RICE UNIVERSITY

THE VELOCITIES OF HF-INDUCED SHORT SCALE STRIATIONS

by

ANTHEA JANE COSTER

A THESIS SUBMITTED IN PARTIAL FULFILLMENT OF THE REQUIREMENTS FOR THE DEGREE

MASTER OF SCIENCE

APPROVED, THESIS COMMITTEE:

W. E. Gordon, Professor of Space Physics and Astronomy and of Electrical Engineering

Chairman

Joseph W. Chamberlain, Professor of Space Physics and Astronomy

Paul A. Cloutier, Professor of Space Physics and Astronomy

HOUSTON, TEXAS

APRIL 1981
Short scale striations are among the phenomena that occur when the ionosphere is heated by high frequency (HF) radio waves with ordinary mode polarization. These striations are electron density irregularities aligned along the magnetic field lines in and near the heated region of the ionosphere. They are a direct consequence of the heating and disappear within seconds after the heater has been turned off. The striations are detected to be moving in the heated region. Experiments were performed in June 1977 to investigate short scale striations. The experiments combined the use of a portable 50 MHz radar located on the island of Guadeloupe with heating facilities at the Arecibo Observatory. The 50 MHz radar had a line of sight that was orthogonal to the magnetic field lines in the F region above Arecibo and was used to detect the striations and their velocities. The 430 MHz radar at Arecibo was used for additional diagnostics. Varying the power of the HF radio wave did not appear to affect the striation velocities. Instead, their velocities appeared to be well correlated with general F-region ionization drifts. On two evenings, shortly after ionospheric sunset, dramatic changes in the striation velocities were observed. These F-region velocity disturbances occur
in conjunction with the intensification of existing sporadic-E regions. Only during these disturbed time periods were reflections from sporadic E detected at greater than 5 MHz. A mechanism whereby the sporadic-E regions cause the alternate coupling and decoupling of the E and F regions in the early evening hours is presented. This mechanism can explain changes in local F-region electric fields which would cause the velocity disturbances.
ACKNOWLEDGEMENTS

Several people aided me in the production of this thesis; it is a pleasure for me now to acknowledge their contributions. First, I would like to thank my thesis advisor, Professor William E. Gordon, for his generous support and guidance; his comments and suggestions during the developing stages of this thesis were invaluable. Mr. Warner Ecklund and Mr. David Carter of the National Oceanic and Atmospheric Administration (NOAA) took special pains to make the Guadeloupe data available to me. Thanks also go to Drs. Frank T. Djuth and Daniel A. Fleisch, both formerly with Rice University, for their patient instruction in the background material of this research. In addition, I benefited greatly from the technical writing instruction provided by Dr. Linda Driskill of Rice University. Special recognition goes to my friend, Faye Boudreaux, and my husband, Steve Boswell, for their constant encouragement which helped me face the task on a day-to-day basis. Finally, I would like to express my sincere appreciation to my parents, Dr. and Mrs. H. P. Coster, for their unwavering faith in me and for all the proofreading dinners that we shared.

This research was partially supported by the Atmospheric Research Section of the National Science Foundation under Grant Nos. ATM76-15550 and ATM79-09088. A portion of this research was conducted at the National Astronomy and Ionospheric Center in Arecibo, Puerto Rico, which is operated by Cornell University with the support of the National Science Foundation.
# TABLE OF CONTENTS

<table>
<thead>
<tr>
<th>CHAPTER 1 - INTRODUCTION</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.1 Parametric Decay Instability</td>
<td>3</td>
</tr>
<tr>
<td>1.2 Resonant Instability</td>
<td>6</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>CHAPTER 2 - THE EXPERIMENT</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>2.1 Geometry</td>
<td>10</td>
</tr>
<tr>
<td>2.2 Facilities</td>
<td>11</td>
</tr>
<tr>
<td>2.2.1 Arecibo Observatory</td>
<td>11</td>
</tr>
<tr>
<td>2.2.2 Guadeloupe Radar</td>
<td>12</td>
</tr>
<tr>
<td>2.3 Beam Intersection</td>
<td>13</td>
</tr>
<tr>
<td>2.4 Data Processing</td>
<td>15</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>CHAPTER 3 - F-REGION DYNAMICS</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>3.1 F-Region Characteristics</td>
<td>18</td>
</tr>
<tr>
<td>3.2 Derivation of Transport Velocities</td>
<td>20</td>
</tr>
<tr>
<td>3.3 Conductivities</td>
<td>23</td>
</tr>
<tr>
<td>3.4 F-Region Electric Fields</td>
<td>25</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>CHAPTER 4 - THE GUADELOUPE DATA</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>4.1 Relationship of the AFAS Velocities to HF Transmitter Power</td>
<td>30</td>
</tr>
<tr>
<td>4.2 Comparison of AFAS Velocities to Natural F-Region Ionization Drifts</td>
<td>32</td>
</tr>
<tr>
<td>4.3 Velocities of Large and Short Scale Striations</td>
<td>35</td>
</tr>
<tr>
<td>4.4 Atypical Data</td>
<td>37</td>
</tr>
<tr>
<td>4.5 E-Region Observations</td>
<td>52</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>CHAPTER 5 - SUMMARY AND CONCLUSIONS</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>REFERENCES</td>
<td>58</td>
</tr>
</tbody>
</table>
CHAPTER 1
INTRODUCTION

In recent years, ionospheric heating experiments have been conducted to study a variety of plasma phenomena. Powerful high frequency (HF) radio waves have been used to transmit energy in a conical beam up to the altitude of reflection in the ionosphere. Reflection occurs at the altitude where the frequency of the HF wave is equal to the plasma frequency. In the reflection region, energy from the HF wave is absorbed by the local plasma, thus disturbing the normal energy balance in this region. The processes by which energy is transferred are currently under intensive investigation as they are related to the problem of plasma heating in laser fusion.

The discovery of short scale striations has been one of the exciting results of ionospheric heating experiments. Short scale striations were first observed in the ionosphere during experiments conducted at the Platteville heating facility in Colorado. The striations were detected in the heated region of the ionosphere when the Platteville heater was operating with ordinary mode polarization at frequencies close to the F-region critical frequency (5-10 MHz). The critical frequency is the highest plasma frequency at any given time in the ionosphere. Electromagnetic waves below the critical frequency are reflected in the ionosphere.
Short scale striations, sometimes known as artificially produced field-aligned striations (AFAS), are electron density irregularities that are generated in the heated region of the ionosphere. They are a direct consequence of the heating and disappear within seconds after the heater has been turned off. These density perturbations, or striations, are aligned along the earth's magnetic field. The striations are highly aspect sensitive—that is, they scatter radiation in certain preferential directions. Radar stations have to be located in positions with the correct geometry in order to detect the striations. Radar stations used in the Platteville experiments detected the striations by operating in the frequency range of 15-435 MHz (Minkoff et al., 1974). This frequency range corresponds to a scale of the AFAS perpendicular to the magnetic field from 10 m to .3 m. Based on the results at Platteville, the AFAS have a calculated minimum length along the magnetic field of 200 m (Fialer, 1974).

The striations appear to be in motion within the heated volume. They are observed to move across (or orthogonal to) the magnetic field lines with velocities varying from 2 m/s to 75 m/s. In the F region of the ionosphere, velocities of electrons perpendicular to the magnetic field are directly proportional to the perpendicular component of the electric field in the local ionosphere. The AFAS velocities may therefore prove to be a valuable tool in furthering our understanding of F region electric fields. This thesis is an investigation of the velocities of short scale striations.
The AFAS velocities appear to be related to the local ionospheric conditions and are not believed to be a direct consequence of the heating of the plasma. Nevertheless, it is useful at this point to present a brief summary of the different theories explaining the generation of the striations, as this introduces the unanswered question concerning their formation.

