatrix} = M \begin{pmatrix} u \\ v \end{pmatrix}, \quad (5.31)$$

$$\begin{pmatrix} u \\ v \end{pmatrix} = M^{-1} \begin{pmatrix} v_{h1} \\ v_{h2} \end{pmatrix}. \quad (5.32)$$

The inverse of M exists only if its determinant is not equal to zero. The determinant of M will only equal zero if θ_1 or θ_2 are equal to zero or a multiple of $\pi/2$. In these cases, the instruments would be directly observing the zonal or meridional winds, and the relationship between the horizontal wind components and the zonal and meridional wind would be much more direct. Since this is not our situation for the RENOIR project, the inverse of M exists.

We now move on to discuss the particular cases of the northern observation region and the southern observation region separately. For the northern region, we have chosen the azimuthal angles of the two instruments such that they are related by

$$\theta_2 = \theta_1 + \frac{3\pi}{2}. \quad (5.33)$$

Using this relationship, we can simplify the transformation matrix to

$$M = \begin{pmatrix} \sin \theta_1 & \cos \theta_1 \\ -\cos \theta_1 & \sin \theta_1 \end{pmatrix}, \quad (5.34)$$

and write the inverse of this matrix as

$$M^{-1} = \begin{pmatrix} \sin \theta_1 & -\cos \theta_1 \\ \cos \theta_1 & \sin \theta_1 \end{pmatrix}. \quad (5.35)$$

Therefore, we can explicitly write the zonal and meridional winds in terms

of the horizontal wind vectors:

$$u = v_{h1} \sin \theta_1 - v_{h2} \cos \theta_1, \quad (5.36)$$

$$v = v_{h1} \cos \theta_1 - v_{h2} \sin \theta_1. \quad (5.37)$$

For the southern observation point, the azimuthal angles have the following relationship:

$$\theta_2 = \theta_1 + \frac{\pi}{2}. \quad (5.38)$$

The transformation matrix then simplifies to

$$M = \begin{pmatrix} \sin \theta_1 & \cos \theta_1 \\ \cos \theta_1 & -\sin \theta_1 \end{pmatrix}, \quad (5.39)$$

and we write the inverse of this matrix as

$$M^{-1} = \begin{pmatrix} \sin \theta_1 & \cos \theta_1 \\ \cos \theta_1 & -\sin \theta_1 \end{pmatrix}. \quad (5.40)$$

As before, this defines an explicit relationship between the zonal and meridional winds and the horizontal wind vectors:

$$u = v_{h1} \sin \theta_1 + v_{h2} \cos \theta_1, \quad (5.41)$$

$$v = v_{h1} \cos \theta_1 - v_{h2} \sin \theta_1. \quad (5.42)$$

5.3.3 Alternative Vertical Wind Calculation

When simulated, the method described for calculating the vertical wind that was described in Section 5.3.1 resulted in large inaccuracies for our equipment setup. The RENOIR setup uses an elevation angle of 64° for the center observation. Therefore, as Figure 4.2 (see page 31) suggests, the vertical wind measurement is extremely imprecise. To understand this, we look more closely at Equation (5.23). This equation can be thought of as performing two distinct operations: extracting the vertical wind component of the measurements and projecting this component onto the z -axis. In the denominator, the horizontal components of v_{dop1} and v_{dop2} cancel out, leaving the vertical wind component and whatever error is in the measurement. The numerator projects this component and error onto the z -axis. With an elevation angle

of 64° , the numerator evaluates to 0.2. Therefore, we end up multiplying the vertical wind component and the composite error of the measurements by a factor of five. This causes errors of great magnitude in the vertical wind calculation.

One method for reducing this error is to return to the assumption of a uniform wind field over the observing area. The measurement averaging method is valid for any pair of opposing measurements, so with the uniformity assumption we can average the northern observation from Site A with the southern observation from Site B because the observation angles differ by π radians. If the observed wind field is uniform, the average of these measurements will accurately produce the vertical wind. Naturally, we can also use the southern observation from Site A paired with the northern observation from Site B. This method has potential for improving the vertical wind speed estimates; however, averaging these results requires assuming that the wind field of the entire viewing area is uniform, which we have already established as a dubious assumption. A further discussion of this method and simulation results are given in Section 5.5.

