Studies of the Tilts of Atmospheric Scatterers by Windprofiler Radars

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Abstract

MST (Mesosphere–Stratosphere–Troposphere) radars are powerful instruments for observing the differences in refractive indexes in the air. By tracking refractive index perturbations, these radars can determine wind speeds and turbulence strengths in the atmosphere from the ground up to 15 km altitude. In this project, we used 6 Doppler Radars to study the correlation between horizontal and vertical winds in the Troposphere to find scatterer tilts. However, instrumental effects due to small tilts in the radar beam can also cause such correlations, thus our studies cover that as well. Scatterers are best known for being the result of turbulence and waves in the Troposphere. Hence understanding the nature of scatterers leads to a better understanding of the gravity wave’s effects in the Troposphere, as they can generate this turbulence.

This thesis found the correlation between horizontal and vertical wind and discovers it changes due to season, month, height, and location. Therefore, it is concluded that this correlation results from the tilt of scatterers, and if there is a tilt in the vertical beam, it is minimal.

Keywords: MST Radars, Doppler Radars, Troposphere, Gravity Waves, Turbulence, Scatterer, Wind, Horizontal, Vertical, Correlation
Summary for Lay Audience

Many geophysical features like sea-shores, lakes-shores, and mountains can generate gravity waves, which happen when the airflow fluctuates by hitting obstacles and is forced to move upward and downward. These gravity waves can be sources of turbulence in the Troposphere. This project aims to understand geophysical and seasonal effects on gravity waves by studying the scatterers induced by turbulence. For this goal, the correlation between horizontal and vertical wind, by using Doppler Radars, is investigated. Studying this correlation gives us valuable information about the scatterers’ tilt. This tilt helps us understand the turbulence that they originate from while providing better information about our radars’ accuracy. The ability to model and predict turbulence is one of the essential skills for weather forecasting.
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Chapter 1

Turbulence and Scatterers

1.1 Introduction

Understanding tropospheric scatterers is a crucial subject for understanding atmospheric phenomena like turbulence, gravity waves and etc. This research tries to understand the shape and tilt of scatterers better as it gives us valuable information about their origin. The shape of scatterers is mostly visualized as isotropic, but in this research, we found they are closer to anisotropic. For this goal, we used MST (Mesosphere- Stratosphere- Troposphere) Doppler radars to analyze the backscattered signal’s data. Also, we concluded the scatterers are tilted in the troposphere. These conclusions led to the final and more significant conclusion that these scatterers are the result of gravity waves, and we provided evidence for the existence of gravity waves in the troposphere. Another objective of this research was to figure out if our radar’s vertical beam is tilted or not. Our results will show that if there is a tilt, it is very small.

Chapter 1 is the description of the problem and issues that had to be addressed for researching the problem. Chapter 2 discusses the fundamentals of radars and the data analyzing method. Chapter 3 talks about gravity waves and how we can conduct an experiment on them. Chapter 4 describes our instruments and their unique features. Chapter 5 is a thorough explanation of our method, and chapter 6 is the result chapter. Finally, chapter 7 is dedicated to the
conclusions of this research. Also, some examples of codes and permissions are displayed in chapter 8 under the name of Appendix.

1.2 The Atmospheric Doppler Spectrum

Now we try to dive more into the problem and explain how we approach it. As turbulence makes scatterers in the atmosphere, we aim to study these scatterers so that later they can give us a better estimation of turbulence and use them in meteorology. In general, turbulence can affect the radial velocity of wind, as the isotropic turbulence produces spatially correlated fluctuations in the wind field as same as the anisotropic turbulence. However the later fluctuations are more directed to the vertical direction for anisotropic ones.

In this project we used windprofiler radars to measure vertical and horizontal velocity of wind as these velocities are interpreted from the backscattered signal from the scatterers. This is being discussed more in the next chapters. For this mean, the vertical and horizontal velocity was measured, and the correlation between them was investigated. The next step is understanding the nature of this correlation and the cause of it. There are multiple explanations for this correlation which, we are going to address. First of all, as we used wind profiler radars we should be aware of their limits. For example if the radar is located close to cities or humid places, it can affect our data. This effect raises the importance of signal to noise to ratio and finding the best means to reduce the noise and external impacts on the spectrum, which was discussed thoroughly in the spectral fitting section. [1]

Another instrumental issue that might affect this correlation the structure of the radar. If the vertical radar beam is not properly vertical, the measured vertical wind can contain some horizontal wind portion. A measured mean wind of 30 \( ms^{-1} \) by the radar with the beam tilt about 0.1° of the nominally vertical beam can produce the apparent vertical wind of 5 \( cms^{-1} \), which can be significant when averaging over the time scale of few days or more. Averaging over the time scale of few day is done to get statistical reliability since short-term averages
have too much scatter. Also, We are most interested in time scales of many hours and days, since those are the weather time scales we are interested in for weather forecasting, so we want to know if these data are contaminated.

The other cause of correlation can be the tilt of the of tropospheric layers around the mountain regions which produces tilted anisotropic scatterers, like anisotropic turbulence and specular reflectors [2]. Similar to the mountain regions, lakeshores and beaches can cause the same effect [3]. They, too, can generate tilted anisotropic scatterers. Therefore, modeling the scatterers is a more complicated process than it was assumed in early literature, before Rotger in 1981 [4], and it needs more care as if the scatterers are isotropic, the correlation can’t be seen. [5]

The general shape of the scatterers defines the way that radio signals are reflected or scattered back. If a reflector is like a mirror, then the signal reflects in the same way that a torch shone perpendicular to a mirror will reflect back into your eyes. However, if the scatterers are like little semi-transparent spheres of the size of wavelength or less, the signal will be scattered in all directions. If the scatterer is a semi-transparent ellipse (with big diameter of $2 - 3 \, m$ and the small diameter about $1 \, m$) the scattered signal will be in between these extremes [6]. So we define the directional dependence of the scattered signal by an effective "polar diagram", which can be written as

$$B(\theta) = e^{-\frac{\sin^2 \theta}{\sin^2 \theta_s}}$$  \hspace{1cm} (1.1)

Here $B(\theta)$ is the power of backscattered signal that is detected with the radar and $\theta_s$ shows the contribution of the nature of the scatterer where $\theta_s = 90^\circ$ shows an isotropic scatterer and $\theta_s = 0^\circ$ represents highly aspect sensitive scatterer; meaning it backscatters the signal differently from different angles of it, thus the receiver receives a non zero signal unlike the isotropic scatterers. The received signal from an isotropic scatterer is zero because the average of backscattered signals is zero as they are homogeneously backscattered [7]. As it has been shown, $\theta_s$ can have variant values that represent variation in the shapes of scatterers. Consid-

Chapter 1. Turbulence and Scatterers

The true model of the scatterers is one of the controversial topics in the literature, as they have been modeled as flat, pancake like or other shapes, therefore this thesis allows us to have a better understanding of the shape of scatterers. It should be noted that each scatterer is highly distorted but we work with the average shape of them. [10]

1.3 Radial Velocity

If it is assumed that the data are collected over time of \( T \), and the radial velocity of the scatterers are constant over time we can write:

\[
y(t) = \lim_{N_s \to \infty} \sum_{l=1}^{N_s} \tilde{A}_l e^{-i2k_0r_{0l}} e^{-i2k_0\nu_{rl}t}
\]

Here \( r_{0l} \) is the distance of the scatterer at time zero, and \( \nu_{rl} \) is the radial velocity of the \( l \)th scatterer. Also, it is assumed scatterers closer to each other have similar radial velocity, echo power and range because of natural continuity of the air in the atmosphere. Now for having the time series in the format of frequencies, we should convert \( y(t) \) with Fourier Transform as:

\[
\tilde{Y}(\omega) = F[y(t)] = 2\pi \lim_{N_s \to \infty} \sum_{l=1}^{N_s} \tilde{A}_l \delta(\omega - \omega_{dl}) e^{-i2k_0r_{0l}}
\]

Here \( \omega_{dl} \) shows the Doppler frequency for the scatter \( l \). Now the power spectrum is calculated with the term of: \( |\tilde{Y}(\omega)|^2 \) and the result function is called Doppler spectrum, which is the power-weighted distribution of Doppler frequencies within the resolution volume of the radar. As the Doppler spectrum contains the radial velocity, the mean of the distribution gives information about the average flow within the volume and the spread of frequency distribution.
1.3. **Radial Velocity**

contains information about the variability of the velocities, which was addressed more in the last section. For reporting of the mean wind flow this notation is used:

\[
\mathbf{u} = [u, v, w]
\]  

(1.4)

Conventionally \( +u \) is the zonal wind and is directed from west to east (Westerly or Eastward). \( +v \) is the meridional wind from south to north (Southerly or Northward), and finally, \( +w \) shows the vertical element of the wind, which is directed upward (Based on glossary.ametsoc.org/wiki). Also, we need to define the zenith angle as \( \theta \) and azimuth angle as \( \phi \).

Using the above notation, the wind velocity is defined as:

\[
\vec{u} = \vec{u} \hat{i} + \vec{v} \hat{j} + \vec{w} \hat{k}
\]

(1.5)

It is being said that \( \vec{u} \) is used to show the total wind vector as a convention and also the zonal component of the wind \([u] \) is not the magnitude of \( \vec{u} \). Now the radial velocity can be written as

\[
v_r = \vec{u} \cdot \left( \frac{x \hat{i} + y \hat{j} + z \hat{k}}{\sqrt{x^2 + y^2 + z^2}} \right) = \vec{u}.[\sin \theta \sin \phi \hat{i} + \sin \theta \cos \phi \hat{j} + \cos \theta \hat{k}]
\]  

(1.6)

where the \( \left( \frac{x \hat{i} + y \hat{j} + z \hat{k}}{\sqrt{x^2 + y^2 + z^2}} \right) \) is showing the unit vector of the direction of beam and the symbol "." represents dot product. By performing the dot product, the radial velocity is written as

\[v_r = u \sin \theta \sin \phi + v \sin \theta \cos \phi + w \cos \theta.\]  

It should be considered that \( \phi \) is the angle clockwise from east in atmospheric work.

Now the correlation between the vertical wind (\( \vec{w} \)) and horizontal wind (which was averaged over the spectrum) \( (U = \vec{u} + \vec{v}) \) is calculated. We look for the combination of \( (\theta, \phi) \) where the correlation maximizes. If we find a significant non-zero correlation, it means the measured vertical velocity had a horizontal component inside it as producing the \( \sin \phi \) or \( \cos \phi \) therefore; there is definitely a \( \theta \) which as mentioned above, shows either there is a tilt in the radar beam or in the scatterer and the nature of this tilt should be explored as it is affected by
many factors like the shape of the scatterers or their scale to the scale of the radar beam or etc. If either the tilt of the radar’s beam or the tilt of layers was zero, then measured vertical velocity should only give us $w$ and therefore there shouldn’t be any components of the $\sin \theta$ and $\cos \theta$.

1.4 Our Approach

In this project, this correlation is studied based on changes of the $\phi$, heights, seasons, and locations to have a better understanding of its nature with giving us a clue about the shape of the scatterers and tilt of the beam of the radar being used. In the next section, we explain our method and how the $\theta$ was investigated.
Chapter 2

Radar’s Principles

Radars are the essential instrument in atmospheric work after world war 2. In this chapter, the principle of MST radars, their design, and how they function, is explained and in the last chapter, the spectral fitting method, which was used for converting data to usable data, is described.

The name of radar comes from "radio detection and ranging". We use atmospheric radars to transmit electromagnetic waves ($E - M$ wave), then observe the received electromagnetic wave after they hit the target and are returned either reflected or scattered. However, we mostly face scattering rather than reflecting in atmospheric work. We can find these scattering and reflection strengths by examining the refractive index changes because of the $E - M$ wave interaction with the phenomena.

There are various types of targets, but in atmospheric observations, we mostly can see aircraft and missiles (which are of little interest to us) and one of the most important ones, changes in humidity and water droplets embedded in the air. At our frequencies (40 – 50 MHz), changes in refractive index are much more critical, which helps us distinguish the water vapour from the rest of the air. The other exciting phenomenon that changes the refractive index of the atmosphere is turbulence. Turbulence is a small-scale, irregular air motions in the atmosphere characterized by winds that vary in speed and direction [11]. It can be defined as perturbation that changes the refractive index of the atmosphere in its area and can backscatter.
the radio wave signal; therefore, it is detectable with radars and it is more discussed in the next chapters. As we can see here in the Figure 2.1 that shows the backscattered power for $SOUSY$-VHF-radar (SOUnding-SYstem) in Germany; an immense received power between $6 - 9\ km$ (Height), is expected due to the wind shear generated turbulence. [12]

All of these scatterers have different behaviours through time and height. Their refractive index’s value can be dependent on the pressure, humidity, temperature, and other factors with unique features, which can help us choose the best method for observation, data acquisition and data analysis in this case. [13]

### 2.1 Simple Radars

Each radar consists of (1)- a radar controller, (2) a transmitter, (3) a transmitter antenna, (4) a receiver antenna, and (5) a receiver. An $E - M$ wave is produced by the transmitter and propagates through the transmitter antenna and into the air. As we can see in Figure 2.2, the signal moves forward, then reaches the target in the fourth step. Most of the pulse continues
Figure 2.2: Schematic picture of the position of a transmitted pulse as a function of time, as it moves away from the radar. In the upper figure, a pulse is transmitted; in the middle one, the pulse encounters the target; in the fourth one, some portion of the pulse is transmitted, and some are reflected. In the last one, the receiver is detecting the returned pulse. Permission’s licence of this figure is available in the Appendix section. [15]

its path, while just a small portion of it is reflected. The typical reflections coefficient ($r$) for an MST radar is $10^{-9}$ to $10^{-3}$. The reflection coefficient tells us the ratio of the reflected wave amplitude relative to the incident amplitude. As here, this number is not big, capturing the backscattered signals needs lots of care. It is also possible that the reflected pulse hits another target and again scatterers. Still, in these situations, the backscattered signal becomes so weak that we can neglect that and ignore multiply reflected pulses [14].
2.2 Echoes

In Figure 2.3.a, we can see three clusters of scatterers and the returned pulse from them; These returned pulses are called *echoes*. The echo from a transmitted pulse can have zero lag, but the typical lags are in the order of microseconds. For example in Figure 2.3.b the second pulse is transmitted at a slightly later time and the scatterers had moved, and perhaps the backscattered signal’s strengths have changed too. When the second pulse is sent in this time lag, we can see echoes now have different amplitudes in comparison with the Figure 2.3.a, as all the signals from an individual cluster of scatterers have almost the same time lags, we can investigate the behaviour at different ranges. Hence if we change the time lags, we can change the ranges we are interested in. *Range Gates* are the ranges that we choose for data acquisition. This is an excellent example to show how radars can distinguish the atmosphere’s behaviour at different ranges. [14].
2.2. Echoes

Figure 2.3: (a) Shows the first pulse; amplitudes vary in strength as the scatterers change shape at different times and locations, and so vary in backscattering strength, and figure (b) The second pulse is sent in this picture but also the scatterers themselves have moved inside their groups. In (c), each pulse is sampled at a particular time delay like $R'$ and a time series is built up. Permission’s licence of this figure is available in the Appendix section. [15]
2.3 Polar Diagram

A transmitted pulse spreads spherically, which means it propagates both directly away from the radar and laterally. The most common radars are the dish radars, in which the radio signal is transmitted from the center of the dish. This is common in Astronomy, for example. Usually, in MST work, a typical antenna consists of several separate transmitting radars connected to each other to work as a big dish; an example of these smaller antennas is Yagi antennas. (Figure 2.4.b) If we set each Yagi antenna with a different radio wave phase, we can change the direction of wave propagation [16]. The radio wave’s power is related to the distance from the antenna \( r \) through the proportionality of \( 1/r^2 \) and the angle from vertical [17], as we can see in Figure 2.5. Usually, the desired radar is designed so that the maximum radiation is close to vertical (darkest colour in the picture). In reality, smaller and smaller amounts are produced at larger off-vertical angles; these secondary increases in signals are called sidelobes.