When short scale striations were first discovered, no theory had predicted their formation. Since then, several theories have arisen, all of which postulate that the AFAS form through thermal interactions between excited plasma waves. The different theories can be broadly categorized into two classes depending on the instability used to generate the striations. One class of theory uses the parametric decay instability which excites longitudinal plasma waves propagating parallel to the magnetic field. The other class of theory invokes the resonant instability which excites plasma waves propagating perpendicular to the magnetic field. The following is a description of these two general classes of theory.

1.1 **Parametric Decay Instability**

The first class of theory (Fejer and Lee, 1978; Perkins, 1974) assumes that the AFAS are generated as a by-product of the parametric decay instability. This instability involves the parametric excitation of natural plasma waves, propagating parallel to the magnetic field, by the HF heating beam. One can think of parametric excitation as a nonlinear process of energy transfer.
According to Fejer (1979), the most important type of nonlinear interaction in the excitation of parametric instabilities in the ionosphere is the three-wave interaction. Here a strong wave characterized by angular frequency $\omega_1$ and wave number $k_1$ causes the growth of two weaker waves with angular frequencies $\omega_2$ and $\omega_3$ and wave numbers $k_2$ and $k_3$ such that the frequency and wave number matching relations $\omega_1 = \omega_2 + \omega_3$ and $k_1 = k_2 + k_3$ are satisfied.

The parametric decay instability is an example of this three-wave interaction process. Two electrostatic daughter waves, one a low frequency ion acoustic wave, the other a high frequency electron plasma wave, are excited by the high frequency pump wave. In the ionosphere, this process requires the initial presence of some energy in the electron plasma wave to modulate the pump wave. The modulation of these two waves produces a low frequency wave. If the phase velocity of the low frequency wave falls within the ion velocity distribution, a transfer of energy occurs. The low frequency wave is Landau damped, giving up its energy to a low frequency ion acoustic wave. The growing ion acoustic wave, in turn, beats back with the high frequency waves to induce another transfer of energy—this time from the ion acoustic wave to the electron plasma wave. Through this process, the high frequency electron plasma wave grows parametrically at the expense of the original high frequency pump wave.

Initially the parametric instability is excited by the electromagnetic pump wave of the heating beam in the ionosphere. The instability is excited only if the pump wave is above threshold which, according to Fejer (1978), is approximately 0.1 to 0.5 V/m for typical
ionospheric conditions. The pump wave does not, however, have to be an electromagnetic wave. An electrostatic wave can be equally effective in the excitation of this instability. Therefore, if the daughter high frequency wave grows in amplitude to the point where it too satisfies the threshold conditions, then it can also excite the parametric decay instability. A second set of daughter waves is thereby generated. A cascading process to slightly lower frequencies is set into motion until the final high frequency daughter wave is no longer above threshold.

This cascade of spectral energy down in frequency is responsible for the saturation spectrum of the parametric decay instability. The saturation spectrum is composed of high frequency electron plasma or Langmuir waves which form at relatively small angles (less than 25°) to the geomagnetic field (Fejer and Juo, 1973; Perkins et al., 1974). This spectrum of high frequency Langmuir waves, at small angles to the magnetic field, is essential to the arguments of Fejer and Lee and of Perkins for the formation of short scale striations.

To produce the striations, another three-wave interaction process is used. Here two of the electron plasma waves couple to produce a low frequency wave. This low frequency wave is not subject to the same force that drives the parametric decay instability. The parametric decay instability is driven by the ponderomotive force, or the force of radiation pressure. In this second three-wave interaction, ohmic heating terms become important. This is because the geometry of the low frequency wave has scale lengths longer than the electron gyro-radius (approximately 2 cm in the F region) across the magnetic field
and longer than the mean free path (approximately 0.2 km) along the magnetic field. In this type of geometry, pressure fluctuations resulting from electron heating exceed the radiation pressure (Perkins, 1974).

The Perkins (1974) theory differs from the Fejer and Lee (1978) theory in the interpretation of how the thermal heating generates the striations. Fejer and Lee argue that the striations form due to collisional dissipation of the heating. Perkins, on the other hand, suggests that the thermal heating induces another parametric instability--thermal coupling. The striations are then formed as a by-product of this instability.

1.2 Resonant Instability

The second class of theory (Vaskov and Gurevich, 1977; Fejer and Das, 1978; Grach et al., 1978) postulates that another instability, the resonant instability (or the slightly different thermal parametric instability), is responsible for the production of short scale striations. Here, small density perturbations aligned along the magnetic field are assumed to preexist in the plasma. Thermal fluctuations within the plasma could cause small density irregularities to form. These naturally occurring perturbations would be field aligned in the F region of the ionosphere because, in the F region, the electronic diffusion rates are much greater along the magnetic field lines than across the magnetic field lines. The resonant instability describes the process through which the HF electric field magnifies these small density perturbations into the striations.
The mechanism of the resonant instability is the following. The HF electric field sets up polarization currents within the small density perturbations. The polarization currents then excite the natural longitudinal oscillations of the plasma. Unlike the parametric decay instability, these plasma oscillations are electrostatic plasma waves that propagate perpendicular to the magnetic field. While plasma waves propagating parallel to the magnetic field are at the plasma frequency, $\omega_p$, electrostatic waves propagating perpendicular to the magnetic field are at the upper hybrid frequency, $\omega_h = \omega_p^2 + \omega_c^2$, where $\omega_h$ = upper hybrid frequency, $\omega_p$ = plasma frequency, and $\omega_c$ = electron gyrofrequency. The equations for the gyrofrequency and the plasma frequency are (in mks units):

$$\omega_c = \frac{eB}{m}$$

$$\omega_p = \left( \frac{Ne^2}{\varepsilon_0 m} \right)^{1/2}$$

$e$ = electric charge
$B$ = magnetic field of the earth
$m$ = mass of the electron
$N$ = electron number density
$\varepsilon_0$ = permittivity constant of free space

The HF wave electric field is thus scattered by the natural density perturbations into the electrostatic plasma waves. The scattering of the HF electric field is greatest at the altitude where the frequency of the HF wave matches the upper hybrid frequency of the
plasma. In the vicinity of the upper hybrid frequency, the group velocity of the excited longitudinal plasma waves is small. In other words, these plasma waves behave similarly to standing waves. As a result, the longitudinal plasma waves dissipate an appreciable fraction of their energy into the small density perturbations.

The electron density irregularities are heated by this dissipation of plasma wave energy. The heating enhances the density perturbations since the electronic diffusion rate and the electronic thermal conductivity are both greater along the magnetic field lines than across the magnetic field lines. Finally, the enhanced density perturbations increase the scattering of the HF electric field and complete the cycle of the resonant instability.

The generation mechanism for the AFAS remains an unsolved problem for plasma physicists. To date, neither class of theory has been proved or disproved. Fejer (1978) has suggested that the true mechanism for the formulation of the striations may be a combination of both instabilities.

So far no theoretical model for the striations has been developed that incorporates a picture of their movement within the heated region. Motion across the magnetic field lines in the F region requires that a gradient in the electric potential be maintained over time scales longer than a second. Therefore, it is assumed that the velocities of the striations are related to electric fields that exist naturally in the ionosphere. Since F-region dynamics are still not completely understood, the striation velocities are of interest to ionospheric physicists.
CHAPTER 2

THE EXPERIMENT

During a two-week period of June 1977, a series of experiments was conducted to further investigate short scale striations. The experiments combined the use of heating facilities at the Arecibo Observatory in Puerto Rico with a portable 50 MHz radar unit located on the French island of Guadeloupe. The radar was used to observe irregularities spaced 3 m apart that formed in the heated region above Arecibo. The observations reported in this thesis were obtained from this set of experiments.

This thesis will address the following specific questions concerning the velocities of short scale striations detected by the Guadeloupe radar:

1) What effect does the power of the HF radio wave have on the short scale striation velocities?

2) How do the velocities of the striations compare to the natural F-region drifts?

3) Can the velocities of the AFAS be used to learn more about natural phenomena in the ionosphere?