5.4 Simulation Details

5.4.1 Input Parameters

Simulations were run to study the bistatic observation method of determining vertical wind speed. The HWM, IRI-07, and NRLMSISE-00 climatological models were used to provide realistic conditions for simulating one night of data collection. The night of 27-28 June 2005 was chosen as a solstice night during moderate solar conditions. Every half-hour, the temperature profile, airglow emission profile, and wind velocity at 250 km were determined. Using a 5-km step size along the LOS of each observation, an image was generated for each temperature and scaled by the intensity of the airglow emission at that altitude. The images were summed into a composite image representing what an FPI would record at that time. The horizontal wind at 250 km was used for each image because the HWM estimates of the wind gradient at these altitudes, and especially at near-equatorial locations, are only approximate [17]. These images are a noiseless representation of the images that an

FPI would produce when observing this wind field.

5.4.2 Image Intensity and Noise Balance

After the entire set of night images was generated for each look direction of each instrument, the images were processed and analyzed. The intensity of each image was scaled to resemble data collected by MiniME. This was done by comparing the U_2 value that the LM NLLMS routine returned for real data to the value it returned for different intensity levels of simulated images. For each image, every pixel was given a base magnitude of 170 analog-to-digital units (ADU) to match typical values, and the maximum at the fringe peaks was linearly reduced from 173.5 to 171 ADUs as the night progressed to approximate the actual loss of intensity. Lastly, noise levels equivalent to those found in real data were added. Gaussian dark noise with a variance of 0.1 ADUs and Gaussian shot noise with a variance of the square root of the pixel magnitude was added. Figure 5.11 shows the intensity resulting from images from a night of data collected by MiniME and the intensity levels used in our simulations. The intensity of the real data shown is that of a single look direction. The simulated intensity is more uniform and is higher than the real data at the beginning of the night. This was done because it is more representative of normal intensity fluctuations. At one point during this night, one look direction of the MiniME data generated a maximum intensity of 0.58. Therefore, the maximum of the simulated intensity is still relatively modest at 0.41. The lower MiniME data intensity from 18:00 to 21:00 LT is most likely due to partial cloud cover. The minimum intensity of the simulated images is 0.098, near the minimum seen in real data of 0.064; so this is a good approximation of typical values seen.

The images were analyzed using the method explained in Section 4.2; and the method outlined in Section 5.3 was used to extract the zonal, meridional, and zenith wind speeds.

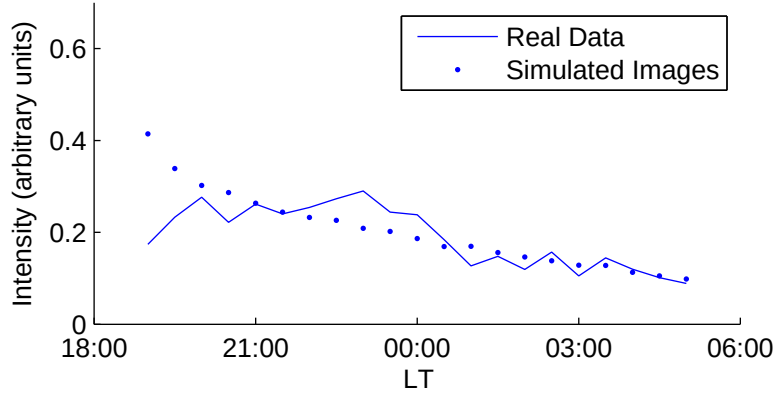


Figure 5.11: Intensities of real data and simulations for an entire night are shown. The value shown is the variable U_2 defined in Equation (4.14).

5.5 Simulation Results

5.5.1 Simulation of Observation and Analysis Techniques

A series of Monte Carlo (MC) simulations were run to test the observation techniques developed above. In each MC simulation, the process of adding noise to images and analyzing them described in the previous section was repeated 100 times in order to obtain statistical data. The MC simulation was run three times: once with no vertical wind, once with a 10 ms^{-1} vertical wind, and once with a 20 ms^{-1} vertical wind. Changing the vertical wind did not result in any statistically significant change in results, suggesting that our method will perform equally well with any reasonable, constant, and uniform vertical wind velocity. For the sake of brevity, only the results from the simulation with a 10 ms^{-1} vertical wind are shown.