Figure 2.5 is called a Polar Diagram. Darker areas are representatives of higher powers, and the beam’s width is a measure of the beam’s concentration. In MST work, we prefer a narrower beam, as it gives better directional information. However, we also try to suppress the side lobes since a strong cluster of scatterers in the side lobes can dominate over a weaker scatterer in the main beam; thus, we might misinterpret the data [18]. Suppressing the sidelobes is achieved by careful modeling of the Yagi antennas’ placement within the radar area (the area which antennas are placed). [19]
2.4 Monostatic Continuous-Wave "Radar"

Figure 2.4: (a) A typical dish antenna, showing how a signal transmits from a point source at the top of the antenna, radiates into the dish, and transmits as a plane wave. (b) Showing how the difference of phase in each element can transmit a plane wave- Permission’s licence of this figure is available in the Appendix section. [15]

Figure 2.5: (a) A density plot shows the transmitted power as a function of an angle. The darker areas are the most potent parts. (b) A "polar diagram" shows the radiated power by curves along the radial direction Permission’s licence of this figure is available in the Appendix section. [15]

2.4 Monostatic Continuous-Wave "Radar"

Figure 2.6 shows a continuous wave (C.W.) radar with one transmitter and a receiver. The carrier wave, the waveform which will convey the information, has the frequency \( f_0 \) and angular frequency \( \omega_0 = 2\pi f_0 \). Typically, these signals are very low level, 30 to 40 Millivolts. At first, we assume the phase of our signal at \( t = 0 \) is zero, and it has the maximum voltage. Thus, we show it with \( \cos(\omega_0 t) \). Then it will be amplified with maximum power in the order of 5kW to 1MW, as we show it as \( p_0(t) = A \cos(\omega_0 t) \). \( p_0(t) \) shows the power and \( A \) represents the amplitude of electromagnetic wave. After the transmitted signal hits the target, it scatters and
returns to the receiver with a time lag, $t_{\text{lag}}$. This time lag makes a shift in the phase, therefore the phase shift is dependent on the distance the E.M wave is travelled. The received signal has an amplitude of $A$, where $\tilde{A} << A$. As $A$ representative of the returned signal’s power, we can see the returned signal is much less powerful, but we can amplify that just as the transmitted signal. The received signal is [20]:

\[
E_R(t) = \tilde{A} \cos(\omega_0 t - \varphi) = \tilde{A} \cos(\varphi) \cos(\omega_0 t) + \tilde{A} \sin(\varphi) \sin(\omega_0 t)
\] (2.1)

Our goal is to find the amplitude ($\tilde{A}$) and the phase ($\varphi$) as these give us the power and the velocity respectively, for which the method is discussed later. Still, we prefer finding $\tilde{A} \cos(\varphi)$ and $\tilde{A} \sin(\varphi)$ as they give us the same information; not only can we use them directly in the Fourier algorithms, but also they are continuous in time in contrast to $A$ and $\varphi$. Figure 2.6 describes it better. In this figure, to achieve this goal, we introduce a $90^\circ$ phase shift to one of the RF signals at point $p$, and they are mixed with the received signal, which is split into

Figure 2.6: A block diagram of a continuous-wave (C.W.) radar. Permission’s licence of this figure is available in the Appendix section. [15]
half at point $S$. Therefore, one of them is mixed with $\cos(\omega_0 t)$ and the other one with $\sin(\omega_0 t)$. At the end of the mixing process, there is a linear term plus a term which is proportional to the square of the sum of the received signal $\tilde{A}\cos(\omega_0 t - \varphi_r)$ and the reference signal (either $\cos(\omega_0 t)$ or $\sin(\omega_0 t)$) and higher orders. All the produced terms are either constant or have frequencies equal to $\omega_0$, $2\omega_0$, or higher. We need the mean signal; therefore, if we calculate the mean values of the linear term, the mean is zero, and the average of the second other terms is $(1 + \tilde{A}^2)/2 + \tilde{A} \cos(\varphi)$ in the case where we mix with the $\cos(\omega_0 t)$ reference. If we set $\tilde{A} << 1$, then we can write the mean values $1/2 + \tilde{A} \cos(\varphi)$ as in the first case and $1/2 + \tilde{A} \sin(\varphi)$ in the case that we mix the signal with the $\sin(\omega_0 t)$ reference. Then a low pass filter is applied and removes all the terms with frequencies $\geq \omega_0$, only time-independent terms are left:

\[ 1/2 + \tilde{A} \cos(\varphi) \] (2.2)

and

\[ 1/2 + \tilde{A} \sin(\varphi) \] (2.3)

With this approach, we find $1/2 + \tilde{A} \cos(\varphi)$ and $1/2 + \tilde{A} \sin(\varphi)$. These are named in-phase and quadrature signals respectively. [21]

More details are presented in Figure 2.7, which shows complex received signals. The upper figure shows the in-phase signal, and the bottom one shows the quadrature-phase signal. Both figures show five successful intervals $T_s$ ($T_s$ is the sample interval equal to the pulse repetition time) that are superimposed to show the signal’s relative change for stationary and moving scatterers. [22]

In Figure 2.8, we address the process with more details. There is a backscattered signal from a scatterer, which is moving towards the antenna. The distance of the scatterer from the antenna at any time is $r = r_0 - v_r t$ and the returned signal has the form

\[ \tilde{A}\cos(\omega_0(t - t_{lag})) \] (2.4)
Figure 2.7: A representative of in-phase and quadrature sampled echo voltage by Argand diagram during five sequences. The pulse-signal peak on the left side of pictures shows stationary scatterers (scatterers not moving relative to each other), and the signal in the middle of the picture shows the development of moving scatterers (scatterers moving relative to each other) during five successive pulses. [22]

Since the radio signal is transmitted first and received next, the time lag is twice the distance of the scatterer from the antenna divided by its speed.

\[
\Delta t = \frac{2r}{c} = \frac{2r_0 - v_r t}{c} \quad (2.5)
\]

Now with this definition, we can define the received signal as

\[
E_R(t) = \tilde{A} \cos((\omega_0 + \omega_r)t - \varphi_0) \quad (2.6)
\]

\[\varphi_0 = \frac{4\pi r_0}{\lambda_0} + \delta \varphi \quad (2.7)\]

is the phase at \( t = 0 \) (which contains distance information) and \( \delta \varphi \) is the additional phase delay with \( \lambda_0 \) being the transmitted wavelength and \( v_r \) the radial velocity of the target.

\[
\omega_r = \frac{2v_r}{c} \quad (2.8)
\]
2.4. Monostatic Continuous-Wave "Radar"

Figure 2.8: Formation of in-phase and quadrature components of a signal in the presence of a Doppler shift in the received signal. This figure applies for all times that the received signal is present. Permission's licence of this figure is available in the Appendix section. [15]

is added to the returned angular frequency, so the Doppler shift is visible here. The final signals, after mixing and applying the low-pass filter, in terms of in-phase signal \( I \) and quadrature \( Q \) are:

\[
I(t) = \tilde{A} \cos(\omega_r t - \varphi_0) \tag{2.9}
\]

and

\[
Q(t) = \tilde{A} \sin(\omega_r t - \varphi_0) \tag{2.10}
\]
Here, the positive radial velocity is defined to be when the scatterer moves away from the radar so that the Doppler frequency and Doppler velocity always have opposite signs.

Figure 2.9 shows the power spectrum of a signal received over many pulses of the radar signal (typically 10 to 30 seconds due to movement of our targets), with the absolute value of the complex Fourier transform shown. In this case, we have more scatterers in the radar’s field of view, all with various radial velocities and amplitudes, so different radial velocities appear as different spectral lines in this figure. Although range detecting is not discussed here, this is one of the most needed data for our work since radars are key tools for measuring the atmosphere, these data are crucial for better interpreting the radar results and therefore this thesis is important to better weather forecasting, thus it is addressed thoroughly in the next section.
Figure 2.9: (a) Typical in-phase and quadrature components received with an MST radar. (b) A typical power spectrum for the time-series shown in (a). In this case, a frequency scale is added only to give an idea of sorts of frequencies that are involved but the actual values depend on the sampling rate and duration of the time series in (a). Permission’s licence of this figure is available in the Appendix section. [15].
Figure 2.10: Further development of Figure 2.6 with more details of backscattered signal. Here the initial pulse is a Gaussian pulse, which amplifies and transmits. In the lower part of the figure, the combination of the received signal after $t_{lat}$ is shown. After applying a low pass filter, the part of the initial signal and the altered one is shown. The result of this process is in-phase and quadrature signals- Permission’s licence of this figure is available in the Appendix section. [15]
2.5 Pulsed Radar

Usually, radars produce a sequence of pulses rather than a continuous wave. A continuous wave transmitter operates without interruption. In contrast, a pulsed transmitter sends very short-duration signals. Figure 2.10 discusses a pulsed radar. Here there is at first an R.F. signal \( \cos(\omega_0 t) \), and the 90° shifted version of it for producing in-phase signal. Quadrature signals are also shown. This signal then passes into a "pulse shaping and timing" unit, where it is multiplied with a pulse-shaping function like a square pulse. In the figure, a Gaussian shaping profile is shown as it is preferred to be used than square pulse because square pulses can produce Fourier harmonics that can interfere nearby devices frequency spectrum and it is forbidden by many governments. Square pulses mathematically are:

\[
p(t) = \begin{cases} 
1 & \text{if } 0 \leq t \leq \tau \\
0 & \text{elsewhere}
\end{cases}
\]

Typical pulse length \( \tau \) is 1\( \mu s \) to 5\( \mu s \) for many atmospheric MST radar applications at frequencies of typically 50\( MHz \) (since the frequency at which we get best signal-to-noise response from the atmosphere is at (30 – 80 \( MHz \)). When this pulse is used as the modulating signal, then the returned signal is in the form of:

\[
E_r(t) = \tilde{A} e^{i[\omega_0 t - (2k_0 r + 2k_0 v_t) + \phi]} p(t - \frac{2r}{c})
\]  \hspace{1cm} (2.11)

Here in the Equation 2.11 the range \( (r) \) is present in both the main exponent and the pulse part \( p(t) \), considering \( p(t) \) shows the envelope of pulse and \( k_0 = 2\pi/\lambda_0 \). If there are two scatterers, at ranges \( r_1 \) and \( r_2 \) both of them modulate the carrier signal, and this effect appears in the returned signal. If we use filters; we had more temporal delays in the order of the inverse of the filter width [23]. As Figure 2.11 represents, if the scatterers are separated by less than the pulse length \( (\tau) \), we cannot visually separate the scatterers. Therefore, \( \tau \) indicates the resolution limits of the radar and is denoted by \( \Delta r = c\tau/2 \). Also \( k_0 \) represents the initial wave number and \( \omega_0 \).
Figure 2.11: A representative of the received signal $E_R(t)$ for two scatterers at ranges $r_1$ and $r_2$ for a square pulse of the temporal length of $\tau$. The envelope of the pulse can differ for various reasons like amplitude or time delay, which themselves can be a function of reflection and range, respectively. Permission’s licence of this figure is available in the Appendix section. [15]

is the initial frequency of the signal. We consider a general form of the transmitted signal as:

$$p(t) = A(t) \cos(\omega_0 t)$$  \hspace{1cm} (2.12)

where $A(t)$ is the envelope. This envelope can be in any form, such as Gaussian and Square. Finding the best shape of envelope is important as it carries information. The off set of it shows the radial velocity and the width of it shows the spread of radial velocities which is averaged over them. Regardless of the shape of the envelope, a radar should transmit a sequence of single pulses with regular intervals (inter-pulse period) to be able to capture different ranges. In this case, all the transmitting, receiving, and amplifying processes are the same as before, except that each received signal should be sampled at distinct ‘range gates’ and stored separately for different delays. As we know, collecting data happens at sampling points, which are called gates and the relationship of the gates is defined as $\Delta r = c \Delta t / 2$ [24]. This suggests that if there is a pulse with a half-power half-width (the angle between half power points of the main beam) of $2 \mu s$ ($c \Delta t / 2 = 300$ meters), and we want to sample from $1.2 \ km$ range to $12 \ km$ range, it would be necessary to sample at temporal delays of $8 \ \mu s$, $10 \ \mu s$, $12 \ \mu s$ up to $80 \ \mu s$. These
would be referred to like 1.2 km, 1.5 km, 1.8 km, ..., 12 km range gates.

2.6 Backscatter as a Convolution

Imagine we have multiple scatterers with different reflection coefficients \( r(z)dz \). The carrier signal is considered with an amplitude of form

\[
p(t) = A(t) \cos(\omega_0 t)
\]

We define the maximum (peak) of the pulse at \( t = 0 \). The strength of the backscattered signal at the time \( t_* \) needs to be measured. Therefore, the amplitude of the received signal at the peaks can be defined as:

\[
S_p(t_*) = r(z_0)p(0)dz
\]

This is the portion of the pulse which reaches the receiver at the time \( t_* \) from range \( z_0 \), but if in order for the equation to apply for the whole of the pulse when they reach at \( z_0 \), we should consider the parts that arrive after or before \( t_* \) at \( z_0 \). We want all the parts of the pulse to arrive at time \( t_* \), so then we should send the parts after the peak to the lower height and the ones before the peak at higher heights so that they all arrive at the time \( t_* \) to the receiver. Generally, it can be shown as:

\[
S_r(t_*) = r(z)p(t')dz
\]

Now we can write the relationship between \( t_* \) and \( t \) as \( t' + 2z/c = t_* \). Then replace it as:

\[
S_r(t_*) = r(z)p(t_* - 2z/c)dz
\]
The equation 2.16 can be simplified and written in such a way that it is dependent on just \( z \) and \( z_0 \) by using \( t_* = 2z_0/c \). Then the functions can be re-defined as \( S_z \) and \( p_z \) by

\[
S_z(z_0) = S_r(t_*)
\]

and

\[
p_z(z_0) = p(t_*)
\]

Finally, we can write the equation as

\[
S_z(z_0) = r(z)p_z(z_0 - z)dz
\]

For the total strength, we should integrate it as:

\[
S_{total}(z_0) = \int_{-\infty}^{\infty} r(z)p_z(z_0 - z) dz = r \ast p_z
\]

Which is the convolution between the reflection coefficient profile \( r \) and the pulse \( p_z \). Range detection is ignored here. As the amplitude drops by a factor of \( \frac{1}{z} \), we have the amplitude \( A_0/z \), at distance \( z \) from the source and \( A_0/z^2 \) for the backscattered signal. Taking the effect of range in the convolution equation we will have:

\[
S_{total}(z_0) = \frac{r}{z_0^{\xi}} \ast p_z
\]

Where \( \xi = 1 \) is for surface reflection and \( \xi = 2 \) is for volume scatter.

### 2.7 Combining the Pulse Radar’s Equation and the Polar Diagram

Every radar measures four parameters: (i) Absolute signal power, (ii) Noise power, (iii) Spectral offset, and (iv) Spectral width. These relate to (i) Reflection, or Scatter cross-section, (ii)
noise limitations as well as absolute calibration, (iii) radial velocity, and (iv) radial velocity variability (Doppler spread). The backscatter cross-section $\sigma_s$ is equal to the power backscattered per unit steradian per unit volume per incident power level. The scattering cross-section is a function of fluctuations of temperature, humidity, and the strength of turbulence. Since the Doppler shift represents the radial velocity (the component of the velocity of targets toward or away from the radar), the Doppler spread produces a spectrum width showing the distribution of radial velocities weighted by their power within the radar’s resolution volume. It is expected that this distribution is a function of beam width, turbulence, and wind shear.

In Figure 2.12, we can see the power spectrum, the spectral width, and the radial velocity of wind for April 13, 2014 from a VHF (Very High Frequency) radar in Costa Rica. VHF radars produce electromagnetic waves from 30 to 300 $MHz$. Here, the strong backscatters are interpreted as echo layers of turbulence. Since they have significant spectral width; the prominent oscillation in the radial velocity in the layers supports the idea of convection/turbulence. [26]

Now, if we mix the signal structure with a polar diagram, we can find both the velocity and the range. The received $E.M.$ wave can then be written as:

$$E_T(r, \theta, \varphi, t) \propto \left(\frac{A(\theta, \varphi)}{r}\right)e^{i(2\pi f_0 - (r - \frac{c}{2}) + \varphi'')}$$

(2.22)

[27] Where:

$\theta$ is angular distance from bore-sight of radar

$\varphi$ is azimuthal angle from the x-axis (generally, but not always, East)

$A(\theta, \varphi)$ is the position dependent amplitude due to polar diagram and scatterer characteristics

$r$ is range which also relates to the pulses time delay

$f_0$ is the transmitter’s frequency

$\varphi''$ is the constant transmitter initial phase

$c$ is the speed of light ($3 \times 10^8 m/s$)

Here $A(\theta, \varphi)$ is related to the antenna’s pattern used for transmission and reception, the strength of scatterers, and their shape. If we want to include this range detecting in the “in-phase” and
Figure 2.12: Plot (a) shows the power spectrum through time for April 13, 2014, in Costa Rica. (B) Shows the Spectral width through time, and (c) Shows Radial velocity’s evolution over time. They are all obtained by fitting a Gaussian model to the Doppler Spectra. The effect of layers and their development are pronounced in all the plots. [26]

"Quadrature notation, we can write them as

\[ s(t) = I(t) + iQ(t) = \tilde{A}e^{i(\omega_r t - \phi_0)} p(t - \frac{2r_0}{c}) \]  

(2.23)

Where \( \omega_r = -2k_0v_r \), \( v_r \) is the radial velocity and \( \frac{2r_0}{c} \) is the phase. It should be noted that this phase change is minimal. For a 6 m wavelength MST radar, a scatterer with a 40 m/s radial velocity would produce less than \( 10^{-4} \) rad phase change for a typical pulse (1 \( \mu s \)). As will be
discussed, we use this phase change between pulses to determine the Doppler shift.

### 2.8 Radial Velocity

To find the radial velocity, we use the phase shift. In order to recognize this phase shift, it is necessary that the scatterer move by much less than a half wavelength in the time interval \( T_s \). Besides, if there are multiple scatterers in the radar’s beam, the spectra become more complicated. The spectrum will have a width and a mean offset from zero Hz. The area under the spectrum is proportional to the total backscattered power, and the mean offset of the spectrum is a measure of the mean radial velocity measured in the radar volume [28]. The radar volume is the region covered by the beam width and the pulse length at the scatter’s height.