A brief description of the geometrical considerations, the experimental facilities, and the data processing of these experiments will now be presented.
2.1 **Geometry**

Certain geometrical constraints played a crucial role in the selection of the radar's site on the island of Guadeloupe. Short scale striations are elongated in the direction of the geomagnetic field in the heated region of the ionosphere. To picture the striations, one can think of very long, thin antennas lying along the earth's magnetic field lines. When scattering radio waves, the striations act as nearly perfect specular reflectors. In other words, the striations, as reflectors, largely obey the rule that the angle of incidence must equal the angle of reflection.

In order to detect a return signal, the radar receiver must have a line of sight to the striations in the heated region that satisfies the angular conditions of the reflection rule. In the case of a monostatic radar system, such as that used at Guadeloupe, the reflection rule is satisfied when the radar has a line of sight that is orthogonal to the magnetic field lines at the altitude where the striations form.

The site of Guadeloupe was chosen to satisfy the geometric constraints discussed above. Guadeloupe has a line of sight that is perpendicular to the geomagnetic field lines over Arecibo at an altitude of 220 to 250 km. The 50 MHz radar located on Guadeloupe was therefore in an excellent position to detect the striations that formed between these altitudes. In actuality, the Guadeloupe radar detected the striations over height intervals from 210 to 290 km (Frey, 1980). This is because the striations are not infinitely long and are therefore not perfect specular reflectors.
2.2 Facilities

2.2.1 Arecibo Observatory

The Arecibo Observatory is located about 15 km south of the city of Arecibo in Puerto Rico and is at geodetic latitude of 18°20'46" north and longitude of 66°45'11" west. The magnetic dip angle at Arecibo is 50° and the magnetic declination is 8°30' west. The antenna at the Arecibo Observatory is a spherical dish with an opening diameter of 305 m. The Observatory is operated by Cornell University with the support of the National Science Foundation.

The Arecibo Observatory was used during these experiments to heat the ionosphere and to measure the plasma motion. Powerful high frequency radio waves were produced by an HF transmitter. The transmitter was connected to an antenna feed that was mounted above the antenna dish. The transmitter could produce 140 kw of power, although, due to periodic arcing of the feed, 100 kw was a more typical output. For these experiments, the HF transmitter operated at frequencies from 4-7.8 MHz. Frey (1980) calculated that the half-power beamwidth in degrees for the HF wave is 7.4° for 7.8 MHz, 8.3° for 6.875 MHz, and 14° for 4.070 MHz.

Throughout the experiment, the ionosphere was monitored by two sweep frequency ionosondes. One is located at the Arecibo Observatory and the other at Los Caños, which is approximately 11 km north of the Observatory. The Los Caños ionosonde was operated at 5-minute intervals.
2.2.2 Guadeloupe Radar

The portable 50 MHz radar situated on the island of Guadeloupe was used to detect 3 m irregularities in the F region over Arecibo. For the experiments described, this radar was located at the geodetic latitude of 16°15'11" north and longitude of 61°11'53" west. The radar was supplied and operated by the Aeronomy Laboratory of the National Oceanic and Atmospheric Administration and the site at Tarara, near Point-a-Pitre, was made available by the Institut de Physique du Globe in Paris.

The Guadeloupe radar consisted of four collinear strings of half-wave dipoles, each string containing forty dipoles. The radar operated at 50 MHz with a peak power of 15 kw and pulse lengths of 20-200 μs. An interpulse period of 10 μs was used. The following are the radar's calculated (or measured) antenna parameters:

- Horizontal two-way beamwidth $2^\circ$
- Vertical two-way beamwidth $10^\circ$
- Elevation angle $16^\circ$
- Azimuth angle $292^\circ$
- Collecting area $1200 \text{ m}^2$
- Antenna gain 420

The receiver noise temperature was 1000°K and the sky noise temperature was estimated to be about 6000°K.

The Guadeloupe radar detects only a fraction of the AFAS produced in the heater region. The 50 MHz wave is scattered by a multitude of striations spaced at varying intervals perpendicular
to the magnetic field. The vast majority of this scattered radiation is eliminated through the destructive interference of waves being scattered from different portions of the heated region. Only striations that have scale sizes of 3 m (half the radar's wavelength) will backscatter waves that constructively interfere. The 50 MHz radar, therefore, selects out the AFAS spaced 3 m apart across the magnetic field lines. The many scatterers combine incoherently to produce the signal the radar measures.

2.3 Beam Intersection

The Guadeloupe radar was capable of detecting only the striations that are formed within the intersection of its radar beam and the Arecibo heating beam. A description of this region of intersection follows. Assume that the frequency of the HF transmitter is adjusted so that the heated region forms at an altitude of 230 km above Arecibo. It can be shown that, at this altitude, the Arecibo beam will heat a disk-shaped volume of approximately 30-50 km in diameter and an estimated 10-15 km in thickness. The Guadeloupe radar intersects the heated volume at the ranges of 680-700 km from Guadeloupe. The Guadeloupe beam is shaped like an elliptical cone. In the region of intersection its 10° two-way vertical beam will view 120 km of altitude while its 2° two-way horizontal beam will view 10 km (N-S). An illustration of the intersection between these two beams is presented in Figure 2.1, Frey (1980).

The Guadeloupe radar observes the striations at different ranges within the heated volume. Along the path of the Guadeloupe
INTERSECTION OF THE 50 MHZ RADAR BEAM WITH THE HF-BEAM

RADAR ELEVATION ANGLE = 16°
RADAR HALFPower BEAMWIDTH = 10° (VERTICAL)

HF HALFPower BEAMWIDTH = 7° to 14°
FOR HF FREQUENCIES = 8 MHZ to 4 MHZ

FIG. 2.1. FREY (1980)
beam, there is a resolution length of 7.5 km when a pulse length of 50 µs is used. Using the above assumptions about the region of intersection, the Guadeloupe radar detects a volume that is approximately a 10 km cube for each of its range gates. When we speak of a range in this thesis, we are referring to a 10 km³ volume.

2.4 Data Processing

At Guadeloupe the Doppler shifted return signal was first mixed with an 80 MHz signal. A 30 MHz signal resulted that was then amplified and mixed with 30 MHz sine and cosine waves. Two baseband signals are produced by this process representing the inphase and quadrature components of the return signal. After the higher frequencies were filtered out, these two signals were recorded on analog tape. The tape was digitally sampled and analyzed at a later point in time. What follows is a discussion of the data sampling process.

Consider a signal sent out from the Guadeloupe radar at time \( t_0 \). The return signal will be sampled at time \( t = t_0 + t_1 \), where \( t_1 \) is called the gate delay. The gate delay corresponds to the first range being sampled, through the relationship \( d = ct_1/2 \), where \( d \) is the distance to the first range and \( c \) is the velocity of light. The return signal is then sampled for additional ranges with a period called the gate width. If the pulse is sampled \( N \) times, for \( N \) different ranges, then the sampling is completed at time \( t = t_0 + t_1 + (N-1)t_2 \), where \( t_2 \) is the gatewidth. At a later time, \( t = t_0 + IPP \), where IPP equals the interpulse period, the entire process is repeated. An illustration of this procedure is provided in Figure 2.2.
The velocity, $v$, of the short scale striations was extracted from the frequency domain representation of the signal using the Doppler shift equation, $2v/c = \Delta f/f$, where $\Delta f$ is the Doppler shift and $f$ the 50 MHz frequency. To shift from the time to the frequency domain, a Discrete Fourier Transform (DFT) was performed on the data using a Fast Fourier Transform (FFT) algorithm. The FFT algorithm requires that the number of samples input be a power of two.

In these experiments a 64-point FFT program was implemented on each of the 16 ranges. The 64 samples in the time sequence were separated by the interpulse period (IPP). The result was a frequency spectrum for each range. The bandwidth, $f_c$, of these spectra is determined by the IPP through the relationship $(IPP)^{-1} = 2f_c$. The standard Guadeloupe IPP was 10 msec, which results in a bandwidth of 50 Hz. Using the Doppler shift equation, the velocities represented by the frequency spectra have a range from $-150$ m/s to $+150$ m/s. To improve the statistics, 10 spectra from each range were summed. Each velocity spectrum is therefore an average over approximately 6 seconds.