Figure 5.12 shows the raw LOS data that was used to simulate the images with the results of one simulation. Here we can observe the potential problem with averaging LOS measurements to determine vertical wind according to the method explained in Section 5.3.3. For this method to be accurate, the Site A northern measurement should be equivalent to the Site B southern measurement reflected across the vertical wind component of the measurement, as should the Site A southern measurement of the Site B northern measurement. Figure 5.13 shows that the average of each pair varies from 8 to 9 ms^{-1} instead of the correct value of 10 ms^{-1} for this simulation. Since this is an error of only $1\text{--}2 \text{ ms}^{-1}$, averaging all measurements together does

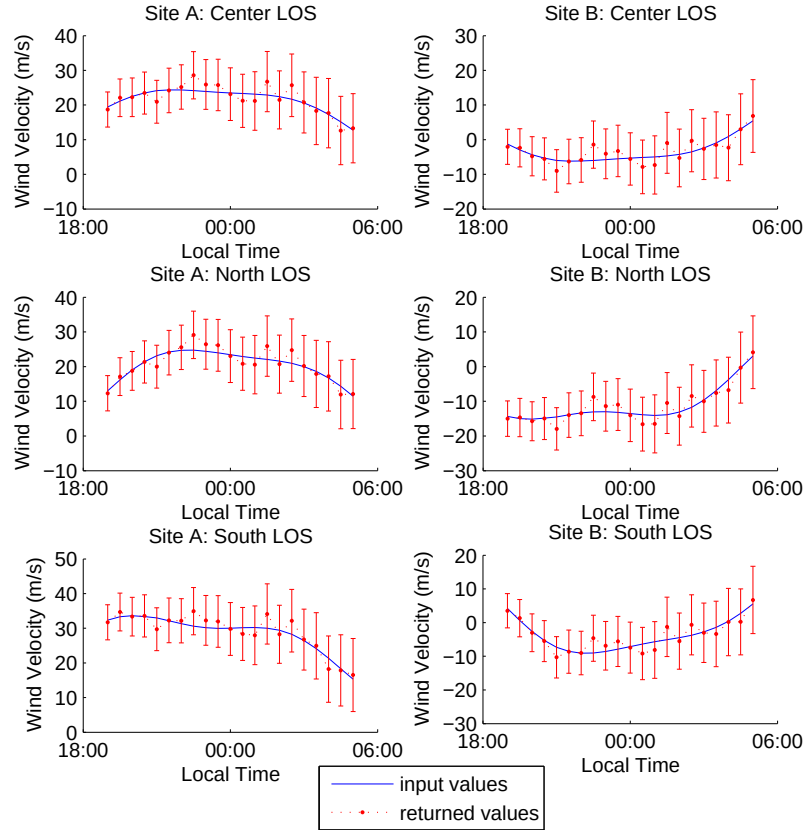


Figure 5.12: The raw LOS data obtained from the HWM are shown with a vertical wind of 10 ms^{-1} . The results of a single simulation of the FPIs are also shown with the uncertainty statistics produced by the LM NLLMS algorithm.

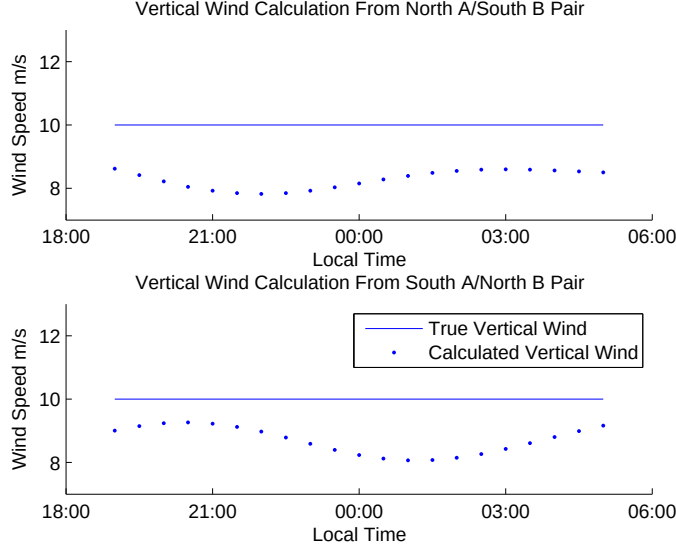


Figure 5.13: The averages of the LOS data from HWM are shown. The average of the northern measurement from Site A and the southern measurement from Site B is the top panel, and the average of the southern measurement from Site A and the northern measurement from Site B is the bottom panel.

not introduce a significant level of error in this simulation, and the loss of accuracy in exchange for precision is well worth it. But it is important to recognize the potential problems with this method. When real data are available, it must be analyzed to determine if this method for improving the estimate of the vertical wind calculation is trustworthy. It could be argued that this method nullifies the advantages of using a BFPIS altogether because full vector wind information at multiple locations has no worth if the assumption is made that the wind at these two points is identical. However, this assumption is only made to estimate the vertical wind. If making this assumption results in a substantial improvement in precision with little loss of accuracy, it is still worthy of use. Additionally, the BFPIS still yields full vector wind information, which cannot be determined using a single FPI.