To determine the wind speed, it is necessary to investigate the radar volume and find the radial velocity. If we find the radial velocity, the horizontal wind speed can be calculated as

\[
v_H = \frac{v_r}{\sin(\theta)}
\]  

(2.24)

where \( v_H \) is the horizontal wind-speed component in the vertical plane in which the radar beam is tilted, \( \theta \) is the tilt of the beam from vertical, and \( v_r \) is the radial velocity, which is defined as

\[
v_r = \frac{-\lambda}{2} f(r)
\]  

(2.25)

At first, it is assumed that the vertical velocity component is zero, and the scatterers are all isotropic for the start [29], which is the base of our work. In Figure 2.13, it is obvious the scatterers are illustrated more isotropic in the center and less isotropic in the edges.

To determine the \( v_r \) we need to find \( f_r \) which can be difficult due to the presence of noise in the spectrum, as noise might interfere in the spectra and hide the \( f_r \). Noise cancelling then becomes essential and can be done through multiple methods which is addressed more, in the next paragraph. [30] In Figure 2.14, a typical spectral power density is shown, \( \Delta f \) represents
the Doppler shift of frequency of the transmitted signal. Here $S_r$ shows the power spectra of the transmitted signal and $S_N$ shows the power spectra of the noise. The dashed line indicates $S_N$, the mean value of the noise and the noise fluctuations level is shown by $\Delta S_N$, $\pm \Delta f_{\text{max}}$ are the limits for the number of points used in the **Coherent Integrating** technique (which is a form of signal averaging) and is determined by the transmitted pulse rate. Coherent integration is one of the noise-cancelling methods, which sums multiple sequential pulses samples collected from a specific echoing volume before spectral analysis [31]. This process increases Signal-to-noise-ratio (SNR) since the noise is randomly distributed and summing the signals cancels out the major part of the noise.

One of the earliest (and commonly used) techniques is to use weighted moments to calculate spectral offsets and spectral widths. For example, the mean offset is found as

$$\bar{f} = \int_{f_{\text{min}}}^{f_{\text{max}}} p_N(f) df$$

An alternative procedure for determining the value of $\nu_r$ is to employ the autocovariance
2.8. Radial Velocity

Figure 2.14: A typical Doppler spectrum (After Balsley, 1978) [31]

The function, given by

\[
\rho_c(\tau) = \sum_{j=1}^{n-k} s'(t_j)s(t_j + \tau)\delta t
\]

(2.27)

where: \( T = n\delta t \) is the data length, and \( \tau \) takes values of \( k \delta t \) for \( k = 0, ..., n - 1 \).

Although this function is computationally slower to evaluate than using the spectrum, it is unnecessary to form the entire function. We can just calculate it at the zeroth and first lag (\( k = 0 \) and 1). The radial velocity is written in the form of:

\[
v_r = -\frac{\lambda}{4\pi} \frac{d\phi}{dt}
\]

(2.28)

where \( \frac{d\phi}{dt} \) is the rate of change of phase of \( \rho_c \) at zero lag [32].
2.9 Spectral Fitting

In this section, we dig more into the process of data acquisition. The radars captured data in time series but not all of the data are beneficial for our work, even some of them should be deleted as they are not originated from atmospheric phenomena like meteors, aircraft, etc. In general, after data are saved in time series they are converted to a spectrum (by using Fourier transforms) that contains frequencies that can be interpreted as velocities. In some cases, after saving data into time series and before converting them to the spectrum, a suitable polynomial is fitted to remove slow drifts in the signal that are not due to atmospheric effects [33]. The mathematics of this process is explained more in the next chapter; here we are more focused on the practical approach of eliminating the noise and finding the best spectra.

After recording voltages for the antennas, the collected data are usually in a 20 – 40 s time series; spectra are formed by online analysis, and noise and effects of phenomena other than turbulence are eliminated. Distinguishing spectra due to air craft, meteors, etc, and eliminating them from the turbulence spectrum is an important topic that will be addressed in this section. We aim to remove the noise, and select the best spectra and fit a Gaussian function [27]. One of the essential characteristics that should be known is the spectral width and the power spectrum. If there is any noise, the power spectrum is non-zero across the whole frequency domain. In addition to the power spectrum, other interfering spectral components can affect the spectral variance. In most cases, a Gaussian function is fitted to the spectrum (as in nature we do expect the backscattered spectrum be in Gaussian shape), and then the least-squares chi-square parameter, $\chi^2$ is calculated. After that, the spectra, which is bi-modal or have other non-atmospheric components like aircraft or radio interference, are rejected. The Gaussian function used is in the form of

$$A_0e^{-(f-f_0)^2/2\sigma^2} + D_0$$  \hspace{1cm} (2.29)

$D_0$ is an offset that shows the level of noise, thus larger values of noise shows larger values
of $D_0$ and also larger values of $\chi^2$. $A_0$ is the amplitude, $f_0$ is the offset of the peak and $\sigma_f$ is the standard deviation of the spectrum in Hz. Rejecting the spectra for that have $\chi^2$ greater than $\chi^2_c$ (which is the specified limit for $\chi^2$), determines acceptable spectra [34]. In the next step, after the determination of spectral width, we look for non-turbulent effects because by eliminating their impact, we are one step closer to calculating the turbulence’s contribution.

In this process, we minimized the usage of coherent integration, but we digitized a large amount of raw data and then applied Fourier Transform. Here up to 65,000 points of data per height and per record are transformed. As large data streams with small time steps between points are used, aircraft’s or meteors’ are detected. For aircraft the radar detection numbers can vary from $1 \sim 10$ per day in remote sites to $100 \sim 200$ per day when close to an airport (such as Harrow, which is close to the Detroit airport), on the other hand for meteors, anywhere from 1000 to 10,000 meteors per day can be determined if the system is designed. However, after our detection algorithms are applied, only maybe 1 or 2 aircraft and at most 4 – 5 meteors will contaminate our signal per day.

This method also allows calculation of 'large frequency band' up to $\pm 100$Hz. This bandwidth allows us to find aircraft from the signal because they have large velocities, and as a result, they produce huge frequencies, so they are shifted out in the spectrum to arrange rejected frequencies. In the next step of the procedure, the algorithms start to search for the maximum peaks in the spectra, then narrows the search in regions where the peaks are. After recording our wide-band spectrum, a low pass filter is applied. The user examines the spectrum, finds the signal peak, then applies a low-pass filter or band-pass filter around these peaks. This method is very efficient for small spectral widths like those resulting from radial velocities that are small. It lets the user choose small bandwidths like 1 Hz to 4 Hz.

On the other hand, the user can use a wider filter for off-vertical beam data as they have larger radial velocities, and therefore they produce larger frequencies. It should be noted that in most of the cases no filter is needed, but the search for the spectral peaks usually is in the range of between $-4$Hz to $+4$Hz. This choice of filter throws out the spectral peaks due to aircraft as
they have large velocities, therefore they make huge peaks in greater frequencies. After finding the spectral peaks associated with the atmospheric echoes, we can apply the Gaussian function. After applying the suitable Gaussian function, the determination of the spectral widths, offsets, and peak spectral densities becomes easy. This procedure allows obtaining the atmospheric echoes from 1 to 13 km height for the vertical beam and 2 to 8 km for the off-vertical beam, directed at 10.9° from the zenith, precisely the angle we are using in the next chapter. This angle is chosen since with this method the first null of the radar beam is placed in the vertical direction and minimizes interference from directly overhead while the beam is pointed off-vertical. The reason that the vertical beam can capture higher altitudes is because of the existence of so-called specular reflectors which are approximately horizontally aligned. Specular reflection is a mirror-like reflection from the tilted layers of the atmosphere. [35]

As an example, Figure 2.15 shows a spectrum that contains atmospheric echoes, meteors, and an aircraft. As can be seen from this figure, the aircraft shifted the spectrum away from 0 Hz. In traditional VHF profiler systems, aircraft’s signals render the data and make it useless. Despite all of the good results of this method, it cannot eliminate the effect of aircraft that move perpendicular to the radar. In this case, the Doppler shift is tiny and it is estimated that only 10 – 15% of aircraft cause such a problem [33]. Figure 2.15 shows the spectra recorded from 7 km altitude with a relatively good signal to noise ratio. Now, it is time to go through all the steps. First, we remove the meteor’s spectrum. Sometimes meteors associated with 80 – 110 km are range aliased into the data as the radars operate at a very high pulse repetition frequency, 10 kHz. This broadens the spectra; therefore, it can corrupt the atmospheric signal. Meteors show their impact by a rapid increase in amplitude followed by a fall back to normal levels in less than 3 – 4 s. Meteors are best projected in the time domain, while the aircraft is best rejected in the frequency domain. [36]

The overall approach, up to here was removing the meteors and strong aircraft from the spectra.

After removing the meteors, the search for spectral peaks begins. We start with the outer
Figure 2.15: This spectrum contains aircraft, meteors, atmospheric signals, and ground clutter recorded by an off-vertical beam. The length of the data series is 90 s in this spectrum. The upper figure shows the middle part of the spectrum, middle ± 4Hz. The meteor’s effect and the atmospheric spectrum are distinguished. The lower graph shows the whole spectrum. The peak made by aircraft is prominent. [33]

part of the spectrum (between −4 to −0.45 Hz and 0.45 to 4 Hz), which we typically expect to contain less atmospheric information, and the central part (−0.2 to 0.2 Hz) is completely removed. In the next step, we calculate the auto-correlation for this part of the spectrum by
applying the Fourier transform. Then the first estimation of the radial velocity is made by finding $\frac{d\phi}{dt}$ close to zero lag for the auto-correlation function. As mentioned before $\phi$ depends on the range, so $\frac{d\phi}{dt}$ gives information about the radial velocity. Next, we go back to the spectrum and search for the largest values in the spectrum. These four peaks are close to each other; usually, they should lie within 1.5 Hz of each other. If they are close to each other, we find the average of their respective frequencies; therefore, we obtain several radial velocity estimates. Later, it is checked if the radial velocity determined by the auto-correlation is close to the ones determined by the spectral peak or not.

Several other tests are applied to the data, which is out of context here, but if all these tests are passed, it is the time for applying the Gaussian function and a constant fitting. After determining the constant, we try to determine the least-squares fit to the data. The initial guess is from our first estimate of the spectral peak position and the power in that region, then by trying different spectral offsets, peak values, and spectral widths, choose the least square fit. If there were no spectral peaks in the spectrum’s outer edge, we would search the central part. There are other measurements taken regarding eliminating other noises like instrumental oscillations and ground clutter. Data will be rejected if the signal to noise ratio is less than 0.3 or if the phase change in the auto-correlation function is non-linear between successive lags as it is acceptable in atmospheric work [36]. After passing multiple tests, the radial velocity is finally reported.

Figure 2.16 is one example of spectra before and after applying the above procedure. The left-hand side column of figures shows the raw spectra, Where data are binned in groups of three or four spectral lines. The right-hand side of columns of pictures shows them after eliminating noise and fitting the best Gaussian function, making it suitable for determining the radial velocity. For example, the 4 km data is rejected after tests because it had a secondary peak at the negative frequencies, and also, the spectra at 8 km are rejected because of the poor SNR. The spectra are accepted in all other cases, and the best Gaussian function is fitted as it is shown. This approach also helps to eliminate specular return effects.
At the final step, there might be some noises that are not eliminated with all these procedures. In such a case, we should do it off-line and look for the outliers in each case specifically. These are aircraft or some ionosphere echoes that can not be detected before, although only 2 – 3% remain, and the rest are wiped out in the previous procedures. However, how can we understand some peaks are outliers? One of the examples is symmetry. If there are some peaks in the eastward direction of the beam, there should be some in the westward beam as well. If not, we can conclude we are looking at outliers as the beams themselves are symmetric, hence if one of them captures an approaching object, the other one captures the same object as a leaving object. In overall, this procedure is very efficient in both eliminating the noise like meteors considering the fact that 400 meteors or even more can be detected per day [37] and keeping the data in high quality simultaneously.

In conclusion, all the above measurements are taken to decrease the effects of noise and have a better spectrum that can produce a better estimation of velocities as this is the data we need.
Figure 2.16: Sample spectra collected by CLOVAR (Canada London Ontario VHF Atmospheric Radar). The left-hand graph shows the raw spectra, and the right-hand side shows the accepted spectra after applying all the tests mentioned above and procedures and applying the best fit Gaussian function. [37]
Chapter 3

Gravity Waves

This chapter describes gravity waves, their origin, and their impacts, and the reasons we are interested in them. Also, here we discussed how gravity waves can produce turbulence and then scatterers. Scatterers are our interested object here as they are detectable with our radars. The relationship between gravity waves and scatterers is described in the last section.

Gravity waves are one of the strong sources of turbulence-generation in the mid atmosphere and lower atmosphere [38]. Thus they are responsible for the transportation of momentum and energy and are extremely important in atmospheric circulation. Many ideas exist about how they originate in the atmosphere [39]. The main sources of gravity waves in the troposphere are convection and frontal systems [40], thunderstorms [39], wind shears [41]. The best place for observing gravity waves is the mesosphere as they have a large amplitude in this altitude [42].

As gravity waves are transferring momentum flux and energy between different points in the atmosphere, they have a significant impact on atmospheric motions. Gravity waves were not properly appreciated prior to the 1960’s, and the first paper that really expressed their importance was that due to Hines in 1960. [43]. Usually, they are responsible for transporting energy from various sources, including mountains, to other points [44]. The impact of gravity waves reaches its greatest in the stratosphere and mesosphere, where they are so important that they can change the direction and the speed of mean wind there. They may also affect the mo-
mentum and temperature at the same time. It can even reverse the direction of mean eastward winds (zonal) or north-south (meridional) winds, which was not considered in earlier models. These waves are still not included in many meteorological models and weather prediction models.

One of the significant gravity waves traits is its large-scale consistency with sufficient averaging over months and geographical coverage. The wave spectra and the spectral powers show only modest variations as a function of season and latitude. This means its changes due to latitude or season is very small. This trait brings up the concept of “universal spectrum“, which means the distribution of wave spectral densities is invariant in terms of amplitude and shape as a function of latitude, longitude, and time [45]. However, this “universality“ is most apparent in the mesosphere and upper stratosphere: the universality in the troposphere is less assured. Also, it should be noted that they are responsible for local phenomena, which makes them more interesting to study as they effect our lives directly.

3.1 Impacts of Gravity Waves

Gravity waves break for various reasons; Specific details will be discussed in the next sections. For the moment, the main thing is that the wave-breaking leads to turbulence in most cases. [46] Another outcome of breaking gravity waves is shedding, which is Convective adjustment [47]. Convective adjustment is often known for situations where that instability is removed because of vertical mixing or diffusion [48]. This can happen in cases where the wave does not break catastrophically, and therefore sheds the energy gradually and maintains a largely time-invariant spectrum.

Another huge effect of gravity waves is the effect of their momentum flux on the mean wind, especially in the mesosphere. The force induced by gravity waves in the upper levels can be strong to the extent to reverse the zonal flow that had been produced because of radiative equilibrium in the upper levels of the atmosphere [49]. This phenomenon leads to the merid-
3.2 Local Impacts of Gravity Waves

Orographical flow causing a rise in the air at the summer pole, and falling the air in the winter pole. This has a surprising result in generating heat in the winter pole and cooling in the summer pole the by adiabatic process. Thus, summer in mesopuase is as cold as $120\, K$ [50].

Another effect of gravity waves, that can be mentioned is their fingerprint on planetary and tidal oscillations. They can even generate new planetary waves, and this also can occur in a reverse manner, meaning the planetary wave oscillations changes the gravity waves [51]. Planetary or Rossby waves are Synoptic scale Waves that happen because of Coriolis force which happens as a result of rotation of earth [52]. The rotation changes the direction of wind current that is moving from equator to the north pole, to right and vise versa in the southern hemisphere. [53]

3.2 Local Impacts of Gravity Waves

Our focus in this thesis will be on tropospheric and lower stratospheric regions, as these are the neglected regions for gravity waves’ search, so we now look in more detail at these lower altitude phenomena as One of the most interesting local effects of gravity waves, at least for this thesis, is ‘Chinook’ Winds. Chinook waves are strong, periodic winds that flow form the mountains to plains of the eastern side of Rockies in Western Canada (as shown in Figures 3.1 and 3.2). They alternate between strong winds, then calm winds, then strong winds again, with the periodicity of a few tens of minutes. This periodicity is a product of gravity waves that have generated over the mountains which are called lee waves. [54] [55]

Moreover, the most significant result of gravity waves, turbulence, should not be neglected. As the gravity wave breaks, the most possible scenario is the production of turbulent motion. It should be noted that scatterers are caused by this turbulence, and the study of such scatterers is the key to investigating gravity waves.
Figure 3.1: Dotted area shows Chinook waves at 1800 GMT, in 15 April, 1963, Denver, Colorado [54]
3.2. Local Impacts of Gravity Waves

Figure 3.2: Clouds that are formed as a result of Chinook winds over the Rockies in Banff National Park in Alberta, Canada. (Image credit bought from: Autumn Sky Photography/Shutterstock)
3.3 What is a Gravity Wave

Gravity waves are mesoscale or synoptic scale, atmospheric waves that propagate over long distances, which are observable by MST radars at heights of 1 to 10 km, and also above the troposphere. They have horizontal wavelengths of few to hundreds of kilometers and vertical wavelengths of 1 to 30 km. They produce velocity, pressure, and density fluctuations in the air, in the form of sinusoidal propagating waves; therefore, they carry momentum flux and energy. They are also called Buoyancy waves or Internal gravity waves. They are responsible for wind and temperature fluctuations with periods of minutes to a few hours. The biggest effect that had been discussed before is the dissipation of kinetic energy of motion to turbulence and then to molecular kinetic energy, which leads to the release of heat.