Velocity spectra in this thesis are plotted in signal-to-noise ratios (S/N). A noise estimate for each spectrum was obtained by averaging 8 points on either end of the spectrum. This method of estimation was valid since the highest velocity detected by the Guadeloupe radar was on the order of $\pm 75$ m/s.
CHAPTER 3

F-REGION DYNAMICS

The object of this thesis is to present an interpretation of the velocities of short scale striations detected by the 50 MHz radar on the island of Guadeloupe. The Guadeloupe radar had a line of sight to the AFAS that was orthogonal to the magnetic field at the altitude where the striations form. Since the striations are variations of the electron density concentration, their velocities represent a measure of the motion of electrons at right angles to the magnetic field lines. To interpret the velocities, one must understand the forces that govern the movement of electrons across the field lines in the F region of the ionosphere. In this chapter a brief review of F-region dynamics will be presented with the aim of establishing a framework to guide our interpretation.

3.1 F-Region Characteristics

We will first present certain key characteristics of the F region that affect the motion of ionization. In the mid-latitudes, the F region can be loosely defined as the region of the ionosphere from the altitudes of 150 km to 600 km. The neutral atmosphere is predominantly oxygen and the ions are \( O^+ \). Gas densities in the F region are lower than the gas densities of other ionospheric regions.
Because of this the collision frequencies of both ions and electrons in the F region are much less than their gyrofrequencies. The gyro-frequency or cyclotron angular frequency is the frequency at which charged particles orbit about the magnetic field lines. Typical mid-latitude values for these frequencies at the altitude of 200 km as given by Rishbeth and Garriott (1969) are, when:

\[
\begin{align*}
\omega_{ci} &= \text{ion gyrofrequency} \\
\omega_{ce} &= \text{electron gyrofrequency} \\
\nu_{in} &= \text{collision frequency of ions with neutrals} \\
\nu_{en} &= \text{collision frequency of electrons with neutrals}
\end{align*}
\]

then,

\[
\begin{align*}
\omega_{ci} &= 190 \text{ rad s}^{-1} \\
\omega_{ce} &= 8.0 \times 10^6 \text{ rad s}^{-1} \\
\nu_{in} &= 4.1 \text{ s}^{-1} \\
\nu_{en} &= 130 \text{ s}^{-1}.
\end{align*}
\]

This notation will be used throughout this chapter. The gyroradii for this altitude are approximately 2 m for ions and 2 cm for electrons.

Using the above values, we obtain the following ratios of the collision frequency to the gyrofrequency in the F region:

\[
\begin{align*}
\frac{\nu_{in}}{\omega_{ci}} &= 2.2 \times 10^{-2} \quad \text{for ions} \\
\frac{\nu_{en}}{\omega_{ce}} &= 1.6 \times 10^{-5} \quad \text{for electrons}.
\end{align*}
\]
The approximation that $v_{en}/\omega_{ce} \ll v_{in}/\omega_{ci} \ll 1$ is extremely important in determining the transport velocities of charged particles in the F region. When these approximations hold, the background magnetic field exerts a dominating influence on the motion of charged particles across the magnetic field lines. It should be noted that $v_{in} \approx \omega_{ci}$ around 140 km and $v_{en} \approx \omega_{ce}$ around 80 km in the ionosphere. Below these altitudes, the above approximations do not hold.

3.2 Derivation of Transport Velocities

A complete treatment of the subject of transport velocities of ions and electrons in the F region can be found in an article by Kendall and Pickering (1969). For our purpose, however, a simpler approach will suffice. The following notation will be used in this general derivation of the transport velocities:

\[ \vec{v}_e = \text{electron velocity} \]
\[ \vec{v}_i = \text{ion velocity} \]
\[ \vec{U} = \text{neutral wind velocity} \]
\[ e = \text{electric charge} \]
\[ \vec{B} = \text{magnetic field of the earth} \]
\[ \vec{E} = \text{electric field} \]
\[ \perp = \text{component of a vector perpendicular to the earth's magnetic field} \]
\[ \parallel = \text{component of a vector parallel to the earth's magnetic field} \]
Following the discussion of Rishbeth and Garriott (1969), we will now derive the transport velocities. The macroscopic force equations for charged particles acted on by electric and magnetic fields are:

\[
m_i \frac{d\dot{v}_i}{dt} = e\dot{E} + e\dot{v}_i \times \dot{B} - m_i v_i n (\dot{v}_i - \dot{U}) \quad \text{for ions (3.1)}
\]

\[
m_e \frac{d\dot{v}_e}{dt} = -e\dot{E} - e\dot{v}_e \times \dot{B} - m_e v_en (\dot{v}_e - \dot{U}) \quad \text{for electrons (3.2)}
\]

We are concerned with the average drift velocities of electrons and ions, over periods much longer than \(v_i^{-1}, v_e^{-1}, \omega_i^{-1}, \omega_e^{-1}\), so we may assume that \(dv_i/dt = dv_e/dt = 0\).

Define the magnetic field \(B\) to be in the positive \(z\) direction in an \(x,y,z\) coordinate system. Using the notation that the applied force on the ions is \(\vec{F}_i = e\vec{E} + m_i v_i n \dot{U}\) and the applied force on the electrons is \(\vec{F}_e = -e\vec{E} + m_e v_en \dot{U}\), we can rewrite the force equations into their separate cartesian coordinates. For ions, these equations are:

\[
F_{ix} = m_i v_i n v_{ix} - e(v_{iy} + v_{iz})B \quad (3.3)
\]

\[
F_{iy} = m_i v_i n v_{iy} + e(v_{ix} + v_{iz})B \quad (3.4)
\]

\[
F_{iz} = m_i v_i n v_{iz} \quad (3.5)
\]
and for electrons, they are:

\[
\begin{align*}
 F_{\text{ex}} &= m_e v_{\text{en}}^2 + e(v_{\text{ey}} + v_{\text{ez}})B \\
 F_{\text{ey}} &= m_e v_{\text{en}} v_{\text{ey}} - e(v_{\text{ex}} + v_{\text{ez}})B \\
 F_{\text{ez}} &= m_e v_{\text{en}} v_{\text{ez}}
\end{align*}
\]

(3.6) (3.7) (3.8)

We can easily solve these equations for the motion of electrons and ions in the \( z \) direction, parallel to the magnetic field \((v_z = v_\parallel)\).

\[
\begin{align*}
 v_{\parallel i} &= \frac{eE_z}{m_i v_{\text{en}}} + U_z \\
 v_{\parallel e} &= -\frac{eE_z}{m_e v_{\text{en}}} + U_z
\end{align*}
\]

(3.9) (3.10)

Using the relations that \( \omega_{ci} = eB/m_i \) and \( \omega_{ce} = -eB/m_e \) and the conditions that \( v_{\text{in}}/\omega_{ci} \ll 1 \) and \( v_{\text{en}}/\omega_{ce} \ll 1 \) (which we found to be valid approximations in the \( F \) region), we find that the component of velocity perpendicular to the magnetic field is approximately:

\[
\hat{v}_L = \frac{\hat{E} \times \hat{B}}{B^2} \quad \text{for both ions and electrons.}
\]

(3.11)

In other words, charged particles can move at right angles to the magnetic field lines in the \( F \) region only if there exists an electric field at right angles to the field lines. This condition on the perpendicular component of velocity is extremely relevant to our discussion of the striation velocities.
Our equations for the transport velocities are in general agreement with the findings of Kendall and Pickering (1969). Their treatment of the subject included pressure and gravity terms in the force equations. They find that motion of charged particles parallel to the magnetic field is due to neutral air winds and to plasma diffusion (a function of pressure and gravity). More important to this discussion is that Kendall and Pickering find, to a high degree of approximation, that the perpendicular component of the velocity is:

$$v_L = \frac{\vec{E} \times \vec{B}}{B^2}$$

(3.12)

which is the exact expression we derived following the more general approach of Rishbeth and Garriott.