Figure 5.14 shows the estimated horizontal winds based on the HWM data along with the estimations from one iteration of the simulation calculated using the single vertical wind and the averaged vertical wind. The error bars are shown for the calculation based on the inline vertical wind calculation and are calculated from the parameters returned from the LM NLLMS algorithm. It can be seen that the errors in the wind calculations are small enough to not

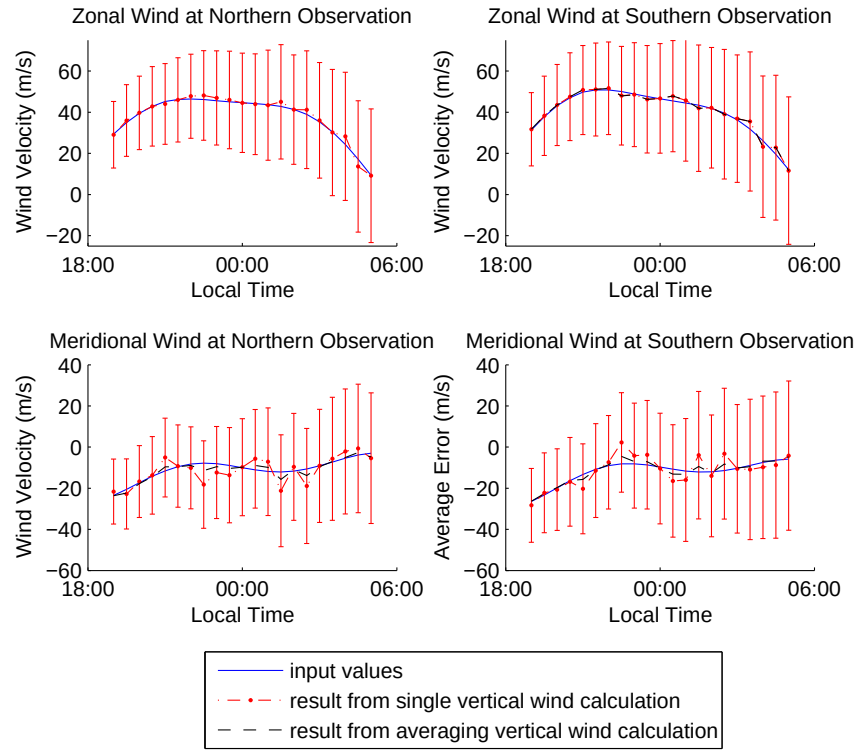


Figure 5.14: The simulated zonal and meridional winds and the northern observation region and the southern observation region are shown. Each figure shows the HWM data with the calculations based on the single vertical wind calculation and the averaged vertical wind.

have a dramatic effect on the resultant estimates of horizontal winds. It can also be observed in Figure 5.14 that the errors in the zonal wind estimation are much smaller than those in the meridional wind estimation and that there is no improvement in the precision of the zonal wind estimation when the averaged vertical wind is used. The reasons for this are discussed in detail below. The error bars shown are calculated from the parameters produced by the LM NLLMS algorithm. The error bars in Figure 5.14 are much larger than those in Figure 5.12 because the process of calculating the horizontal wind vectors magnifies error.

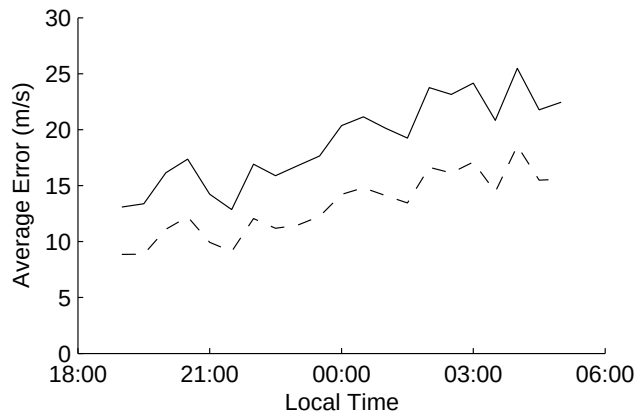


Figure 5.15: The standard deviation of vertical wind speed estimates as determined by a Monte Carlo simulation of 100 runs is shown. The vertical wind is assumed to be a constant for the entire field, so only one plot is shown. The solid line is the error resulting from using the single vertical wind calculation and the dashed line is the error resulting from the averaging method.