3.4 The Generation of Gravity Waves

Gravity waves can be best understood by considering a simple model in which they are generated by uniform flow over a corrugated boundary. In this case, the air flows by oscillating over the corrugation and hence causes a propagating wave. These waves are unusual in a way that their phase velocity and group velocity are perpendicular to each other. As they rise in the air, the atmospheric density decreases; therefore, energy conservation dictates that their velocity amplitude increases from a few centimetres per second to several tens of meters per second at heights 70 km and above [56].

It is easiest to understand the wave generation if we consider a corrugated surface (representing the mountains) being forced to move horizontally through the air, as shown schematically in Figure 3.3. In this figure, there is a tip of a mountain at point A, which forces the air above it to move upward, so now this particle of air has a vertical velocity of $w'$ and horizontal velocity of $u'$; therefore the particle of air is pushed along the purple line with the slope pointing to right and up. Consequently, the air along this line is compressed, and pressure increases (showing by the gray area and with letter $H$). This gray area can be called a high-pressure area.
On the other hand, the area above point B has a lower pressure, as the surface beneath this point is falling. This region is called low-pressure region and representing with letter L; Hence the air velocity is downward and to the left, opposite to point A. Also, since the whole structure is moving forward to an outside observer, the wave-front seems to move in the direction of the big gray array (labelled as: Wave-fronts appear to move this way) [57]. This arrow shows the apparent propagation direction of the wave with the rate of phase speed. Do not forget as the air above point A is pushed to the up-right, it transports energy in this direction. The tiny arrow which is labelled as *Energy propagation* shows it, and energy direction and phase direction are unlike sound waves. If we look at the phenomena from the frame of corrugation, the whole system appears stationary, but from the point of view of the air, it is seen as a propagating wave.

### 3.5 Mathematical Definition

Considering Navier-Stokes equation:

\[
\frac{D\vec{u}}{Dt} + 2\vec{\Omega} \times \vec{u} - \vec{g} + \frac{1}{\rho} \nabla p - \nu \nabla^2 \vec{u} = 0 \quad (3.1)
\]
And combination of the first law of thermodynamic and the second law of Newton and continuity gives:

\[ \frac{D\rho}{Dt} = \frac{1}{C_s^2} \frac{Dp}{Dt} \]  (3.2)

the continuity equation itself:

\[ \frac{D\rho}{Dt} + \rho \nabla \cdot \vec{u} = 0 \]  (3.3)

and finally Fick’s law of heat transport:

\[ \frac{D\Theta}{Dt} = \frac{k}{\rho} \nabla^2 \Theta \]  (3.4)

Here notations are: The total velocity is \( \vec{u} \), the density is \( \rho \), \( \vec{\Omega} \) is the Earth’s angular velocity (with magnitude \( \Omega \), \( \times \) means cross product, \( \vec{g} \) is the acceleration due to gravity = [0, 0, \(-g\)], \( p \) is the pressure, \( c_s^2 \) is the speed of sound squared, \( \nabla \) represents the gradient differential operator and \( \cdot \) means the dot product. \( \Theta \) represents potential temperature, \( k \) the heat diffusion coefficient, and \( \nu \) is the kinematic viscosity coefficient (which is of course, just the molecular viscosity coefficient divided by the atmospheric density). Now for solving and using these equations, we use the gravity wave’s approach. Buoyancy waves exist at angular frequencies less than Brunt-Väisälä frequency or \( \omega_B \). One solution for the form of gravity waves can be

\[ \Psi = \Psi_0 e^{i(\vec{k} \cdot \vec{x} - \omega t + \phi)} \]  (3.5)

where \( \Psi \) can be any component of velocity such as pressure, density or temperature, although it can be complex. The application of complex numbers to represent these variables is just a standard technique used to optimize analysis of sinusoidal oscillations. [20] Here \( \omega \) is the frequency of the wave and \( k \) is the wave number vector \( (= [k, l, m]) \), while the position \( \vec{x} \) is defined as \((x, y, z)\). Here it is assumed that \( w \) and \( k \) are real, but as the wave might grow
or decay while rising up, \( m \) is assumed to be complex; also, the wave is propagating in \( x - z \) plane. For solving first order perturbations, we have:

\[
-\iota \omega \hat{u} - f_c \hat{v} = -i k \hat{\psi} \tag{3.6}
\]

\[
-\iota \omega \hat{v} + f_c \hat{u} = 0 \tag{3.7}
\]

\[
-\iota \omega \hat{w} + \hat{\psi} g = -(im \hat{\psi}/H) \tag{3.8}
\]

\[
-\iota \omega \hat{r} - \hat{\psi} \omega_B^2/g = -i \frac{\omega \hat{\psi}}{C_s^2} \tag{3.9}
\]

\[
-\iota \omega \hat{r} + i k \hat{u} - \hat{w}/H + im \hat{w} = 0. \tag{3.10}
\]

[20]

Where \( \iota = \sqrt{-1}, c_s^2 \) is the mean squared speed of sound at the height of the wave, \( \hat{\psi} \) is first order perturbed \( \psi \), and \( f_c \) is the Coriolis parameter, equal to \( 2\Omega \sin \theta \), where \( \theta \) is the latitude. The parameter \( m \) is the vertical wave number, which is complex, equal to \( m_R + im_I \), where

\[
m_I = -1/(2H) \tag{3.11}
\]

and describes an exponential increase in amplitude with increasing height. Introducing \( H \) is the scale height, we can use this relation: \( H = \frac{kg_T}{m_g} \) with \( K_B \) the Boltzmann's constant and \( m \) the mean molecular mass. \( \hat{\psi} = \frac{\hat{p}}{\hat{T}} \) and \( \omega_B \) is the Brunt-Väisälä frequency:

\[
\omega_B^2 = \frac{g}{H} \frac{\partial T}{\partial z} + \epsilon \frac{g}{H} \frac{\partial \theta}{\partial z} = \frac{g}{H} \frac{d \theta}{dz} \tag{3.12}
\]

where \( \epsilon = R/c_p \), and \( R \) is the gas constant and \( c_p \) being specific heat constant of air. Also symbol \( \theta \) represents potential temperature [58]. While solving these equations, considering
dispersion equation is necessary. The dispersion equation is the relationship between the wave frequency and wave number. If the initial angular frequency, \( \omega_i \), was equal to \( f_c \), then the inertial period is: \( T_i = \frac{2\pi}{f_c} \).

Now it is time for the most general but simplified dispersion equation:

\[
m_R^2 = \frac{\omega_B^2 - \omega_i^2}{\omega^2 - \omega_i^2} k^2
\]

or even simpler version:

\[
m_R \frac{k}{\omega} = \frac{\omega_B}{\omega}
\]

Getting help from polarization equations, we can get:

\[
\hat{\omega} = -\frac{k}{m_R} \hat{\omega} = \frac{\omega}{(\hat{u} - c_\Phi) m_R} \hat{\omega}
\]

\[
\hat{\nu} = -i \frac{\omega_i}{\omega} \hat{\omega}
\]

\[
\hat{\Theta} = -i \frac{\omega^2}{\omega} \hat{\Theta} \hat{\xi} = \frac{\omega}{g} \hat{\Theta} \hat{W} = \frac{-i}{m_R(\hat{u} - c_\Phi)} \frac{d\Theta}{dz} \hat{\xi}
\]

where \( \hat{\xi} \) is the vertical displacement and:

\[
\frac{\hat{\Theta}}{\Theta} = -\frac{\hat{\rho}}{\rho}
\]

\[
\hat{\rho} = \hat{\rho}(c_\Phi - \hat{u})
\]

In these equations, \( c_\Phi - \hat{u} = \omega/k \) is the so-called “intrinsic phase speed” of the wave. Here true these equations are true only if \( \omega_i \ll \omega \ll \omega_B \). \( \hat{\Theta} \) represents the difference between its own potential temperature and that of its immediate environment. The most important parameters used for gravity wave studies are often the velocity fluctuations, the temperature, density, pressure fluctuations, and the direction and speed of propagation. The other characteristics of them also are wave periods (or frequencies), and vertical, horizontal, and total wavelengths. [59]
3.6 Gravity Waves, Turbulence, Scattering

It has been mentioned before, as the wave rises into the air, it grows and becomes unstable, so generates relatively small amounts of turbulence (compared to the total energy of the wave) and loses energy [60]. By reducing its amplitude, it becomes stable again, and this process happens multiple times until all the waves reach the amplitude that they break. Finally, turbulence occurs; On the other hand, instability might not happen, which can lead to shedding in energy and momentum in the direction of the wind [61]. One of the best criteria for measuring if turbulence occurs or not is Richardson Number. Richardson number is defined as:

$$Ri = \frac{\omega_B^2}{(\frac{du}{dz})^2}$$  \hspace{1cm} (3.20)

Which here $\omega_B^2$ is the Brunt-Väisälä frequency as shown before and $\frac{du}{dz}$ shows the stratification of the environment. For Boussinesq fluid (kind of approximation that variations in fluid properties other than density are ignored) the critical Richardson number ($Ri_c$) is $1/4$, meaning if $Ri < Ri_c$ or in this case $Ri < 1/4$ there will be a turbulence in this fluid [62].

An unstable atmosphere happens when a parcel of air that is rising adiabatically reaches the point that it is warmer than its environment; therefore it continues to rise rather than oscillate. In this thesis, we are interested in turbulence as our radars are designed to use turbulence to produce the backscatters. Important parameters that need to be measured are kinetic and potential energy dissipation rate, ($\chi$), refractive index structure constant ($c_n^2$), the mean potential refractive index, and the degree of anisotropy, among others [63]. Structure function gives essential information about the turbulent structure as spectra can be driven from it later. The structure function usually is measured by a probe moving in a line through the fluid. The total structure function is defined as:

$$D_{tot}(r) = |\vec{n}(\vec{x} + \vec{r}) - \vec{n}(\vec{x})|^2$$  \hspace{1cm} (3.21)
In equation 3.21, \( \vec{n} \) is the refractive index and \( \vec{r} \) is any direction we are looking at medium. Now the relationship of the structure function and the potential refractive index structure constant \( (C_n) \) is in the form of:

\[
D_{tot}(r) = C_n^2 r^{2/3}
\]  

(3.22)

By determining the structure function, then the potential refractive index structure constant \( (C_n) \), the kinetic energy dissipation rate \( (\bar{\varepsilon}) \) can be determined as described below:

\[
\bar{\varepsilon} = \left( \gamma C_n^2 \frac{\omega_B^2}{F_t^{1/3} M_n^2} \right)^{3/2}
\]  

(3.23)

Where \( \omega_B^2 \) is the Brunt-Väisälä frequency, \( F_t \) is the fraction of the radar volume which is filled by turbulence and \( \gamma \) is assumed to be a constant (but a debatable assumption) and \( M_n \) is the potential refractive index gradient [65].

Determining the degree of anisotropy was our final goal in this project, as it is crucial in the troposphere. For instance, Hocking found a strong correlation between near-isotropic turbulence and rainfall occurrence, which can play an essential role in meteorology [15]. As discussed, the refractive index is the most crucial atmospheric parameter for MST radars. This refractive index is correlated with electron density, pressure, temperature, etc. The refractive index is the ratio of an electromagnetic wave’s speed in a vacuum \( (c) \) over its speed in our interested medium [66]. When the radar’s signal enters the atmosphere, it frequently hits different types of regions with different refractive indexes; these changes in the refractive index cause a measurable fraction of the radio wave to be reflected [67]. Reflected signals return to the radar’s receiver and give us our primary data, as was discussed in the first chapter.

However, determining the backscattered signal strength is not as easy as it appears. If these scatterers have various shapes, like anisotropic shapes, guessing the vertical velocity and measuring turbulence gets more complicated. Shapes of scatterers alter the measurements of vertical velocity [68]. For example, if the atmospheric scatterers are not horizontal exactly and have a small tilt, then identifying the scatter’s direction is not as easy as it was discussed in chapter 1. If a scatterer exists in a slightly off-vertical position, or if the scatterer is tilted, it
leads to confusion in interpreting the scattered radial velocities. Also, having smaller scatterers than the beam width produces some issues in understanding vertical velocity measurement. In this project, we try to get closer to understanding these scatterers hoping they can be modeled someday and can reduce the effect of these issues on the radar measurements.
Chapter 4

Our Radars

This Chapter describes our radars and their features and locations. As Dr.Wayne Hocking uniquely made these radars for detecting scatterers in the troposphere, it is essential to understand how they are designed.

It is better to look at the radars used in our studies closely. We use the O-QNET (Ontario-Quebec Network) radars and the Eureka radar for our studies. The O-QNET radars are located over a wide area in Ontario and Quebec and cover many different locations, varying from lakeside to hilly and flat regions. These radars are in Abitibi Canyon, Aumond, Clovar (London), Egbert, Gananoque, Harrow, Markstay, McGill, Negro Creek, Walsingham, Wilberforce, as well as in Eureka in the arctic Figure 4.1 and Figure 4.2. I have described that these are wind profilers, which we use to measure the wind's speed in the troposphere and lower stratosphere. Typically, they have been used for measurements from altitude 0 to 13 km above the ground with a vertical resolution of 0.5 km.

The data from the Walsingham radar was analyzed extensively as it had the most clean data, and two other radars that were studied in considerable detail were Wilberforce and Harrow. The first reason for choosing these radars was the quality of data, since these radars have the cleanest data. The second reason for choosing these sites is that they are in variant locations meaning they experience variant geophysical effects for example Walsingham and Harrow are
Figure 4.1: Locations of Ontario’s radars including Negrocreek, Wilberforce, McGill, Walsingham, and Harrow.

sites that are close to lakes but Negrocreek and Wilberforce are not close to any lake or seas. Furthermore, Negrocreek and McGill are chosen because of their unique traits. Eureka is an arctic site which is on a frozen lake and McGill experiences many unpredictable weather phenomena. Here, the Walsingham radar’s principal parameters are shown in Table 4.1 as an example, but these radars have many common features. Examples of common features are the peak power, measurement modes, duty cycle, height resolution, number of beams, and ranges covered. Adjacent radars are separated by typically 200 km. The raw data are recorded at typically 1000 – 3000 Hz, depending on the radar operation mode. This frequency band is chosen
Figure 4.2: Location of all the sites within Canada and the Canadian Arctic

as a balance between avoiding range-aliasing which keeps the pulse-repetition-frequency low and optimizing the signal-to-noise ratio which requires a high pulse-repetition-frequency. Then these data are processed in real-time to condense the information to a more compact form. For example, coherent integration [which we talked about it in the chapter 1, at page 29] is typically performed over 16 points, making the frequency range in the spectral domain of the order of $30 - 100 \, Hz$ to deal better with noise. The radar beams are steered to 4 azimuthal direction sequentially (North- South- East- West) and then vertically. One complete cycle typically takes 5 minutes. These radars used 20 – 40 s duration time series of data (as it gives the optimum
signal reliability) and analyzed them spectrally like calculating their spectral width.

Spectral fitting procedures are used to determine the mean spectral offsets as discussed in chapter 1, spectral widths, and power levels, and these data are stored in a more compact form for posterity. Winds’ velocity and turbulence strengths are then calculated using other software. We use these final products in our analysis.

The various radars do differ in some ways, such as frequency, antenna configuration, and, subsequently, the beamwidth. Figure 4.3 shows the two main different setups used in our radars. Figure 4.3(a) shows the design used in Walsingham, and Figure 4.3(b) shows the design used for the Harrow and Negro Creek radars. Walsingham’s layout is called (type I) radar, which means it has a larger cross-structure, resulting in a narrower beam with a larger side lobe. On the other hand, the layout at Harrow and NegroCreek is more compact (type II). The spacing of NegroCreek’s radar is 1.25 wavelengths diagonally between quartets, which makes a broader main beam, and Harrow has an intermediate beam as seen in Table 4.2

As shown in the layouts, our radars are not big dishes, but instead, they are arrays of coupled antennas, which work like one big dish. The antenna elements are called ”Yagi antennas,” and the whole array typically consists of 128 antennas as it was talked about in chapter 1. Here are 128 antennas in each case. It should be noted that the signals from all antennas are coupled together with suitable phase shifts and fed into a single receiver in the main radar building.