3.3 Conductivities

Before discussing possible sources of F-region electric fields it is instructive to review the conductivity equations that affect the transport of electric currents in the ionosphere. An ionized medium, in the presence of a magnetic field, has anisotropic conductivities. Therefore, the ionosphere, under the influence of the geomagnetic field, acts as a non-isotropic conductor. The different conductivities, parallel and perpendicular to the magnetic field, play an important role in F-region electric fields.

Ohm's law states that the current density $\vec{J} = \sigma \vec{E}$, where $\sigma$ is the conductivity tensor and $\vec{E}$ the electric field. Making the assumption of approximate charge neutrality (the number density of
ions is equal to the number density of electrons, $N_i = N_e = N$), we can also write:

$$\vec{J} = Ne(\vec{v}_i - \vec{v}_e).$$  \hspace{1cm} (3.13)

Using the velocities solved by equations 3.1 and 3.2, we can derive a tensor expression for the conductivity:

$$\begin{pmatrix}
\sigma_1 & +\sigma_2 & 0 \\
-\sigma_2 & \sigma_1 & 0 \\
0 & 0 & \sigma_0
\end{pmatrix}$$  \hspace{1cm} (3.14)

The different elements in the conductivity tensor represent:

- $\sigma_0$, the direct, longitudinal conductivity, parallel to both $\vec{B}$ and $\vec{E}$,
- $\sigma_1$, the Pederson conductivity, perpendicular to $\vec{B}$, parallel to $\vec{E}$,
- $\sigma_2$, the Hall conductivity, perpendicular to both $\vec{B}$ and $\vec{E}$.

The equations for these conductivities are:

$$\sigma_0 = \left( \frac{N_e}{m_e \nu_{en}} + \frac{N_i}{m_i \nu_{in}} \right) e^2$$  \hspace{1cm} (3.15)

$$\sigma_1 = \left[ \frac{N_e}{m_e \nu_{en}} \left( \frac{\nu_{en}^2}{\nu_{en}^2 + \omega_{ce}^2} \right) + \frac{N_i}{m_i \nu_{in}} \left( \frac{\nu_{in}^2}{\nu_{in}^2 + \omega_{ci}^2} \right) \right] e^2$$  \hspace{1cm} (3.16)

$$\sigma_2 = \left[ -\frac{N_e}{m_e \nu_{en}} \left( \frac{\omega_{ce} \nu_{en}}{\nu_{en}^2 + \omega_{ce}^2} \right) + \frac{N_i}{m_i \nu_{in}} \left( \frac{\omega_{ci} \nu_{in}}{\nu_{in}^2 + \omega_{ci}^2} \right) \right] e^2$$  \hspace{1cm} (3.17)
Using the expressions \( \sigma_{oi} = N_i e^2/m_i v_{in} \) for ions and \( \sigma_{oe} = N_e e^2/m_e v_{en} \) and the approximation \( v_{en} \ll \omega_{ce} \) and \( v_{in} \ll \omega_{ci} \), we can express \( \sigma_1 \) and \( \sigma_2 \) in terms of \( \sigma_{oi} \) and \( \sigma_{oe} \):

\[
\sigma_1 = \sigma_{oe} \frac{v_{en}^2}{\omega_{ce}^2} + \sigma_{oi} \frac{v_{in}^2}{\omega_{ci}^2} \tag{3.18}
\]

\[
\sigma_2 = \sigma_{oe} \left( \frac{v_{en}}{\omega_{ce}} \right) + \sigma_{oe} \left( \frac{v_{in}}{\omega_{ci}} \right) \tag{3.19}
\]

Since the approximation that \( v_{en}/\omega_{ce} \ll v_{in}/\omega_{ci} \ll 1 \) is valid for the F region, we see that \( \sigma_1 \ll \sigma_2 \ll \sigma_0 \). In other words, the conductivity in the direction of the magnetic field lines far exceeds the conductivities across the field lines. This high conductivity, \( \sigma_0 \), in the direction of the magnetic field extends down into the E region and above to the magnetosphere.

The large conductivity along the field lines requires that the electric potential is virtually equal at all points along a given field line. The F region is thus electrically coupled to both the E region below and to the magnetosphere above. This electric coupling plays an important role in the determination of the source of F-region electric fields.

3.4 **F-Region Electric Fields**

Electric fields are produced in the ionosphere in a very simple fashion. Picture an equal number of positive ions and electrons...
embedded in a neutral gas. The neutral gas is moved by winds of different origins across the magnetic field lines. To some degree the electrons and ions share in the motion of the neutral wind. The movement of charged particles across the geomagnetic field is similar to that of a conductor being moved through a magnetic field. A current is produced by such movement which, due to the property of non-isotropic conductivity in the ionosphere, is not free to flow in all directions. Changes will accumulate in the presence of velocity or density gradients, resulting in electrostatic polarization fields.

This process can also be described by the electric field induced by the movement of the neutral wind across the magnetic field lines, \( \mathbf{E} = \mathbf{U} \times \mathbf{B} \). This electric field drives a current \( \mathbf{J} = \sigma(\mathbf{U} \times \mathbf{B}) \), where \( \sigma \) is the conductivity tensor. The current produced by the neutral wind does not have to satisfy the equation \( \nabla \cdot \mathbf{J} = 0 \). At any point where \( \nabla \cdot \mathbf{J} \) is not equal to zero, electric charges will accumulate which build up a gradient in the electric potential. Electrostatic polarization fields result.

As was pointed out by Behnke and Hagfors (1974), F-region electric fields are produced by winds in three different regions of the atmosphere. The first source of F-region electric fields is located in the earth's magnetosphere. Here, the interaction of the solar wind with the geomagnetic field produces electrostatic polarization fields. These fields are mapped down to the F region along the highly conducting magnetic field lines.

The second source of F-region electric fields can be found in the E region of the ionosphere. In this region neutral tidal winds
blow across the geomagnetic field. The resulting polarization fields are mapped up into the F region along the magnetic field lines. This generation mechanism of electric fields in the E region was first suggested by Balfour Stewart in 1882 and is known as the dynamo theory.

The final source of F-region electric fields is located in the F region itself. Thermospheric neutral winds in the F region blow across the magnetic field lines, giving rise to a very small differential drift of ions and electrons. Although small, this \( \frac{\mathbf{v} \times \mathbf{B}}{eB^2} \) drift can cause electric charges to accumulate in the presence of velocity or density gradients. The charges then set up electrostatic polarization fields which can move the plasma across the field lines.

Behnke and Hagfors (1974) point out that at Arecibo electric fields produced in the magnetosphere are unlikely to penetrate down to the F region since Arecibo is so well inside the plasmasphere (L = 1.4). This leaves us with two possible sources for F-region electric fields: neutral winds in either the E region or the F region.

Rishbeth (1971) points out that, in the daytime, electrostatic fields produced in the F region are likely to be immediately short-circuited since the F region is connected to the highly conducting E region via the magnetic field lines. At night, the E region conductivity is greatly reduced which allows electric fields generated in the F region to exist. The results of Behnke and Hagfors (1973) suggest that F-region electric fields are produced by tidal winds in the E region during the daytime and by thermospheric winds in the F region at night.
Our discussion of F-region dynamics can be summarized by the following. Electrons and ions move across the magnetic field lines in the F region only when there exists an electric field at right angles to \( \mathbf{B} \). The electric fields responsible for this perpendicular motion, at Arecibo, may be produced in either the E region or the F region. The high conductivity along the geomagnetic field lines allows E-region electric fields to be mapped up into the F region.
CHAPTER 4
THE GUADELOUPE DATA

During the vast majority of the experiments conducted in June, 1977, the AFAS velocity spectra for each range were clean, Gaussian-shaped curves from which a single velocity component could be selected. Moreover, the velocities thus determined were fairly constant in space over all detected ranges and in time over intervals of 5 to 10 minutes. Therefore, for most of the data, we can determine an average velocity of the striations over these time intervals and throughout the heated region. On two days, however, 5 June 1977 and 9 June 1977, during the early evening hours at Arecibo between 19:00 and 21:00 Atlantic Standard Time (AST), fairly dramatic reversals in the striation velocities were observed over time scales on the order of minutes.