The average error in each wind component using each of the two vertical wind calculation methods is shown in Figures 5.15 and 5.16. As the night goes on the error increases, as is expected because of the falling image intensity. The averaging vertical wind calculation method reduces the error in zenith wind calculation from $\pm 20 \text{ ms}^{-1}$ to $\pm 13 \text{ ms}^{-1}$ on average. The averaging method reduces the average error on the meridional wind calculation from $\pm 4 \text{ ms}^{-1}$ to $\pm 1.6 \text{ ms}^{-1}$. However, this approach has no effect on the average error of the zonal wind calculation, which stays steady at 1.0 ms^{-1} . This was not initially expected. A look at the location of the instruments and the angles of their observation reveals that the meridional winds are calculated by scaling and summing the vectors from each instrument, while

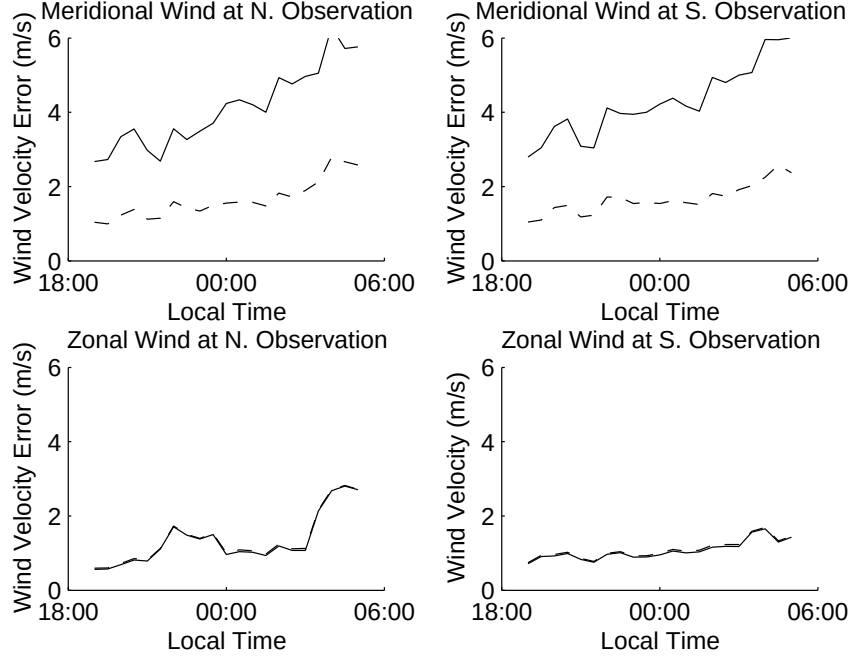


Figure 5.16: The standard deviation of zonal and meridional wind speed estimates as determined by a Monte Carlo simulation of 100 runs is shown. The zonal and meridional wind errors are shown for the northern site and the southern site. The solid line in each figure is the error resulting from using the single vertical wind calculation and the dashed line is the error resulting from the averaging method.

the zonal winds are calculated by scaling and subtracting the vectors from each instrument. The subtraction causes the effect of the vertical wind to be minimized. There are two increases in the zonal wind error at the northern observation that are not present in the southern observation. These are the result of outlying data points. If the simulation were run for a larger number of iterations, those anomalous points should be reduced in size. It is also surprising that the high magnitude of error in the vertical wind estimation does not propagate through to the horizontal wind measurements more strongly. Looking more closely at this, we observe that we are using elevation angles in the range of 52 to 64 degrees (see Table 5.1. Figure 4.2 shows that angles in this range result in large uncertainties for vertical wind calculation. However, large elevation angles result in higher accuracy in horizontal wind measurements. Equations (5.26) and (5.27) show that the vertical wind factor that contributes to the horizontal wind calculation is multiplied by $(1 - \sin \alpha)$. As the elevation angle increases towards $\pi/2$, the effect of the vertical wind is

reduced. So the large vertical wind error and the small horizontal wind error are the effect of the large elevation angles being used.