Each Yagi antenna itself is both a transmitter and a receiver, and each of them is made up of three horizontal bars Figure 4.4, comprising, from the top: the director, the driven element, and the reflector. To steer the beams of these arrays, there must be time delays in the signal sent to some of the Yagi’s. Each antenna is fed by a cable, which is an integer number of wavelengths from the transmitter. If there are no net delays, the main beam is pointed precisely perpendicular to the radar plane. However, by adding delays using the feeder, it adds phase shifts to its signals. A time delay of approximately 20 ns applied to these antennas results in a one-wave period, or 2π radians of phase. Usually, phase delays are less than this, and by appropriate choice of phase delays, the beam can be steered around the sky. This method
Figure 4.3: (a) Layout of type-I radar and (b) layout of type-II radar. In both layouts, antennas are clustered in groups of 4 with a separation of half a wavelength between the antennas. There are 128 antennas in each case. The signals from all antennas are coupled together with suitable phase shifts and fed into a single receiver in the main building. [36]

results in approximately a plane wave (At distances greater than one Fresnel zone) radiating from the array of antennas, with different directions for the different beam directions.
Figure 4.4: Yagi Antennas in Eureka, NU.

Figure 4.5: Eureka’s station. Mountains are seen in the background.
## Table 4.1: Walsingham’s Radar configuration.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Location</td>
<td>Walsingham(42.6377°N; 80.5730°W)</td>
</tr>
<tr>
<td>Operating frequency</td>
<td>44.5 MHz</td>
</tr>
<tr>
<td>One-way half-power half-beam width</td>
<td>2.3°</td>
</tr>
<tr>
<td>The total area of the antenna field</td>
<td>4000 m² (partially filled)</td>
</tr>
<tr>
<td>Mean power</td>
<td>3200 W</td>
</tr>
<tr>
<td>Peak power output</td>
<td>40 kW</td>
</tr>
<tr>
<td>Gain</td>
<td>25 dB</td>
</tr>
<tr>
<td>Wind measurement mode</td>
<td>Doppler</td>
</tr>
<tr>
<td>Pulse length</td>
<td>500, 1000 (m)</td>
</tr>
<tr>
<td>Mean power aperture product</td>
<td>$1.6 \times 10^7$ Wm²</td>
</tr>
<tr>
<td>Duty cycle</td>
<td>0.5 – 10%</td>
</tr>
<tr>
<td>Height resolution beams</td>
<td>0.5 – 1 km</td>
</tr>
<tr>
<td>Number of beams</td>
<td>$5(Vert. + 10.9° off – vert.toN, S, E, W)$</td>
</tr>
<tr>
<td>Range (off-vertical beam)</td>
<td>0.4 – 14 km</td>
</tr>
<tr>
<td>Range (vertical beam)</td>
<td>0.4 – 14 km</td>
</tr>
<tr>
<td>Digitizer aliasing frequency number</td>
<td>&gt; 100 Hz</td>
</tr>
</tbody>
</table>

Duty cycle: Pulse length (seconds) divided by the time between pulses (seconds).
<table>
<thead>
<tr>
<th>Radar Parameters</th>
<th>Walsingham</th>
<th>Harrow</th>
<th>Negro Creek</th>
</tr>
</thead>
<tbody>
<tr>
<td>Location</td>
<td>42.6377°N and 80.5730°W</td>
<td>42.039°N and 82.892°W</td>
<td>44.632°N and 80.859°W</td>
</tr>
<tr>
<td>Operating frequency</td>
<td>44.5 MHz</td>
<td>40.68 MHz</td>
<td>48.92 MHz</td>
</tr>
<tr>
<td>One-way half-half beam width</td>
<td>2.3°</td>
<td>2.75°</td>
<td>3.30°</td>
</tr>
<tr>
<td>Two-way half-half beam width</td>
<td>1.63°</td>
<td>1.95°</td>
<td>2.34°</td>
</tr>
</tbody>
</table>

Table 4.2: Walsingham, Harrow and Negro Creek’s configuration.

Figure 4.6: The radar station at Negro Creek, ON.
Chapter 5

Method and Techniques

This Chapter is dedicated to explaining our experiment and the problems that had to be solved to conduct this experiment. Doppler radars are used to measure vertical and horizontal wind’s velocity. However, here it is explained how we can understand the correlation between vertical wind and horizontal wind and figure out how much of this correlation results from tilted layers or the scatterers inside them and how much of it is because of the tilt of nominally vertical beam.

5.1 Experimental Studies’ Description

We will assume that we have a Cartesian coordinate above the ground in which the $x − y$ plane represents the horizontal plane, and the $z − axis$ represents the vertical direction. If we determine the vertical wind average over a month, the vertical velocity of wind should then be close to zero, as it expected downward wind and upward wind in be produced homogeneously but this mean can be non-zero for various reasons like gravity waves or wind shears [69].

Furthermore, if the turbulent scatterers are isotropic, and if the vertical beam is truly vertical, and if all of the motions of the scatterers are due to turbulence carried along by a horizontal mean wind speed, then cross-correlating the hourly mean velocities measured with the vertical beam against the horizontal hourly-mean winds measured with the off-vertical beam should
produce zero correlation coefficient.

In the following discussion, we will measure winds with vertical and off-vertical radar beams and cross-correlate the hourly mean vertical and horizontal winds (as it is the suitable mean for measuring the desired changes and ideal for forecasting use). We will refer to the values measured on the vertical beam as ”nominally vertical” winds since we will at times be considering the possibility that the vertical beam is not truly vertical. We can just use this method for the determination of nominally vertical beam’s tilt, because in the accuracy of a fraction of a degree, this is the only way and it can not be determined by other means or due to non moving objects. There are various reasons why these two sets of motions might be correlated, and the purpose of this thesis is to ascertain what types of scenarios might lead to such a correlation and determine which (if any) is most applicable to our radars. In the paragraphs below, we will discuss possible scenarios that might lead to non-zero correlations.

The first possibility will be to consider whether the nominally vertical beam which is used to measure the vertical wind $W_i$ is truly vertical. If it is not vertical (e.g. with a tilt error of say $0.10\degree$ to $0.50\degree$), then the measured vertical velocity will contain a horizontal component, which will bias the vertical wind’s mean. Figure 5.1 demonstrates this assumption.

Some other possibilities might also account for non-zero correlations, which are discussed below.

A second possibility is that the vertical beam is truly vertical, as it had been designed, and the scatterers layer itself is tilted from horizontal, or that individual scatterers are anisotropic and tilted within the layers. Therefore, the nominal vertical wind would contain a component of the horizontal wind’s component. In contrast to the first proposal, which is instrumental, if this second case were right, it could teach us essential geophysical information about the physics of the scattering layers.

Another possibility is that the motions of air, above the radar, are not dominated exclusively by turbulence and contain gravity waves. As discussed in chapter 3, the horizontal and vertical velocities in a gravity wave are correlated by nature. Again, this would be a geophysical reason.
A fourth possibility is that we face a combination of some or all of the issues discussed above. In such a case, we have a combination of instrumental and geophysical effects.

For example, if we consider combining the first and second possibilities, both the scattering layer and the vertical beam may be slightly off from their expected plane. In this case, again, a component of horizontal wind is recognizable and but if this is the case, we should investigate how the effect of each phenomenon is. It is schematically shown in Figure 5.2.

There are some distinct differences between these scenarios, which we can use to discriminate these models. For example, if the effect were purely due to the instrumental effects, it would be independent of height and month. Other geophysical effects may show temporal and height-dependent changes.

In the next section, we will begin to present some experimental results.
Figure 5.1: An illustration that shows how the vertical wind will contain a horizontal component if the vertical beam is not truly vertical, and it is off-vertical even though the scatterers are in a horizontal plane.
Figure 5.2: The yellow scatterers are tilted in this schematic figure. The pink beam shows a real vertical beam, and the blue beam shows the tilted beam from vertical.
5.2 The Experiment

To investigate these assumptions, we assumed that the nominally vertical radar beam had a tilt from vertical of $\theta$. Assuming a horizontal wind speed of $U$, and assuming that the wind had a direction $\phi_0$ relative to north, then we can determine the nominal vertical wind which we would expect the radar to measure on the nominally vertical beam through the relation:

$$W_u = U \sin \theta \cos(\Phi + 180 - \phi_0) \quad (5.1)$$

where $U$ is the magnitude of the horizontal wind, $\Phi$ is the angle from which the wind blows, and $\phi_0$ is the azimuthal angle of the beam relative to north. We then can cross-correlate the measured vertical wind ($w$) and horizontal wind ($U$) (or being more specific the correlation between $w$ and $U \sin \theta \cos(\Phi + 180 - \phi_0) = W_u$) and see if they are at all correlated.

Both the nominal vertical wind ($w$) and the horizontal wind ($U$) are measured by Doppler radars. We use off-vertical beams for measuring horizontal wind and nominally vertical beams for measuring the nominal vertical velocity. The off-vertical beam is illustrated in Figure 5.3.

Afterward, we then plotted the least-square best-fit line between the measured hourly vertical wind and the hourly horizontal wind in the direction of $\phi$ for each month. For each month and height, we changed $\Phi_0$ from 0° to 360° degrees to maximize $R^2$ (Pearson Correlation Coefficient), then find out at what azimuthal angle the correlation is maximum, and then finally found the zenithal direction of the tilt. Pearson correlation coefficient is defined as:

$$R^2 = \frac{\sum_{i=1}^{n}(w_i - \bar{w})(W_{u_i} - \bar{W}_u)}{\sqrt{\sum_{i=1}^{n}(w_i - \bar{w})^2} \sqrt{\sum_{i=1}^{n}(W_{u_i} - \bar{W}_u)^2}} \quad (5.2)$$

Figure 5.4 shows one of the example plots that fits the best fit line to the vertical ($W$) and horizontal wind ($U \sin \theta \cos(\Phi + 180 - \phi_0)$). This figure shows a correlation with $R^2 = 0.4$ at $\phi_0 = 50^\circ$. On the other hand, Figure 5.5 shows one of the examples of plots that can not show any correlation between the vertical ($W$) and horizontal ($U \sin \theta \cos(\Phi + 180 - \phi_0)$) wind. Here, not having a correlation happens at $\phi_0 = 5^\circ$. Not having a correlation is not a bad thing,
surprisingly having correlation in some $\phi_0$ and not having it in other ones is in support of our assumption. It shows the correlation is dependent on the direction of $\phi_0$, meaning there is not a constant correlation everywhere or every time which leads to the fact that this correlation is not just because of tilt of the beam which is an instrumental cause and is constant everywhere as all the radar’s setup’s are similar.

In my initial calculations, we assumed that $\theta$ was 1.0°. Then the slope of the scatter plot at the optimum azimuthal angle tells us the real value of $\theta$ (in degrees (or more specifically the slope is $\sin \theta$, but at our small tilt angles it is suitable to use simply $\theta$ : the slope measured is just the ratio of $\sin \theta$ to $\sin(1°)$)

1. We have plotted the scatterer plots for heights from 0.5 to 13 km and different azimuthal directions $\phi_0$ from 0° to 360° for different sites. These figure shows the correlation plot as density plots for the Negrocreek site for May of 2009. Here it can be seen that there is a strong dependence on height. Here we chose $\theta = 1°$. As we can see here, the
5.2. The Experiment

Figure 5.4: Scatter Plot of Vertical wind Vs. Horizontal wind, with $\theta = 10^\circ$ and $\phi_0 = 50^\circ$ for Jan 2010, Walsingham

maximum correlation at height 10 km, happens at $\phi_0 = 100^\circ$, while for altitude 5 km, the max correlation happens around $320^\circ$.

We should point out the important phenomenon in these plots is that we see a red pattern, which is the maximums, but we see a similar blue pattern, which is the minimums. This blue pattern is not giving us any new information; because the angles used here are the same as the red part but from backward: $360 - \phi_0$. This makes sense since all we have done is reverse the mean wind direction when we rotate it by $180^\circ$, so we expect the opposite sign in the correlation. In other words, because we were using a sinusoidal equation, we had to expect this maximum and minimum.
2. We repeated the scatter plots as the first figure, but we considered the slopes of best fit lines between the vertical wind ($w$) and horizontal wind ($W_u$ or this time). Again, we chose $\theta = 1^\circ$ Here we can see again for Negro Creek in May of 2009 in the altitude $10\ km$, where the $\phi_0 = 100^\circ$, which the angle where the correlation is maximum, the slope is about one and for height $5\ km$, in the phase where the correlation is optimum ($\phi_0 = 320^\circ$), the slope is about $0.6^\circ$ Here it shows the absolute slope is $1 \times 0.6^\circ = 0.6^\circ$

3. Similar studies for other times of the year represent different values for $\phi_0$ for example here in Figure 5.8, in June of the same year, again for Negro Creek, for chosen $\theta = 1^\circ$, we can see the maximum correlation at height $10\ km$, happens at $\phi_0 = 120^\circ$ while for
5.2. The Experiment

Figure 5.6: This plot shows correlation for each height by variation of phi ($\phi_0$) for Negrocreek, May, 2009.

Altitude 5km, the max correlation happens around 60°.

4. We repeated the scatterer plots as the first figure, but we considered the slopes between vertical wind ($W$) and horizontal wind ($W_u$ or $U \sin \theta \cos(\Phi + 180 - \phi_0)$) this time. Again, we chose $\theta = 1^\circ$. Here we can see again for Negrocreek in June of 2009 in the altitude 10 km, where the $\phi_0 = 120^\circ$, which the angle where the correlation is maximum, the slope is about one and for height 5 km, in the phase where the correlation is optimum ($\phi_0 = 320^\circ$), the slope is about one again. Here it shows the real value of $\theta$ is:

$$1.0 \times 1^\circ = 1.0^\circ$$

There is an essential point that in almost all correlation plots, the correlation from 0 to 5 km is weak. This is true for almost all of the sites and all the times, indicating that the
beam is truly vertical at these heights. These results suggest that the effect is not due to the radar but to the geophysical effects. This is a region of high-quality data. Therefore, the result can not be interpreted as a poor data consequence. It seems most likely that the beam is truly vertical, and the effects at other heights are geophysical. (It can also be probable that the effects that we observe above 5 km are due to the tilt of the beam, but here, below 5 km, we have some geophysical effects that null the beam’s effect, but this is a less likely situation). Comparing Figure 5.6 and Figure 5.8, we can determine not only that the trend of correlations and the angle where the optimum \( \phi_0 \) happens for each height is not the same for different months of the same year and the same sites, but also it is highly dependent on the time and season as they are plots for the same site and year but for two different months. In the next chapter, these variables are addressed extensively

Figure 5.7: This plot shows slope for each height by variation of phi (\( \phi_0 \)) for Negrocreek, May, 2009
Figure 5.8: This plot shows correlation for each height by variation of phi ($\phi_0$) for Negrocreek, Jun, 2009

and here all the plots are just represented for comparison between each other.
Figure 5.9: This plot shows correlation for each height by variation of phi ($\phi_0$) for Negocreek, Jun, 2009.
Chapter 6

Results

In this chapter, I present the results of our studies. Many variables like location, season, or height are investigated in this experiment to understand each of their contributions. Therefore, there are many diagrams and statistical analyses, so I wish to make some comments here.

While in a regular study like this, it is standard to seek out the means of various parameters and compare them. I will do that regarding correlations, azimuthal orientations, and zenithal tilts.

But a word of warning is needed. Since we are especially interested in gravity waves, there may often be no preferred “mean value”, especially in the presence of time-varying tilts. Tilts will change on scales of minutes and hours. In such circumstances, the mean value is of little importance, and the key parameter will be the variation of tilts - this will be a better measure of wave activity. More significant wave activity will lead to a larger variation in tilt angles. So towards the end of the chapter, we will start to focus not on the mean values but rather the standard deviations as our primary parameter for looking at wave activity. The errors will be found using the standard deviations of the standard deviations or as being called as "spread of standard deviations."

We, therefore, wish to forewarn the reader of this unusual but valid approach so that it does not catch them unaware.
6.1 Maximum Correlation and Corresponding Phi’s (Azimuth) and Slopes (Zenithal angles)

After we found out there are correlations between horizontal and vertical wind, we found the maximum correlation between the horizontal wind and vertical wind at each height sequentially by changing $\phi_0$ for all the 6 sites from 0° to 360°. Once we had found the azimuth at which the correlation was maximum, then I calculated the slope of vertical wind vs. horizontal wind at that azimuth angle. As already mentioned, this relates to the tilt angle (Zenith): $\theta$.

Some figures are selectively chosen for the goal of comparisons and pointing out the differences between sites or seasons or etc., therefore we are more interested in the big picture rather than addressing all the details like outliers in every single plot. For instance, Figure 6.1 shows how the maximum correlation behaves for different heights.

Figure 6.2 represents the behavior of azimuths for each month for all the altitudes with resolution the of 0.5 km at which correlation is maximum.