In this chapter we will present our interpretation of the AFAS velocity data of the Guadeloupe experiments. We will examine the relationship between the power output of the HF transmitter and the striation velocities. We will also discuss the general patterns of natural F-region drifts and compare them to AFAS velocities. In addition, we will present Duncan's correlation between the velocities of large and short-scale striations.

Also in this chapter we will examine the atypical velocity data and relate the sharp changes detected in the AFAS velocities to
other natural phenomena in the ionosphere. A mechanism that could induce the AFAS velocity disturbances is outlined. For further insight, observations made by Ecklund, Carter and Balsley of similar velocity disturbances detected in E-region irregularities will also be discussed.

4.1 Relationship of the AFAS Velocities to HF Transmitter Power

On 6 June 1977, between 17:50 and 19:30 AST, both the power and the frequency of the Arecibo HF transmitter were changed over time intervals of 5 to 10 minutes to test the effect on the striations and their velocities. The short time periods were chosen in hopes that the background ionosphere would remain fairly constant. Changing the frequency of the HF transmitter adjusts the altitude at which the striations form, while altering the power output of the transmitter affects the amount of energy deposited in the heated region. The 50 MHz radar on Guadeloupe was used to detect the striations over Arecibo and their velocities normal to the earth's field lines.

On this day the striations appeared to move uniformly across all detected ranges so that an estimate of an average velocity for the AFAS can be determined. The striation velocities observed during this time period were all positive. Throughout this thesis positive velocities will refer to motion towards Guadeloupe, while negative velocities will refer to motion away from Guadeloupe. Figure 4.1 is a plot of the AFAS velocities averaged over one-minute time intervals and over all detected ranges versus the power and the frequency of the HF transmitter.
Figure 4.1 VELOCITY VERSUS HF POWER AND HF FREQUENCY
The data presented in Figure 4.1 shows that AFAS velocities appear to have no direct relationship to either the frequency or the power of the HF wave. The velocity fluctuations that are detected can easily be explained by normal variations in background electric fields. While examining this data, it is instructive to note that no scattering was detected from the striations when the HF power output was 5 kW and only weak scattering was detected when the power output was 11 kW. Frey (1980) has interpreted this to represent a 10 kW threshold power to produce the AFAS in the F region above Arecibo.

4.2 Comparison of AFAS Velocities to Natural F-Region Ionization Drifts

The data of Figure 4.2 is an estimate of the average AFAS velocities plotted as a function of time of day. The striation velocities were averaged in time over 10-minute intervals and in space over all detected ranges for the entire period of the experiments, 5-15 June 1977. Since the Guadeloupe radar detects the motion of the striations across the magnetic field in a predominantly east-west direction, we may refer to the positive velocities as eastward drifts and the negative velocities as westward drifts.

Although F-region ion drifts can be measured using the 430 incoherent scatter radar at Arecibo, none were measured during these experiments. Therefore, the comparisons we make to the natural F-region ionization drifts must be limited to that of magnitude and general direction. The Guadeloupe radar could measure AFAS velocities from -75 m/s to +75 m/s. These velocities are of the correct order of
Figure 4.2  Average AFAS Velocities Versus Time of Day. June 5-15, 1977
magnitude for F-region ionization drifts perpendicular to the magnetic field.

With the exception of two days, the AFAS velocities measured between 16:00 and 21:45 AST were positive or in the eastward direction. The days for which the velocities were negative during this time period were the atypical days that will be discussed later in this chapter. This pattern of eastward drifts before midnight agrees nicely with the findings of Behnke and Hagfors (1974). The Behnke-Hagfors data shows that in general the east-west component of the ion drift velocity perpendicular to the magnetic field is eastward before midnight and westward after midnight. Recall that, in the F region, electrons and ions drift across the magnetic field lines at approximately the same rate so that a comparison of the ion drift velocity to the AFAS velocities is valid.

Unfortunately, due to the difficulty in scheduling the time for use of the HF transmitter at Arecibo, we lack a significant amount of data between the hours of 10:00 PM to 2:00 PM (22:00 to 14:00 AST). The data we do have do not show westward drifts until 9:00 AM. As a result, our findings are not in complete agreement with those of Behnke and Hagfors. However, the Guadeloupe radar also detects a small fraction of the AFAS motion in the north-south direction perpendicular to the magnetic field. Behnke and Hagfors have measured this component of the ion drift velocity to be positive, or southward, after midnight. This might explain the discrepancy between our data and that of Behnke and Hagfors.
4.3 The Velocities of Large and Short Scale Striations

An interesting sideline to this discussion of AFAS velocities is Duncan's correlation (personal communication) between the velocities of short and large scale striations presented in Figure 4.3. Large scale striations are on the order of kilometers wide and are thought to be caused by the HF wave becoming collimated or striated as it propagates through the ionosphere.

The mechanism of thermal self-focusing is thought to produce the large-scale striations. This mechanism can be described as follows. Natural density fluctuations in the ionosphere cause small variations in the index of refraction. As the HF wave propagates through these fluctuations, a larger percent of its energy is refracted into the regions of smaller electron density, thus focusing the HF wave. Because of ohmic heating and the ponderomotive force, more plasma is pushed out of the focused regions, thereby creating a new perturbation downstream from the first fluctuation. The width of these large scale striations is determined by both ionospheric conditions and the power density of the HF wave.

The velocities of the large scale striations detected in these experiments were measured by Duncan (1978). He used the technique of scanning the 430 MHz radar at Arecibo through a set of striations, then quickly reversing the beam direction and sweeping back through the same striations. Using this method, Duncan was able to estimate both the size and the velocities of large scale striations.
Comparison of Large and Small Scale Striation Velocities

- Dotted line: Large Scale (500m) Irregularities
- Solid line: Small Scale (3m) Irregularities

Figure 4.3

June 6, 1977
Short scale striations are thought to form inside of the large scale striations. It is assumed that the correlation between the velocities of large and short scale striations shows that both sets of striations drift at the natural speed of ionization in the F region. The small scale striations would be packaged within the large scale striations.

4.4 Atypical Data

We will now focus our attention on the data for which unusual disturbances in the velocity spectra of the striations were noticed. On 9 June 1977, between 19:00 and 19:30 AST, and on 5 June 1977, between 19:20 and 19:50 AST, disturbed patterns in the velocity spectra were detected. Figure 4.4 shows the abrupt shift in the peak of the velocity spectra for one range from Guadeloupe on 9 June. The peak shifts dramatically from +20 m/s towards Guadeloupe to -70 m/s away from Guadeloupe in a period of 2 minutes.

A closer inspection reveals that the velocity spectra for these time periods have more detail than those of ordinary spectra. Figures 4.5 through 4.7 are plots of the signal-to-noise intensity of the velocity spectra for one range versus time on three different days. The velocity spectra of 7 June, a normal or undisturbed day, for the range of 690-697 km from Guadeloupe, are plotted in Figure 4.5 over an approximate 30-minute time interval. During this time period the Arecibo transmitter was operating at 50 kw of power and at a frequency of 7.8 MHz, and switching between ordinary (0) and extraordinary (X) mode polarization at 2-minute intervals. The striations did not form
Range: 679-687 Km
Date: June 9, 1977
Time: 19:00-19:19 AST
Time Intervals 12 secs.

\[ \Delta t = 2 \text{ minutes} \]

Figure 4.4 PLOT OF VELOCITY SPECTRA PEAK VERSUS TIME
Range: 690-698 Km  
Time Resolution: 6.4 secs  
June 7, 1977  
Start Time: 19:03:00 AST

Figure 4.5 VELOCITY PLOT FOR TYPICAL DAY

Stop Time: 19:28:08 AST
when the heater was transmitting X-mode polarization which accounts for the regular intervals between the data points. Nevertheless, the velocity spectra appear smooth with little variation between one time period and the next.