Finally, Figures 5.15 and 5.16 also highlight the advantage of using averaging to determine the vertical wind. By assuming that the wind field is uniform, we are able to significantly reduce the uncertainty in our calculations at the cost of being unable to identify variability in the wind field. Alternatively, using only the inline measurement to calculate the vertical wind allows us to observe changes in horizontal wind velocities between different observation regions, but also results in greater uncertainties. One way to improve this option is to move the instrument sites further away so that the elevation angles are reduced, thereby decreasing measurement uncertainties without a loss of spatial resolution.

CHAPTER 6

CONCLUSION

6.1 Conclusions

In this thesis, we explained the importance of thermospheric winds in the development of equatorial plasma bubbles and how the Fabry-Perot interferometer can be used to measure these winds. Then we outlined a method for using two FPIs in a bistatic formation to improve measurement precision and obtain full vector wind information at multiple locations. Finally, we simulated this method to demonstrate its usefulness and feasibility.

In Chapter 2, we introduced basic concepts of the mechanics of the upper atmosphere. We explained the basic composition of the thermosphere and ionosphere. We also described the F region dynamo and showed how thermospheric winds have a significant contribution to this phenomenon. We then went on to discuss how equatorial spread F disrupts the propagation of low-power trans-atmospheric radio waves and the potential effects that thermospheric winds have on the development of ESF. In this chapter, we also reviewed the chemical processes that produce the 630.0-nm nighttime airglow emission. Finally, we use climatological models to generate typical electron density profiles, 630.0-nm emission profiles, and the typical concentrations of chemicals relevant to the 630.0-nm emission.

In Chapter 3, we discussed the details of the Fabry-Perot interferometer. We began with a brief history of the development of the first FPI. We then reviewed the physical principles that the FPI is based on and explained the details of the FPI etalon. This chapter concluded with a detailed description of the design of the MiniME FPI that is employed for the RENOIR project.

In Chapter 4, we continued to explain how to apply an FPI to atmospheric science. We described typical observation strategies for a single FPI used for estimation of neutral winds. Then, we explained the Levenberg-Marquardt

nonlinear least squares algorithm that we use for analyzing data images. We concluded Chapter 4 with a discussion of the inherent limitations of using an FPI to estimate thermospheric winds.

In Chapter 5, we explained the development of a bistatic FPI system and its advantages. We began by highlighting the advantages of a BFPIS over the use of a single FPI. We then explained how the observation strategy was chosen and how the proper azimuthal and elevation angles for this strategy were determined. Next, we introduced a method for determining full vector wind information from the FPI data. Finally, we described simulations that were run to test this implementation of the BFPIS and present the results of these simulations.

6.2 Future Work

The most glaring omission in this thesis is the comparison of simulations and theories to results from real data. As of the writing of this thesis, one of the RENOIR FPIs in Brazil has not been functioning properly since its installation in August of 2009 despite several attempts to repair it. When this issue is resolved, the BFPIS will be implemented and data will be gathered. Analysis of this data will inform the reliability of using uniform wind field assumptions. We will then be able to compare the results of using only the center measurements to determine the vertical wind speed to the results of averaging all measurements together. It will also enable researchers to write more accurate models of the thermosphere. Used in conjunction with the other instruments installed in Brazil, this data will provide evidence to support or disprove the theoretical effects of neutral winds on the RTI and the formation of EPBs.

Our simulations showed that the high elevation angle used for the center observation will cause the error on a LOS measurement to be multiplied by a factor of five when determining the vertical wind. According to Equation (5.23) reducing the elevation angle from 64° to 53° would reduce the multiplicative factor to 2.5. This would require the instruments being located ~ 375 km apart instead of 230 km. Therefore, any future systems would ideally be deployed with a greater distance between instrument sites.

The derivation here assumed that the instrument bias β is a constant,

which is far from the case. A more advanced technique would involve feeding the vertical wind measurement back into the model to improve the calculation of the instrument bias.

A significant part of the advantage of the BFPIS over the use of a single FPI is that of the reduction of the assumptions necessary to obtain wind estimates. A test case using data from HWM suggests that the proposed averaging method of determining vertical wind could be quite useful. However, this cannot be known for certain until real data is obtained. This data can be evaluated using only the center measurement to determine the vertical wind. The resulting wind speed estimates can then be compared to determine if the horizontal winds are similar enough to warrant the use of averaging.

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