Figure 6.3 shows how the slope changes over heights at phase angles that gave us the maximum correlations for each month of the year 2010 for site Harrow. Other sites are also analyzed, but only Harrow-2010 is shown as a sample here.
6.1. Maximum Correlation and Corresponding Phi’s (Azimuth) and Slopes (Zenithal angles)

Figure 6.1: Maximum Correlations between vertical wind ($w$) and horizontal wind ($W_u$)
Figure 6.2: Azimuths where correlation between vertical wind ($w$) and horizontal wind ($W_u$)
6.1. **Maximum Correlation and Corresponding Phi’s (Azimuth) and Slopes (Zenithal angles)**

Figure 6.3: Zenithal angles where correlation between vertical wind ($w$) and horizontal wind ($W_u$) is maximum
6.2 Seasonal Moving Averaged Correlation

Next, we calculated the moving average over height for all the months of each year-site in order to the better observe the patterns of behaviour as a function of height and time. We calculated the moving average of the maximum correlation value with a window of 9 (nine heights in each average) as it makes the plots smoother and cleaner, which means in every plot, we have 18 points for each month. Some of the plots are shown here for the sake of comparison. We repeated the same sequence for phi’s (azimuths) where correlation maximizes and for the value of the slope at where the correlation is maximum. The approach of moving average was chosen to smooth the graphs and make them more suitable for comparisons.

However, we also separated the seasons to investigate the effect of seasons in selected years. Moving averages of maximum correlation for different heights in winters (December, January, February) are plotted in red, in springs (March, April, May) are plotted in yellow, in summers (June, July, August) are plotted in green, and for falls (September, October, November) are plotted in blue.

As can be seen in Figure 6.4, the pattern for the correlation for all the months is different, and it is also different among various sites. We can also find a similar pattern in all of them. Nominally as the height increases, the correlation increases, and in the higher heights, it lies somewhere between 0.4 and 0.6.

The noticeable fact here is that the maximum correlation pattern is different from site to site, but it is similar within the sites with slight differences for different years. The fact that where this maximum correlation is happening (the azimuth angle) for each site-year, for all heights is calculated and compared in the next section.

Also, the different behaviour of the maximum correlation for different seasons should not be neglected.
Figure 6.4: Seasonal moving averages over height with window of 9-Correlation-Negro Creek- Typical error bars are ±0.02
Figure 6.5: Seasonal moving averages over height with window of 9-Correlation-Wilberforce- Typical error bars are $\pm 0.02$

Figure 6.6: Seasonal moving averages over height with window of 9-Correlation-Eureka- Typical error bars are $\pm 0.02$
6.3 Seasonal Phis (Azimuth)

Here again, we repeated the same procedure as above for phi’s but this time our parameter of interest is the value of phi (azimuth) at which the correlation maximizes, rather than the value of the correlation itself. The moving average of phase at which the maximum correlation happens, over heights, with a window of 9 is colour-coded based on seasons.

For all plots, we can find that for different months, the optimum phi’s are different. However, they revolve around a specific angle. For instance for Negrocreek (Figure 6.7), the optimum phase angle is somewhere between 50° to 120° but mostly around 90°, on the other hand for Wilberforce (Figure 6.8), the optimum phase angle lies between 80° to 160°, more inclined to 120°. Again, like the correlation plots, these maximums get closer to each other in the upper part of the troposphere. For Wilberforce, the azimuthal rotation of Doppler Beams relative to the north is 22.5°. If we subtract it from the phase angle of maximum correlation (120°), we find a difference of 97.5°.

For Negrocreek, the azimuthal rotation of Doppler Beams relative to the north is −27° degrees, which means if we subtract it from the azimuth where the correlation is maximum at (90°) it gives 117° as our absolute optimum phase.

This comparison shows the optimum phase angles are different for different sites, suggesting a difference in every individual site’s situation as both radars’ setups are the same. If the observed effect was due to the setup, then we should have witnessed a similar offset relative to the beam azimuth everywhere, which is not the case here.

Again by looking at colors, it can be said months considered in the same season have similar behaviors and the phase angles in the same-season months are relatively closer to each other than non-same-season months.
Figure 6.7: Seasonal moving averages over height with window of 9-Phi-Negrocreek
6.3. Seasonal Phis (Azimuth)

Figure 6.8: Seasonal moving averages over height with window of 9-Phi-Wilberforce

Figure 6.9: Seasonal moving averages over height with window of 9-Phi-Eureka
6.4 Table of Azimuths (Phi’s) Where the Maximum Correlation Happens

<table>
<thead>
<tr>
<th>Site</th>
<th>Year</th>
<th>Set up Phi</th>
<th>Optimum Phi</th>
<th>Absolute Optimum Phi</th>
</tr>
</thead>
<tbody>
<tr>
<td>Wilberforce</td>
<td>2009</td>
<td>22.5°</td>
<td>140°</td>
<td>117.5°</td>
</tr>
<tr>
<td>Wilberforce</td>
<td>2010</td>
<td>22.5°</td>
<td>120°</td>
<td>97.5°</td>
</tr>
<tr>
<td>Walsingham</td>
<td>2009</td>
<td>−14°</td>
<td>70°</td>
<td>84°</td>
</tr>
<tr>
<td>Walsingham</td>
<td>2010</td>
<td>−14°</td>
<td>80°</td>
<td>94°</td>
</tr>
<tr>
<td>Walsingham</td>
<td>2011</td>
<td>−14°</td>
<td>60°</td>
<td>74°</td>
</tr>
<tr>
<td>Walsingham</td>
<td>2012</td>
<td>−14°</td>
<td>45°</td>
<td>59°</td>
</tr>
<tr>
<td>McGill</td>
<td>2009</td>
<td>−48.7°</td>
<td>110°</td>
<td>158°</td>
</tr>
<tr>
<td>McGill</td>
<td>2010</td>
<td>−48.7°</td>
<td>130°</td>
<td>178°</td>
</tr>
<tr>
<td>Negro creek</td>
<td>2009</td>
<td>−27.0°</td>
<td>95°</td>
<td>122°</td>
</tr>
<tr>
<td>Negro creek</td>
<td>2010</td>
<td>−27.0°</td>
<td>90°</td>
<td>117°</td>
</tr>
<tr>
<td>Harrow</td>
<td>2009</td>
<td>−41.0°</td>
<td>50°</td>
<td>91°</td>
</tr>
<tr>
<td>Harrow</td>
<td>2010</td>
<td>−41.0°</td>
<td>60°</td>
<td>101°</td>
</tr>
<tr>
<td>Eureka</td>
<td>2009</td>
<td>−30.0°</td>
<td>100°</td>
<td>130°</td>
</tr>
<tr>
<td>Eureka</td>
<td>2010</td>
<td>−30.0°</td>
<td>80°</td>
<td>110°</td>
</tr>
</tbody>
</table>

Table 6.1: Absolute Optimum azimuths

Table 6.1 shows that every site and year, the absolute phase angle that on average the maximum correlation happens for each site and year is different, which is strong evidence that this correlation is NOT just due to the set up of the radars, but also to the tilt of the layers or both tilts.
6.5 Seasonal Slopes (Zenithal Angular Offsets)

Here in Figure 6.10, the slope at phase angles where the correlation is maximum, is somewhere between $0.3^\circ \times 1 = 0.3^\circ$ to $1^\circ \times 1 = 1^\circ$. It is noticeable that for this site (Negrocreek), the slope is between $0.3^\circ$ to $1^\circ$ which is entirely different from the slope for Wilberforce in the same year (2010), $0.5^\circ$ to $1.3^\circ$ (Figure 6.11). As all the radars’ setup is the same and they had been set up with the same angle from vertical (actually tried to be vertical), it is logical to assume that we are looking at the tilt of layers in these plots or a combination of tilt of the layers and the tilt of nominally vertical beam.

Obviously, this tilt is different for other sites and years. But almost all of them have this repetitive pattern that there is a change in slope in the middle altitudes, still this altitude is different for each site; It is also noteworthy to mention if the tilt were solely due to an error in the pointing of the vertical beam, then all tilts would be identical at all heights. It can also be recognized that these patterns have slightly different forms for each year for the same sites. These patterns in the change of slopes improve the idea that they are mostly due to atmospheric phenomena rather than the contribution of radar.

The last important issue that catches our eyes is how to spread out the angles over months for each year-site. A quick comparison of Eureka with Negrocreek shows that the minimum tilt for Eureka (2009) is $0.6^\circ$ (Figure 6.12), and the maximum is $1.6^\circ$. On the other hand, the minimum slope for Negrocreek (2010) is $0.3^\circ$, and the maximum is $1^\circ$ as mentioned above; hence it is noticeable that the difference between the maximum and minimum slope for Eureka is $1^\circ$, but for Negrocreek, it is $0.7^\circ$. This difference in various sites and years led us to calculate the running standard deviation of correlations, phi’s and slopes from month to month and also from season to season to investigate the effect of seasons on the standard deviations and how seasons impact the correlations, phi’s (azimuths) and slopes (zeniths). Therefore in the next step, the seasonal effects are addressed.
Figure 6.10: Seasonal moving averages of over height with window of 9-Slope-Negrocreek
6.5. Seasonal Slopes (Zenithal Angular Offsets)

Figure 6.11: Seasonal moving averages of slope over height with window of 9-Slope-Wilberforce

Figure 6.12: Seasonal moving averages of over height with window of 9-Slope-Eureka
6.6 Monthly Running Means for Correlations

![Moving Averaged Height Vs.Moving averaged correlation](image)

Figure 6.13: Monthly Moving Averages with Window of 3 × 3-Correlation-Negrocreek

In this step, the moving average of correlation with a window of 3×3 is calculated, meaning the average over 3 months and over 3 month was calculated, then moved one step forward and the average for the next 3 months and next 3 heights and so on were calculated. A correlation of about 0.5 is happening around 6 km for almost all the months and all the years and sites which can suggest greater tilt in layers in this height. It should be noted that some months have more variances than others, which is addressed with more details in the standard deviation’s section. It is also clear that there are similar patterns inside every site, which is different from other sites.

This approach also makes the plots smoother both over months and altitudes. The vital
6.6. Monthly Running Means for Correlations

Figure 6.14: Monthly Moving Averages with Window of 3×3-Correlation-Walsingham

Figure 6.15: Monthly Moving Averages with Window of 3×3-Correlation-Eureka
issue that should be pointed out here is that the altitudes here goes up to 12.5 km meaning it started from 0.5 km then goes upward with the resolution of 0.5 km, but in the previous sections, the altitudes were going to 26 just because they were multiplied by 2, it looked like they were like points rather than altitudes, but the truth is they should be divided by 2 for understanding the exact height in km.

6.7 Monthly Running Means for Phi’s

Here the moving average of phi’s that the correlation is maximum at are observed. For example, in Figure 6.23, the optimum phi is 60° to 80° for Negro creek in 2010 but for Walsingham in 2011 (Figure 6.26), the optimum phi is around 80° to 100° and for Wilberforce in
2010 (Figure 6.27), the optimum correlation happens around 130°. Also, Walsingham encounters lots of variations around the average of averaged phi’s, which enhances this theory that there are many variations in this area because of its location, close to the sea, and volatile weather.
Figure 6.17: Monthly Moving Averages with Window of 3 × 3-Phi’s-Walsingham

Figure 6.18: Monthly Moving Averages with Window of 3 × 3-Phi’s-Wilberforce
Here again, the moving averages for 3 months and 3 heights were calculated, and the number beside them in the plot represents the middle month of each three-month pack. As monthly averaging made the graphs smoother, it is clear from all the plots that slopes changes as moving upward and the range of value of them is different for each site-year and there are more tilts in some height in comparison with other heights.
Figure 6.20: Monthly Moving Averages with Window of $3 \times 3$ -Slope-Walsingham

Figure 6.21: Monthly Moving Averages with Window of $3 \times 3$ -Slope-Eureka
6.9 Running Standard Deviations for Correlations

In the previous section, the spread of the monthly moving averages caught our eye. It needs to be recognized that mean values of slopes and azimuthal directions might not be the best parameter to examine since the weather can vary. They might make sense if the site is close to a lake or mountain so that specific wind flows are ubiquitous (e.g. lake breezes). However, at other sites, where there are no systematic flows, it might be more prudent to look not at the mean values but more at the variability. For example, if waves are important, we might expect enhanced variability but no systematic mean values of azimuth or zenithal angles like it is illustrated in the Figure 6.22. Therefore, in this section, the monthly moving standard deviations are addressed. As similar as above, again, monthly standard deviations with a window of $3 \times 3$ are calculated simultaneously with calculating moving standard deviations over the heights with a window of 3 and over the months with window of 3 and plotted with different colours. Again, each number beside the colours represents the middle month of the averaging window, and the numbers on the height axis show the middle height in the averaging window.

Moving standard deviation in all cases is relatively small for the correlations, but it seems some months are more variant than others.
Figure 6.22: An illustration of movement of typical scatterer and its propagation over the beam
Figure 6.23: Moving standard deviations with window of $3 \times 3$ over months and heights-Correlation-Negro Creek
Figure 6.24: Moving standard deviations with window of $3 \times 3$ over months and heights-Correlation-Walsingham

Figure 6.25: Moving standard deviations with window of $3 \times 3$ over months and heights-Correlation-Eureka
6.10 Running Standard Deviation’s Table for Correlations

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<th>Site-Year</th>
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<tbody>
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<tr>
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<tr>
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Table 6.2: Running Standard Deviations’ Table for correlations with window of $3 \times 3$

6.11 Running Standard Deviations for Phi’s

The standard deviation for phi’s are significant relatively, as in some months it goes up to $90^\circ$, but overall they are mostly around $0^\circ$ to $10^\circ$. These significant standard deviations are more recognizable in the window for which January is the middle month (December- January- February).
Figure 6.26: Moving standard deviations with window of $3 \times 3$ over months and heights-Phi’s-Negro creek

Moving Averaged Height Vs. Moving Standard Deviation of Phi

- 201001
- 201002
- 201003
- 201004
- 201005
- 201006
- 201007
- 201008
- 201009
- 201010
- 201011

Averaged Height (km)

Moving Std of Phi ($^\circ$)

Negro creek 2010
Figure 6.27: Moving standard deviations with window of $3 \times 3$ over months and heights-Phi-Walsingham

Figure 6.28: Moving standard deviations with window of $3 \times 3$ over months and heights-Phi-Eureka
6.12 Running Standard Deviation’s Table for Phi’s

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Table 6.3: Running Standard Deviation’s Table for Phi’s with window of $3 \times 3$

6.13 Running Standard Deviations for Slopes

The moving standard deviations of slopes shows a similar pattern as the pattern of moving standard deviations for correlation and phi’s, so that some months are more variant than the others; looking at Eureka, 2009 (Figure 6.31) is the most variant one overall for each month.
Figure 6.29: Running Standard Deviations with window of $3 \times 3$ - Slope-Negrocreek
Chapter 6. Results

Figure 6.30: Running Standard Deviations with window of $3 \times 3$ -Slope-Walsingham

Figure 6.31: Running Standard Deviations with window of $3 \times 3$ -Slope-Eureka
For the purpose of comparison, we made a solid number for each year and site to compare. We squared all the calculated standard deviations for a site-year, sum them up and divided them by their count, finally took the square root and reported the value in the table. The interesting fact here is that the slope’s standard deviation for Walsingham and Harrow is more significant than other sites. This result concludes that lake breeze can be an influential factor in producing these deviations as these sites are close to Lake Erie. Another interesting fact is that we cannot say anything specifically about McGill as it is good evidence that this city has variant weather.
6.15 The Spread of Standard Deviation of slopes

In Table 6.4, the spread of standard deviations is shown. For 7 Site-Year out of 14 Site-year, the spread of standard deviation is significantly greater in November than June. This suggests that these standard deviations are originated because of seasons. As expected, June is a calm month in summer, and November is one of the least stable months in winter, so the spread of standard deviation in a winter monthly average is more significant than a summer monthly average. It should be remembered that November is representative of averages of October, November, and December. There is not enough information left for the other 7 Site-Years; data for either June or November is missing. Besides, we cannot conclude anything about some of them like Negrocrreek-2010. Again, it can be acknowledged here that McGill is one of the sites that makes it harder to predict its pattern.
Table 6.5: Standard Deviation’s Spread for Wilberforce

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Table 6.8: Standard Deviation’s Spread for Walsingham and Eureka

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6.15. The Spread of Standard Deviation of slopes

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(b) McGill-2010

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<tr>
<td>201003</td>
<td>0.083</td>
</tr>
<tr>
<td>201004</td>
<td>0.126</td>
</tr>
<tr>
<td>201005</td>
<td>0.108</td>
</tr>
<tr>
<td>201006</td>
<td>0.075</td>
</tr>
<tr>
<td>201007</td>
<td>0.068</td>
</tr>
<tr>
<td>201008</td>
<td>0.128</td>
</tr>
<tr>
<td>201009</td>
<td>0.101</td>
</tr>
<tr>
<td>201010</td>
<td>0.083</td>
</tr>
<tr>
<td>201011</td>
<td>0.141</td>
</tr>
<tr>
<td>201012</td>
<td>0.248</td>
</tr>
</tbody>
</table>
6.16 Analysis of Variance (ANOVA)

The one-way analysis of variance (ANOVA) is used to determine whether there are any statistically significant differences between the means of three or more independent (unrelated) groups or not [70].

Here one-way ANOVA is used to make sure these variances which are observed are not due to noise or other random phenomena. ANOVA compares the variance between individual groups to the variance inside of each sample, because there might be some vast variances between sample groups but if also each sample group is variant inside themselves, this variance is not giving us any information.