Our next plot, Figure 4.6, shows the velocity spectra of 9 June for the range of 675-682.5 km from Guadeloupe in the time interval of 19:00-19:26 AST. The HF transmitter at Arecibo was on continuously until 19:25 AST, operating at a frequency of 6.78 MHz and a power level of 60 kw. These velocity spectra do not have a smooth continuous appearance. At approximately 19:05 AST, there is a dramatic switch from a single positive velocity component to three peaks in the velocity spectrum, two of which are negative. At a later time, 19:20 AST, there appears to be another shift towards higher negative velocities. This second shift in velocity is smoother and more continuous than the first. Recall that during this time period, the early evening hours before midnight, F-region ionization drifts are typically positive or eastward.

The symmetric nature of the peaks in the spectra can be interpreted by two different physical pictures. Because of the aspect sensitivity of the AFAS, a circular motion would be detected as two components of motion. In other words, the radar would detect only the motion towards and away from Guadeloupe. The other possibility is that of two counterstreaming layers of striations. Note that the middle peak occurs at a later point in time, and evolves into the dominant peak. It is possible that this velocity component (\( -20 \) m/s)
Range: 675-683 Km
Time Resolution: 6.4 secs
June 9, 1977
Start Time: 19:00:00 AST
Stop Time: 19:26:48 AST
is the result of collisional dissipation between two counterstreaming layers.

Figure 4.7 shows the velocity spectra for 5 June between 19:06 and 19:32 AST for the range 675-682.5 km from Guadeloupe. During this time period, the HF heater was operated at a frequency of 7.8 MHz and a 50 kw power output. The transmitter was being pulsed on and off at 2-minute intervals, with an exception at 19:15 AST when it was pulsed once at a half a minute. In this data a velocity disturbance can be detected at approximately 19:15 AST that continues until 19:22 AST. It is unfortunate that the heater was off during most of this time period.

An investigation of the 5 June and 9 June data was conducted to determine possible causes for the disturbed velocities. No unusual activity was noticed in the geomagnetic field on these days. Magnetograms from the Baker Lake, Great Whale River and Churchill Observatories (all located in Ottawa, Canada) were quiet throughout the experiment. In addition, the Kp number (or planetary K index) which measures the degree of disturbance in the magnetic field was consistently low during these time periods.

The investigation revealed two characteristics that the 5 June and 9 June data shared in common. The first of these characteristics is related to the time of day that the unusual velocity spectra were detected. Both sets of data were taken in the early evening hours after sunset. The experiments were run on a total of six evenings during the 5-15 June period.
Range: 675-683 Km
Time Resolution: 6.4 secs
June 5, 1977
Start Time: 19:06:09 AST

Stop Time: 19:31:48 AST

Figure 4.7 VELOCITY PLOT - DISTURBED DAY
Throughout the experiments, in the early evening hours between 18:00 and 21:00 AST, disturbed intensity signals from the heated region were received by the Guadeloupe radar. The velocity data taken on 5 June and on 9 June both occurred at times when the return signal appeared irregular. Figure 4.8 is a plot of the intensity of the return signal detected at Guadeloupe on 9 June as a function of time and range. The entire plot represents approximately 2 seconds of data. If we examine this plot, we see that the intensity of the scattered signal appears to oscillate back and forth between different ranges in the heated volume. In particular, the return intensity sometimes appears strongest from the range of 712-720 km from Guadeloupe, while at other times the intensity is strongest in the range of 682-690 km. These ranges are separated by about 25 km, which results in a horizontal separation of about 24 km and a vertical separation around 7 km.

If data such as those presented in Figure 4.8 are averaged over 1-minute intervals, double and triple peaked intensity profiles result. When examining the bulk of the Guadeloupe data, multiple peaked intensity profiles resulted every evening the experiments were run. Multiple-peaked intensity profiles also occurred twice for short periods of time at 14:00 AST and once at 08:00 AST. The odd-shaped profiles during the morning and evening hours are probably related to the changing ionospheric conditions at sunset and sunrise.

The relationship of these disturbed intensity profiles to the unusual velocity spectra is not clear. Multiple-peaked intensity profiles do not necessarily imply disturbed velocity spectra. However,
Intensity Plot of 16 Ranges: 600-720 Km
Each Range is 7.5 Km
Time Resolution: 10 msec
June 9, 1977
Start Time: 19:22:48 AST

Stop Time: 19:20:50 AST
since the intensity of the scattered power signal appears to oscillate between different regions in the heated volume, we can perhaps postulate two distinct layers of ionization that could move at different velocities. This could explain the data of 9 June presented in Figure 4.9. This plot shows a shift in velocities on the order of \(-70\) m/s detected in the range 675-682.5 km from Guadeloupe to velocities on the order of \(-30\) m/s detected in the range of 690-697.5 km.

The other distinguishing feature that the data taken on 5 June and 9 June had in common was the strong sporadic E detected by the Los Caños ionosonde. Sporadic E refers to dense layers or patches of ionization in the E region that possess horizontal dimensions between 100 and 1000 km. These ionospheric irregularities occur frequently in the evening hours at Arecibo during the summer months. The sporadic-E patches drift at a rate on the order of 50 m/s (Rishbeth and Garriott, 1969). According to Djuth (1979), a significant fraction of the ions present in the sporadic-E layers above Arecibo are metallic, the most common of which are Fe\(^+\), Mg\(^+\) and Si\(^+\). Metallic ions in the ionosphere are thought to be meteoritic in origin. At the top and bottom of the sporadic-E layers are large vertical density gradients. These electron density gradients cause sporadic-E to act as a partial, or even total, reflecting layer for high-frequency emissions.

The Los Caños ionograms for the time periods of the disturbed velocity data on 5 June and 9 June show partial reflection of the HF waves off sporadic-E layers located at approximately 100 km above the Los Caños dish. Figure 4.10 presents three ionograms taken on 9 June 1977 at 19:00, 19:01 and 19:05 AST. The broad line at the bottom of
AFAS VELOCITIES AS A FUNCTION OF TIME AND RANGE

Start Time: 19:00:00 AST
Time Resolution: 6.4 secs
June 9, 1977

Range: 675-683 Km  Range: 682.5-690.0 Km  Range: 690.0-697.5

Stop Time: 19:09 AST

Figure 4.9  AFAS VELOCITIES AS A FUNCTION OF TIME AND RANGE
Figure 4.10  IONOGRAMS FOR JUNE 9, 1977
each ionogram represents reflection of the probing waves off the sporadic-E layer. The highest frequency reflected by this layer increased from 4 MHz at 19:01 to 9 MHz at 19:25 AST. Note that the first disturbance detected in the velocity spectra of 9 June (see Figure 4.6) occurred at 19:05 AST.

Figure 4.11 shows three ionograms taken on 5 June 1977 at 19:10, 19:15 and 19:20 AST. Here the highest frequency reflected by the sporadic-E layer increased from approximately 4 MHz at 19:10 AST to 9 MHz at 19:20 AST. Again note that the disturbance detected in the velocity spectra of 5 June (see Figure 4.8) occurred at approximately 19:15 AST. Sporadic E was detected on numerous occasions throughout the experiment. However, only on 5 June and 9 June were reflections from frequencies higher than 5 MHz detected in the Los Caños ionograms.

A mechanism by which the sporadic E could affect the velocities of the AFAS in the F region will now be discussed. Electric fields are generated in the ionosphere by the motion of conducting layers across the geomagnetic field lines. During the daytime, the E region acts like a conducting layer, about 30 km thick, that is moved by tidal winds across the field lines. The electrostatic fields that result are mapped up into the F region along the magnetic field lines. Shortly after sunset, however, the E region decays.