The analysis of variance is conducted in Python version 3.7 using `scipy.stats.f_oneway` for slopes. `f_oneway` returns two values, first the F score and then P-value. F value calculates the variance of the group means (Mean Square Between) / mean of the within-group variances (Mean Squared Error)

1. ANOVA for Slopes, 2009:

First, the ANOVA was performed on all the slopes for all the 6 sites and the year 2009.

**F Score** = 66.24

**P value** = $6.33 \times 10^{-60}$

As the degree of freedom is more than 130, the F-score is far beyond the critical value for any $\alpha$.

2. ANOVA for Slope, 2010:

Results of ANOVA for 6 sites for year 2010

**F Score** = 63.03

**P value** = $1.41 \times 10^{-60}$

Again, the F-Score and P-value are far beyond the critical value.
3. ANOVA for Std of slopes, 2009:

ANOVA for the standard deviation of slopes:

In this step, ANOVA was performed on the moving standard deviations of slopes for our 6 sites for 2 years.

F Score = 4.77

P value = 0.0004

4. ANOVA for Std of slopes, 2010:

F Score = 4.98

P value = 0.0003

As the degree of freedom is more than 20 for both analyses, any F Score greater than 2.9 (the critical value) is good, which is right in our cases.

5. ANOVA for the spread of Std of slopes, 2009: Also, the P-value is much less than 0.05 which confirms that with high probability, the alternative hypothesis is true, meaning these standard deviations are coming from different samples, and the Null Hypothesis can be rejected with high probability.

6. ANOVA for the spread of Std of slopes, 2009:

In this step, ANOVA was performed on the spread of standard deviations (Standard deviation of standard deviations), with about 10 – 12 points in each array I have:

F Scores = 7.24

P value = 0.00010

7. ANOVA for the spread of Std of slopes, 2010:

F Score = 9.49

P value = $3.74 \times 10^{-6}$
As the degree of freedom is more than 10 for all, any f score greater than 4.5 (the critical value) can reject the null hypothesis, which is met again.

8. ANOVA for Walsingham:

The ANOVA for Walsingham for 4 years was calculated since it is the only site that its 4 years data for is analysed. The goal of this procedure was to investigate if the data of different years for one site are different or not, as the result can talk about the time effects.

F Score = 3.45

P value = 0.019

It slightly passes the F test and p-value test. They are kind of from the same sample and somehow not from the same sample!

In overall, these result shows that the seasonal/annual changes at one site are modest, but site-to-site variations are huge, which is what was expected.

The results of the analysis of variance for slopes, standard deviation of slopes and the spread of standard deviation of slopes and slopes for different years of Walsingham are summarized in the next section’s table to make it easier for comparison.

6.17 Summary of ANOVA’s Result

<table>
<thead>
<tr>
<th>Source of Variance</th>
<th>Year</th>
<th>F Score</th>
<th>P value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Slopes</td>
<td>2009</td>
<td>66.24</td>
<td>6.33 × 10⁻⁶⁰</td>
</tr>
<tr>
<td>Slopes</td>
<td>2010</td>
<td>63.03</td>
<td>1.41 × 10⁻⁶⁰</td>
</tr>
<tr>
<td>Standard Deviation of slope</td>
<td>2009</td>
<td>4.77</td>
<td>4.55 × 10⁻⁴</td>
</tr>
<tr>
<td>Standard Deviation of slope</td>
<td>2010</td>
<td>4.98</td>
<td>3.04 × 10⁻⁴</td>
</tr>
<tr>
<td>Spread of Standard Deviation</td>
<td>2009</td>
<td>7.24</td>
<td>1.04 × 10⁻⁴</td>
</tr>
</tbody>
</table>
### 6.17. Summary of ANOVA’s Result

<table>
<thead>
<tr>
<th>Spread of Standard Deviation of slope of Slope</th>
<th>2010</th>
<th>9.49</th>
<th>$3.74 \times 10^{-6}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Slopes for Walsingham</td>
<td>2009 – 2010 – 2011 – 2012</td>
<td>3.45</td>
<td>$1.95 \times 10^{-2}$</td>
</tr>
</tbody>
</table>

Table 6.12: ANOVA
Chapter 7

Conclusion

Overall, in this research, the correlation between horizontal wind and vertical wind in the troposphere with Doppler radars, in 6 sites, was investigated. Variables such as location, month, season, height, etc., were explored to better understand this correlation’s nature. This correlation is tested to understand if it is due to the tilt of atmospheric scatterers, which itself is proof of the existence of gravity waves or tilt in the nominally vertical beam.

Here is the conclusion of our research:

1. There were significant correlation between the horizontal and vertical wind for all the sites and months but this correlation was variant between 0 to 0.7. As it has been addressed in the Result chapter we showed that the azimuth which maximum correlations occurs is different for different sites in different months and years, specially considering the fact that all the radars have been set up with variant azimuthal orientation. Also the values of the slope at azimuths, where the correlation maximized was shown to be variant from site to site. This means this correlation is not just based on the radar’s design or just because the vertical beam of the radar is slightly off vertical. We conclude the scatterers themselves are tilted and if the nominally vertical beam is tilted, its tilt is very small (fraction of a degree) and tilt of scatterers have the major contribution in our observation.
2. The slopes where the azimuthal angle produces maximum correlation are typically between 0.6° to 1.6° which also means the angle (θ), that have been addressed thoroughly in this thesis, is typically between 0.6° to 1.6°.

3. The running standard deviation of slopes for Walsingham and Harrow and Eureka sites had significantly greater standard deviation. Since these sites are close to lakes, they experience frequent lake breezes which may also produce gravity waves. While lake-breezes occur below heights of 1 – 2 km, they can propagate more into the troposphere and reflect back which can make standing waves, thus they can be responsible for tilt of layers [71]. The fact Walsingham and Harrow are both on the lake shores, empowers this idea since there is a north–south wind breeze in both sites, below 2 km altitude.

4. Sea breeze is a mesoscale wind (2 to 2000 km) which is the origin of local circulation in coastal areas. In our case, with sites close to the Great Lakes, lake-breezes behave similarly. These breezes happen as the solar radiation makes a huge temperature difference between land and sea, thus it makes a pressured gradient force toward the land. Cool air above the sea moves toward the land and the air above the land lifts vertically and it may make a closed circuit which have a return descending flow toward the lake at distances of tens of kilometers. There are several means for measuring the wind velocity that is generated by sea breeze but they are all dependent on calculating the temperature gradient. Understanding sea breeze is one of the hardest jobs in forecasting as many factors have an impact and changes can occur fast as an hour. Also, hence they are responsible for causing mixing between the marine air and land air and making turbulence, they are really interesting phenomena to study. This work helps meteorologist to have a better estimation when sea breeze can happen, and if it happens, how strong it is and in which direction and finally far in can spread [71].

5. Eureka, is the only arctic site which is very close to sea-water but the sea is frozen for the most of year. These special characteristics of Eureka also made lots of our data
usable but the one usable year of data showed high variance in slopes, which again can be interpreted at this location, sea breeze affected our data. Also the mountains in the north-east side of the site increases the chance of production of more gravity waves in this region.

6. McGill is one the most interesting sites which suffer from different phenomena so it shows both high and low standard deviations for different years, thus it is more complicated to study this site.

7. Wilberforce and Negrocreek have more minor standard deviations as they are not close to any lakes or mountains, suggesting this low level of variety in the slopes is because gravity waves are not generated in these areas. Since without nearby lake effects and mountain effects, gravity-wave production is weaker.

8. In conclusion, the spread of standard devotions for summer months (e.g. June) is smaller than the spread of standard deviation for winter months (e.g. November). This was exactly what was expected if it was assumed the tilt is a result of gravity waves. The troposphere is more turbulent in winters and there are more chances that this turbulence is induced by gravity waves.

9. The Analysis of variance verifies that all of these sites are different from each other and the observed patterns are not due to noise or other random factors. Also analysis of variance of Walsingham for multiple years showed exactly the results we expected; i.e the slopes were different from year to year, but the differences were not as huge as differences between sites.

10. Finally; the scatterers tilts are a strong evidence of existence of gravity waves in the Troposphere, at the same time suggesting the probable tilt of radar is small [69].

11. Receiving backscattered signals from the scatterers is a strong proof that these scatterers are not isotopic, because if they were isotropic the average received backscattered signal
would be zero as isotropic scatterers, backscatter the signal homogeneously [72]. Also it should be emphasised that anisotropic scatterer is a product of gravity waves and these is another evidence for existence of gravity waves in the troposphere.

12. A future recommendation for further work on this project could be to perform numerical modeling to understand how much of this tilt is a contribution of tilt of the scatterers and how much of it is a tilt of the nominally vertical beam.
Chapter 8

Appendix

8.1 Python Scripts-Correlation

```python
import numpy as np
import math as mt
import pylab
import matplotlib
import matplotlib.cm as cm
import matplotlib.mlab as mlab
import matplotlib.pyplot as plt
from mpl_toolkits.mplot3d import Axes3D
import matplotlib.ticker as ticker
import scipy.interpolate as si
import pandas as pd

matplotlib.rcParams['xtick.direction'] = 'out'
```


matplotlib.rcParams['ytick.direction'] = 'out'

##### Time array #####
# xat = np.loadtxt('e:/Data/re200805.walsingham/longdatat.dat', delimiter = ',', unpack=True)
# print(xat)
# read # rows and columns from rowcol array - used for # all files EXCEPT the time file, which has 5 columns

nc = np.loadtxt(r'C:\Users\Farnoush\Desktop\Data\rowcol.dat', delimiter=' ', unpack=True)
# print(nc)
# Note nc[0] is not actually used but keep it for sanity checks.

# print(nc[0], nc[1])

# split up string into the desired 2-D array
# -create a new 2-D array with 5 columns and nc[1] rows

time_array = np.reshape(xat, (1, nc[1]))
# ==================================================================================

##### Repeat for 3 remaining arrays ######
# - note nc[] is good for all arrays.

Site='Negro Creek'
Year='2009'
month='06'

address=r'C:\Users\Farnoush\Desktop\Data\LaCie\Sites\Negro Creek\2009\Jun\re200906.negro creek'

# Wind magnitudes
xav = np.loadtxt('{}/longdatav.dat'.format(address), delimiter='\,' , unpack=True)
#print(xav)

# split up string into the desired 2-D array
# -create a new 2-D array with nrc[0] rows and nrc[1] columns
# (though the program re-calculates nrc[0] for itself)
vmag_array = np.reshape(xav, (-1, int(nrc[1])))

# Wind directions
xaph = np.loadtxt('{}/longdataph.dat'.format(address), delimiter=',', unpack=True)

# split up string into the desired 2-D array
vphi_array = np.reshape(xaph, (-1, int(nrc[1])))

## vertical winds
xaw = np.loadtxt('{}/longdataw.dat'.format(address), delimiter='\,' , unpack=True)
xaw=xaw/10.0

# print(xaw)
# split up string into the desired 2-D array
w_array = np.reshape(xaw, (-1, int(nrc[1])))

# sample extraction of rows and columns for vmag
# a row is data for all heights at a given time
# a column is all data for all times at a given height
# ** Note first rows and columns are index = 0 (ZERO!)
n=1
COR=[]
SLOPE=[]
while n< int(nrc[1]):
    vcol=vmag_array[:, n]
    wcol =w_array[:, n]
    phcol=vphi_array[:,n]
    tlen=len(wcol)

teta = 1.0  # angle of beam from vertical
p = np.pi

Sumw=[]
Sumwmag=[]
ncount=[]
averw=[]
avermag=[]

phi0=0.
Cor=[]
Slope=[]

while phi0 <360.0:
    Wcoli = []
    Wmag=[]
    I = []
    wmag2=0.0
    sumw=0.0
sumwmag=0.0
sumwmag2=0.0
multisum=0.0
ncount=0.0

i=0

while i < tlen:
    if vcol[i] >-1000.0:
        if phcol[i]>-1000.0:
            if wcol[i]>-100.0:
                I.append(i)
                Wcoli.append(wcol[i])

    #print (i,vcol[i],phcol[i],wcol[i],phi0)

    wmag = vcol[i]* (np.sin(teta * p / 180)) * (np.cos((phcol[i] + 180 - phi0) * p / 180))
    Wmag.append(wmag)
    ncount=ncount+1.0

    sumw=sumw+wcol[i]       #sum of measured w
    sumwmag=sumwmag+wmag    #sum of modeled w (wmag)
    wmag2 = wmag2 + wmag * wmag

    multisum=multisum+wcol[i]*wmag

    i=i+1

summulti=sumwmag*sumw
8.1. Python Scripts-Correlation

```python
sumwmag2 = sumwmag * sumwmag

averw = sumw/ncount
avermag = sumwmag/ncount

slope=((ncount*multisum-summulti)/(ncount*wmag2-sumwmag2))

sumwdev2=0
sumwmagdev2=0
sumcross=0
j=0
while j<len(Wcoli):
    wdev=Wcoli[j]-averw
    wmagdev=Wmag[j]-avermag

    cross=wdev*wmagdev
    sumcross=sumcross+cross

    wdev2=wdev*wdev
    wmagdev2=wmagdev*wmagdev

    sumwdev2=sumwdev2+wdev2
    sumwmagdev2=sumwmagdev2+wmagdev2

)

j=j+1

rw=mt.sqrt(sumwdev2)
rwmag=mt.sqrt(sumwmagdev2)
```
cor = sumcross/(rw*rwmag)

phi0 = phi0 + 10.0
Cor.append(cor)
Slope.append(slope)

COR.append(Cor)
SLOPE.append(Slope)

n = n + 1

CORarray = np.reshape(COR, (-1, 36))
CORARRAY = np.array(CORarray)
SLOPERshape = np.reshape(SLOPE, (-1, 36))
print(len(SLOPERshape))
SLOPERESHAPE = np.array(SLOPERshape)

# Finding the Maxes
MaxCors = np.amax(CORARRAY, axis=1)

print('Max cores are:', MaxCors)

Indcore = np.argmax(CORARRAY, axis=1)
Maxphcor = [(i) * 10 for i in Indcore]

print('phases where correlation is max are:', Maxphcor)
MaxSlopes = np.amax(SLOPERESHAPE, axis=1)

print ('Max slopes are:', MaxSlopes)
Indslope = np.argmax(SLOPERESHAPE, axis=1)
Maxphslope = [(i)* 10 for i in Indslope]

print('phases where slope is max are:',Maxphslope)

# Finding slopes where the correlation is max
MaxSlopeCor= SLOPERESHAPE[np.arange(len(CORARRAY)), np.argmax(CORARRAY, axis=1)]

print('These are the slopes where correlations are max:',MaxSlopeCor)

# saving and reading results of phies in a csv file
Maxphcorearray=np.array(Maxphcor)

#df = pd.DataFrame([])
df = pd.read_csv(r'C:\Users\Farnoush\Desktop\Data\LaCie\Maxes\Phies\Negrocreek\MaxP-+Year+-'+Site+'.csv')
print('Maxphcorearray:',Maxphcorearray,'lenght:',len(Maxphcorearray))
df[Year+month] =Maxphcorearray
df.to_csv(r'C:\Users\Farnoush\Desktop\Data\LaCie\Maxes\Phies\Negrocreek\MaxP-+Year+-'+Site+'.csv',index=False)

df = pd.read_csv(r'C:\Users\Farnoush\Desktop\Data\LaCie\Maxes\Phies\Negrocreek\MaxP-+Year+-'+Site+'.csv')
Negrocreek\MaxP-‘+Year+-’+Site+.csv')
column_labels =list(df.columns.get_values())
Phiarray=[]
for column in df:
    #print(np.array(df[column]))
    Phiarray.append(np.array(df[column]))

Phiarray=np.array(Phiarray)
PHIARRAY=np.reshape(Phiarray,(-1,26))

print(PHIARRAY)
print(Phiarray.dtype,len(PHIARRAY),column_labels)

######################################################
# saving and reading slopes where the correlation is max
######################################################

MaxSlopeCorarray=np.array(MaxSlopeCor)
MAX=float(format(np.amax(MaxSlopeCorarray), '.2f'))

dff = pd.read_csv(r'C:\Users\Farnoush\Desktop\Data\LaCie\Maxes\Slopes\Negrocreek\MaxS-’+Year+-’+Site+.csv')
dff[Year+month] =MaxSlopeCorarray
dff.to_csv(r'C:\Users\Farnoush\Desktop\Data\LaCie\Maxes\Slopes\Negrocreek\MaxS-’+Year+-’+Site+.csv',index=False)

dff = pd.read_csv(r'C:\Users\Farnoush\Desktop\Data\LaCie\Maxes\Slopes\Negrocreek\MaxS-’+Year+-’+Site+.csv')
column_labels =list( dff.columns.get_values())
slopearray=[]
for column in dff:
    #print(np.array(df[column]))
slopearray.append(np.array(dff[column]))
Slopearray=np.array(slopearray)
SLOPEARRAY=np.reshape(Slopearray,(-1,26))
SLOPEARRAY=np.around(SLOPEARRAY, decimals=2)
print(SLOPEARRAY)

############################################# saving and reading results
for maximum correlation
#############################################
MaxCors=np.array(MaxCors)
MAX=float(format(np.amax(MaxCors), '.2f'))

dff = pd.read_csv(r'C:\Users\Farnoush\Desktop\Data\LaCie\Maxes\Corrs\Negrocreek\MaxC-\'+Year+'-'+Site+'.csv')
dff[Year+month] =MaxCors
dff.to_csv(r'C:\Users\Farnoush\Desktop\Data\LaCie\Maxes\Corrs\Negrocreek\MaxC-\'+Year+'-'+Site+'.csv',index=False)

dff = pd.read_csv(r'C:\Users\Farnoush\Desktop\Data\LaCie\Maxes\Corrs\Negrocreek\MaxC-\'+Year+'-'+Site+'.csv')
column_labels =list( dff.columns.get_values())

MaxCorsarray=[]

for column in dff:
    #print(np.array(df[column]))
MaxCorsarray.append(np.array(dff[column]))
MaxCorsarray=np.array(MaxCorsarray)
MAXCORRSARRAY=np.reshape(MaxCorsarray,(-1,26))
MAXCORRSARRAY=np.around(MAXCORRSARRAY, decimals=2)
print(MAXCORRSARRAY)

#############################################################
CORRELATION PLOT
#############################################################

fig = plt.figure(figsize=(6, 3.2))
ax = fig.add_subplot(111)
#ax.set_title('Correlation ')
plt.title('Correlation')
plt.xlabel('Angle( )')
plt.ylabel('Altitude(km)')
scale_x = 0.1
scale_y =2
ticks_x = ticker.FuncFormatter(lambda x, pos: '{:g}'.format(x/scale_x))
ax.xaxis.set_major_formatter(ticks_x)
ticks_y = ticker.FuncFormatter(lambda x, pos: '{:g}'.format(x/scale_y))
ax.yaxis.set_major_formatter(ticks_y)
pylab.pcolor(CORARRAY, cmap='jet',vmin=-0.8, vmax=0.8)
ax.set_aspect('auto')
plt.colorbar(orientation='vertical')
ax.minorticks_on()
ax.grid(which='major', linestyle='-', linewidth='0.5', color='black')
ax.grid(which='minor', linestyle='-', linewidth='0.5', color='black')
txt = Site + 'C' + Year + 'C' + month
fig.text(.8, .01, txt, ha='center')
plt.savefig(r'C:\Users\Farnoush\Desktop\Data\LaCie\Sites\Negrocreek\2010\Nov\C'+Year+month+Site+'.png')

fig = plt.figure(figsize=(6, 3.2))
ax = fig.add_subplot(111)
plt.title('Slope')
plt.xlabel('Angle( )')
plt.ylabel('Altitude(km)')
scale_x = 0.1
scale_y = 2
ax.xaxis.set_major_formatter(ticks_x)
ax.yaxis.set_major_formatter(ticks_y)
levels = np.arange(0.0, 2, 0.1)
cc = pylab.pcolor(SLOPERESHAPE, cmap='jet')
cc.cmap.set_over('brown')
ax.set_aspect('auto')
cb = plt.colorbar(cs, orientation='vertical')

plt.rcParams['axes.facecolor'] = 'white'
plt.rcParams['axes.edgecolor'] = 'white'
plt.rcParams['grid.alpha'] = 1
plt.rcParams['grid.color'] = '#cccccc'

ax.minorticks_on()
ax.grid(which='major', linestyle='-', linewidth='0.5', color='black')
```python
import subprocess
import numpy as np
import matplotlib.pyplot as plt
import pandas as pd
import numpy as np
import matplotlib as mpl
import csv

Site = 'Negrocreek'
Year = '2009'
output_dir = r'C:\Users\Farnoush\Desktop\Data\LaCie\Maxes\plots\Negrocreek\Moving\Averages2\Slopes\2009'

dff = pd.read_csv(r'C:\Users\Farnoush\Desktop\Data\LaCie\Maxes\Slopes\Negrocreek\MaxS-2009-Negrocreek.csv')

column_labels = list(dff.columns.get_values())

slopearray = []

for column in dff:
```
# print(np.array(df[column]))
slopearray.append(np.array(dff[column]))
Slopearray = np.array(slopearray)

SLOPEARRAY = np.reshape(Slopearray, (-1, 26))

SLOPEARRAY = np.around(SLOPEARRAY, decimals=3)

SLOPEARRAY1 = SLOPEARRAY[1:len(SLOPEARRAY) - 1]

#---------------------------------- dff

def dff['Heights']

dff.drop(dff.columns[[12]], axis=1, inplace=True)

dff['LastColumn'] = np.nan

#adjusted_columns (11)
#print('dff:', dff)

########################################### dff

DataFrame (Running Mean) ############

dfm = dff.rolling(3, axis=1, min_periods=1, closed=None).mean()
dfm.drop(dfm.columns[[0, 1]], axis=1, inplace=True)

#print('dfm:', dfm)

RolMean = []

for column in dfm:
    RolMean.append(np.array(dfm[column]))
RolMean = np.array(RolMean)

ROLMEAN = np.reshape(RolMean, (-1, 26))
```
# print('ROLEMEAN', ROLMEAN, len(ROLMEAN))

y = [0.5, 1.5, 2, 2.5, 3, 3.5, 4, 4.5, 5, 5.5, 6, 6.5, 7, 7.5, 8, 8.5, 9.5, 10, 10.5, 11, 11.5, 12, 12.5, 13]
i = 0

while i < len(ROLMEAN):
    fig = plt.figure(1)
    # print(ROLMEAN[i])
    plt.plot(ROLMEAN[i], y, '.-', label=column_labels[i + 2])
    i = i + 1

plt.xlabel('Moving Mean')
plt.ylabel('Height')
plt.title('Height Vs. Moving average')
plt.xlim(left=0)
plt.ylim(bottom=0)
plt.xticks(np.arange(0, +2, 0.25))
plt.yticks(np.arange(0, max(y) + 1, 2.5))
plt.grid(b=None, which='both', axis='both')
plt.minorticks_on()
plt.legend()

txt = Site + ' ' + Year
fig.text(.8, .01, txt, ha='center')

####################################### Saving the Graph

# Exception
# Saving the Graph

def mkdir_p(mypath):
```
from errno import EEXIST
from os import makedirs, path

try:
    makedirs(mypath)
except OSError as exc:  # Python >2.5
    if exc.errno == EEXIST and path.isdir(mypath):
        pass
    else:
        raise

mkdir_p(output_dir)

fig.savefig('{}/Height Vs.Moving_average.png'.format(output_dir))

# DataFrame (Running STD) ##############

dfs = dff.rolling(3, min_periods=1, axis=1).std()
dfs.drop(dfs.columns[[0,1]], axis=1, inplace=False)

RolStd = []

for column in dfs:
    # print(np.array(df[column]))
    RolStd.append(np.array(dfs[column]))
RolStd = np.array(RolStd)
ROLSTD = np.reshape(RolStd, (-1, 26))

y = [0.5, 1, 1.5, 2, 2.5, 3, 3.5, 4, 4.5, 5, 5.5, 6, 6.5, 7, 7.5, 8, 8.5, 9,]
```
9.5, 10, 10.5, 11, 11.5, 12, 12.5, 13]

i = 0

while i < len(ROLSTD):
    fig = plt.figure(2)
    plt.plot(ROLSTD[i], y, '.-', label=column_labels[i +1])
    i = i + 1

plt.xlabel('Moving_STD')
plt.ylabel('Heights')
plt.title('Height_Vs.Moving_STD')
plt.xlim(left=0)
plt.ylim(bottom=0)

plt.xticks(np.arange(0, 1, 0.1))
plt.yticks(np.arange(0, max(y) + 1, 2.5))

plt.grid(b=None, which='both', axis='both')
plt.minorticks_on()
plt.legend()

txt = Site + ' ' + Year
fig.text(.8, .01, txt, ha='center')

#!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!  Saving
# the Graph #!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!!

def mkdir_p(mypath):
    from errno import EEXIST
    from os import makedirs, path

    try:
```
```python
makedirs(mypath)

except OSError as exc:  # Python >2.5
    if exc.errno == EEXIST and path.isdir(mypath):
        pass
    else:
        raise

# output_dir = r'C:\Users\Farnoush\Desktop\Data\LaCie\Maxes\plots\Wilberforce\Moving Averages1\Slopes\2009'
mkdir_p(output_dir)

fig.savefig('{}/Height Vs.Moving_STD.png'.format(output_dir))

# (3) DataFrame (Running Mean both dimensions) ######
dfm.loc[len(dfm)] = 'Nan'

dfmm = dfm.rolling(3, min_periods=1, axis=0, closed=None).mean()
dfmm = dfmm.drop([0, 1], axis=0)
# print('dfmm:', dfmm)

dfmmT = dfmm.T
# print(dfmmT)
# dfmT.plot();

RolMean1 = []

for column in dfmm:
    RolMean1.append(np.array(dfmm[column]))
RolMean1 = np.array(RolMean1)
```
```python
ROLMEAN1 = np.reshape(RolMean1, (-1, 25))
# print('ROLEMEAN1', ROLMEAN1, len(ROLMEAN1))

y = [1, 1.5, 2, 2.5, 3, 3.5, 4, 4.5, 5, 5.5, 6, 6.5, 7, 7.5, 8, 8.5, 9, 9.5, 10, 10.5, 11, 11.5, 12, 12.5, 12.75]
i = 0

while i < len(ROLMEAN1):
    fig = plt.figure(3)
    plt.plot(ROLMEAN1[i], y, '.-', label=column_labels[i + 2])
    i = i + 1

plt.xlabel('Moving Average')
plt.ylabel('Average Height')
plt.title('Moving Average Height Vs. Moving average in both directions')
plt.xlim(left=0)
plt.ylim(bottom=0)

plt.xticks(np.arange(0, +2, 0.25))
plt.yticks(np.arange(0, max(y) + 1, 2.5))

plt.grid(b=None, which='both', axis='both')
plt.minorticks_on()
plt.legend()

txt = Site + ' ' + Year
fig.text(.8, .01, txt, ha='center')

################################################################### Saving the Graph

def mkdir_p(mypath):
```
8.2. Python Scripts-Standard Deviations-Slopes

```python
from errno import EEXIST
from os import makedirs, path

try:
makedirs(mypath)
except OSError as exc:  # Python >2.5
    if exc.errno == EEXIST and path.isdir(mypath):
        pass
    else:
        raise

# output_dir = r'C:\Users\Farnoush\Desktop\Data\LaCie\Maxes\plots\Wilberforce\Moving Averages1\Slopes\2009'
mkdir_p(output_dir)

fig.savefig('{}/Moving Average Height Vs.Moving average both.png'.format(output_dir))

######################################################## (4)
DataFrame (Running STD both dimensions) ##########
dfs.loc[len(dfs)] = 'Nan'

dfss = dfs.rolling(3, min_periods=1, axis=0, closed=None).std()
dfss = dfss.drop([0,1], axis=0)
# print('dfss:', dfss)

dfssT = dfss.T
# print(dfssT)

RolStd1 = []
```
for column in dfss:
    # print(np.array(df[column]))
    RolStd1.append(np.array(dfss[column]))
RolStd1 = np.array(RolStd1)
ROLSTD1 = np.reshape(RolStd1, (-1, 25))
# print('ROLESTD1 ', ROLSTD1, len(ROLSTD1))
y = [1, 1.5, 2, 2.5, 3, 3.5, 4, 4.5, 5, 5.5, 6, 6.5, 7, 7.5, 8, 8.5, 9, 9.5,
     10, 10.5, 11, 11.5, 12, 12.5, 12.75]
i = 0
while i < len(ROLSTD1):
    fig = plt.figure(4)
    # print(i, ROLSTD1[i])
    plt.plot(ROLSTD1[i], y, '.-', label=column_labels[i + 1])
    i = i + 1

plt.xlabel('Moving STD')
plt.ylabel('Average Height')
plt.title('Moving Average Height Vs. Moving STD in both directions')
plt.xlim(left=0)
plt.ylim(bottom=0)
plt.xticks(np.arange(0, +1, 0.1))
plt.yticks(np.arange(0, max(y) + 1, 2.5))
plt.grid(b=None, which='both', axis='both')
plt.minorticks_on()
plt.legend()

txt = Site + ' ' + Year
fig.text(.8, .01, txt, ha='center')

Saving the Graph

```python
def mkdir_p(mypath):
    from errno import EEXIST
    from os import makedirs, path

    try:
        makedirs(mypath)
    except OSError as exc:  # Python >2.5
        if exc.errno == EEXIST and path.isdir(mypath):
            pass
        else:
            raise

#output_dir = r'C:\Users\Farnoush\Desktop\Data\LaCie\Maxes\plots\Wilberforce\Moving Averages2\Slopes\2010'
mkdir_p(output_dir)

fig.savefig('{}/Moving Average Height Vs. Moving STD both.png'.format(output_dir))
```

Averages of STD

```python
RolStd1 = RolStd1[~np.isnan(RolStd1)]
# print(RolStd1)

PowStd = np.power(RolStd1, 2)

total = np.sum(PowStd)
```
```python
count = len(PowStd)

wtotal = total / count

rwtotal = np.sqrt(wtotal)

print('Powstd:', PowStd, 'len_powstd:', len(PowStd), 'count:', count, 'total:', total, 'wtotal:', wtotal, 'rwtotal:', rwtotal)

mydict = {Site + ' ' + Year: rwtotal}

with open(r'C:\Users\Farnoush\Desktop\Data\LaCie\Maxes\Slopes\STDs2.csv', 'a', newline='') as f:
    fieldnames = ['Sites-Year', 'STD']
    writer = csv.DictWriter(f, fieldnames=fieldnames)
    writer.writeheader()
    data = [dict(zip(fieldnames, [k, v])) for k, v in mydict.items()]
    writer.writerows(data)

# Standard Deviation of Standard Deviations
STDstd = []

STDstd = {Site + ' ' + Year: dfss.std(axis=0, skipna=True)}
Dff = pd.DataFrame(STDstd)
#Dff=Dff.drop('201012')
Dff.to_csv(r'C:\Users\Farnoush\Desktop\Data\LaCie\Maxes\Slopes\STDstd.csv', mode='a')
```
8.3  Python Scripts-ANOVA

```python
import pandas as pd
import numpy as np
import scipy as sci
from scipy.stats import f_oneway

# 2009

dfNegro2009 = pd.read_csv(r'C:\Users\Farnoush\Desktop\Data\LaCie\Maxes\Slopes\Negrocreek\MaxS-2009-Negrocreek.csv')
dfNegro2009 = dfNegro2009.drop(['Heights'], axis=1)
Negro2009 = dfNegro2009.to_numpy()
Negro2009 = Negro2009[~np.isnan(Negro2009)]
Negro2009Len = len(Negro2009)
print(Negro2009, Negro2009Len)

dfWilber2009 = pd.read_csv(r'C:\Users\Farnoush\Desktop\Data\LaCie\Maxes\Slopes\Wilberforce\MaxS-2009-Wilberforce.csv')
dfWilber2009 = dfWilber2009.drop(['Heights'], axis=1)
Wilber2009 = dfWilber2009.to_numpy()
Wilber2009 = Wilber2009[~np.isnan(Wilber2009)]
```

---

8.3  Python Scripts-ANOVA
Wilber2009Len = len(Wilber2009)
print(Wilber2009, Wilber2009Len)

dfWal2009 = pd.read_csv(r'C:\Users\Farnoush\Desktop\Data\LaCie\Maxes\Slopes\Walsingham\MaxS-2009-Wal.csv')
dfWal2009 = dfWal2009.drop(['Heights'], axis=1)
Wal2009 = dfWal2009.to_numpy()
Wal2009 = Wal2009[~np.isnan(Wal2009)]
Wal2009Len = len(Wal2009)
print(Wal2009, Wal2009Len)

dfMcgill2009 = pd.read_csv(r'C:\Users\Farnoush\Desktop\Data\LaCie\Maxes\Slopes\Mcgill\MaxS-2009-Mcgill.csv')
dfMcgill2009 = dfMcgill2009.drop(['Heights'], axis=1)
Mcgill2009 = dfMcgill2009.to_numpy()
Mcgill2009 = Mcgill2009[~np.isnan(Mcgill2009)]
Mcgill2009Len = len(Mcgill2009)
print(Mcgill2009, Mcgill2009Len)

dfHarrow2009 = pd.read_csv(r'C:\Users\Farnoush\Desktop\Data\LaCie\Maxes\Slopes\Harrow\MaxS-2009-Harrow.csv')
dfHarrow2009 = dfHarrow2009.drop(['Heights'], axis=1)
Harrow2009 = dfHarrow2009.to_numpy()
Harrow2009 = Harrow2009[~np.isnan(Harrow2009)]
Harrow2009Len = len(Harrow2009)
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```python
print(Harrow2009, Harrow2009Len)

dfEur2009 = pd.read_csv(r'C:\Users\Farnoush\Desktop\Data\LaCie\Maxes\Slopes\Eureka\MaxS-2009-Eur.csv')
dfEur2009 = dfEur2009.drop(['Heights'], axis=1)
Eur2009 = dfEur2009.to_numpy()
Eur2009 = Eur2009.flatten()
Eur2009 = Eur2009[~np.isnan(Eur2009)]
Eur2009Len = len(Eur2009)
print(Eur2009, Eur2009Len)

print(F2009)
```

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