During this time period electrostatic fields generated in the F region become the dominant source for F region electric fields. The presence of sporadic E, a highly conducting layer in the E region, could change this situation. Electrostatic fields produced by the
Figure 4.11 IONOGRAMS FOR JUNE 5, 1977
movement of the sporadic-E layer across the field lines could be mapped up into the F region along the field lines. This introduction of a different electric field in the local F region would affect the motion of striations across the field lines. Thus the sporadic E presents a plausible explanation for the disturbed velocity spectra seen on 5 June and 9 June 1977.

It should be pointed out that the magnetic field lines in the E region that map up into the F region above Arecibo are located approximately 100 km away from the Los Caños ionosonde. Nevertheless, this should not present a problem to the above mechanism since sporadic-E layers are on the order of 100-1000 km in the horizontal dimension.

Certain characteristics of the second velocity disturbance detected in the velocity spectra of 9 June 1977 (see Figure 4.6) can be explained by another ionospheric phenomenon. This second velocity disturbance has a smooth, wave-like structure similar to that seen in a small-scale Travelling Ionospheric Disturbance (TID). According to Gossard and Hooke (1975), small scale TID's occur frequently at temperate latitudes. The horizontal wavelengths for small scale TID's are on the order of tens or hundreds of kilometers. The corresponding wave periods range from ten to fifteen minutes up to several hours.

The data presented in Figure 4.6 fits nicely with this picture of small scale TID's. A smooth, wave-like structure in the velocity spectra for 9 June was detected in four ranges, 668 km to 705 km from Guadeloupe. From this we can say that the wave appeared to have spatial coherence over the entire heated volume. The data appears to
exhibit a wave-like structure for half of an eleven-minute wave period, which is similar to a time length for a wave period of a small-scale TID. Nevertheless, a TID cannot explain the first discontinuity in the velocity spectra of 9 June, which does not appear to have a wave-like form.

To conclude this chapter, we will present the observations of E-region irregularities made by Ecklund, Carter and Balsley (personal communication). Their observations may shed more light on the role of sporadic E.

4.5 E-Region Observations

During the June 1977 experiments, echoes were occasionally observed by the 50 MHz radar over a narrow band of ranges between 210 and 270 km from Guadeloupe. These ranges are approximately one-third of the distance between Guadeloupe and the F region above Arecibo and correspond to E-region heights. At these ranges, the 50 MHz radar beam is perpendicular to the geomagnetic field at the altitudes of 102 to 108 km. The Guadeloupe radar is capable of detecting signals from E-region altitudes over a much wider band of ranges, although the beam is not perpendicular to the geomagnetic field. Since the echoes are confined to the ranges where the beam is perpendicular to the magnetic field, it is concluded that the echoes are caused by naturally occurring 3m irregularities that are aligned along the magnetic field lines in the E region.

Backscattered signals from the E region were observed by Ecklund, Carter and Balsley on five occasions. All of the observations
occurred during the early evening hours between 18:30 and 22:00 AST. The signals were detected over time intervals from 5 to 15 minutes. Sporadic-E layers are known to form frequently in the evening hours in June at approximately 100 km in altitude above Puerto Rico. Ecklund et al. (1981) suggest that the steep electron gradient regions associated with sporadic-E layers might enhance the generation of 3m field-aligned irregularities. The 5 to 15-minute duration of these echoes may then be explained as the drifting of a sporadic-E patch through the Guadeloupe radar beam.

The E-region observations of most interest to us are the disturbances detected in the velocities of these 3m irregularities. On two occasions, Ecklund et al. observed sudden reversals in the velocities of the E-region irregularities. On 6 June 1977, from 20:05 to 20:10 AST, the detected E-region drift velocity changed from -75 m/s to +115 m/s. On 7 June 1977, from 19:12 to 19:13 AST, the E-region drift velocity switched from +70 m/s to -100 m/s. During both of these time periods the corresponding F-region velocities were smooth and regular. Nevertheless, the E-region velocity changes are similar in magnitude to our F-region velocity disturbances. Indeed, there may be a relationship between the sudden reversals in the drift velocities of both regions.

Ecklund et al. observed backscattered signals from the E region on only one of the two days in which F-region velocity disturbances were detected. No E-region scattering was observed on 9 June 1977. On 5 June 1977, from 19:15 to 19:30 AST, a fairly smooth pattern of
E-region drifts was detected by the 50 MHz radar. These drifts had speeds ranging from +50 m/s to +75 m/s. The F-region velocity disturbance occurs at approximately 19:15 AST.

The lack of direct correlation between the E- and F-region velocity disturbances is to be expected from the geometry of these observations. The section of the E region that is electrically coupled to the F region above Arecibo is about 500 km away from where the E-region drifts were measured. Only if the sporadic-E layer had horizontal dimensions greater than 500 km would we expect the E- and F-region drifts to be correlated. Since sporadic-E layers vary between 100 and 1000 km in horizontal extent, it is not surprising that the Ecklund data occasionally suggest reasonable agreement between the undisturbed E- and F-region drifts.

It is quite possible that the E-region velocity reversals are caused by the same phenomenon that induces the sharp changes in the F-region velocities. Ecklund et al. (1981) suggest that, during the early evening hours, the sporadic-E layer may cause the intermittent coupling and decoupling of the E and F regions. This is basically the same mechanism that we suggested for the disturbed F-region velocities. If the E and F regions are alternately coupled and decoupled, the velocity disturbances in both ionospheric regions can easily be explained.
In concluding this thesis, we may summarize our results as follows:

1) Once the power of the HF wave is above the threshold at which the striations are produced (approximately 10 kw), no apparent relation between the HF power and AFAS velocities has been detected.

2) The magnitude and direction of the AFAS velocities are similar to those of general F-region ionization drifts.

3) There is a good correlation between the velocities of large and short scale striations.

4) Velocity disturbances detected in both E- and F-region irregularities suggest that there is a relationship between sporadic-E layers and E- and F-region coupling.

The first three of these results were expected. AFAS velocities had been assumed to be related to the background electric fields in the ionosphere. We would like to confirm this result by an additional test which would require repeating the experiment. The test would be to monitor the background ion drifts in the F region with the 430 MHz radar at Arecibo. The 50 MHz radar on Guadeloupe would be used to observe the AFAS velocities in the heated volume of the
F region. A direct comparison could then be made between the short scale striation velocities and the natural ionization drifts in the F region.

The last of our results, which suggested that sporadic-E layers play an important role in E- and F-region coupling, was unexpected. Clearly, no definitive claim can be made until more experimental work has been done. Another set of experiments is planned to study the formation of short scale striations in the heated E region. A new heating facility at Arecibo capable of transmitting 800 kw of power in the frequency ranges of 3 to 12 MHz will soon be completed. The 50 MHz radar is situated on the island of St. Croix. This geometric location allows the 50 MHz radar to have a perpendicular line-of-sight to the magnetic field lines at approximately 105 km above Arecibo. The radar will thus be able to detect the AFAS that form in the vicinity of this E-region altitude.

One experiment that could be done would require the presence of sporadic E detected by the radar on St. Croix. In this case the 50 MHz radar would observe the velocities of the naturally occurring 3m irregularities in the E region. The F region could then be heated and the large scale striation velocities monitored by the 430 MHz radar at Arecibo. The E- and F-region velocities could then be compared as they were during the Guadeloupe experiments. Unfortunately, large scale striation velocities take minutes to determine and only average F-region velocities could be obtained using this method.
Ecklund, Carter and Balsley have suggested that simultaneous E- and F-region drifts on a common field line could be measured in these experiments. The 50 MHz radar on St. Croix could be pointed to the north to monitor the E-region drifts. The 430 MHz radar at Arecibo could be pointed west to the region where the field line observed by the St. Croix radar intersects with the F region. The 430 MHz radar would measure an F-region ion drift component. Observing the E- and F-region drifts on a common field line would give us direct information on the nature of E- and F-region coupling.

The results of the Guadeloupe experiments clearly indicate the need for more experimental work. The question of E- and F-region coupling in the presence of sporadic E remains unanswered. It is hoped that observations in the upcoming experiments will provide additional information to this and other unresolved questions concerning short scale striations.
REFERENCES


