ABSTRACT

NUNALEE, CHRISTOPHER GARNER. A Dynamical Characterization of Atmospheric von Kármán Vortex Streets Induced by Bluff Topography. (Under the direction of Dr. Sukanta Basu.)

Since the 1950s, Earth-orbiting satellites have regularly observed mesoscale vortex street structures imprinted in marine stratocumulus cloud decks, in maritime environments all around the world. These vortex street structures occupy a large range of spatial scales and are always associated with lower tropospheric flow around bluff mountainous islands. Due to qualitative similarities, atmospheric vortex streets are typically referred to as the atmospheric analog of classical low-Reynolds number von Kármán vortex streets (VKVSs), which have been long observed in flow around two-dimensional cylinders in laboratory studies. Despite the obvious similarities between the two fluid phenomena, there are several distinct differences. For example, atmospheric VKVSs have an additional non-homogeneous dimension (i.e., the vertical axis), an entirely different formation mechanism, and an enormous Reynolds number difference (i.e., due to viscosity and length scale differences). As a result of these differences, it is unclear whether the dynamical nature of atmospheric VKVSs share the same quantitative character as their classical counterparts.

In this dissertation, a dynamical characterization of atmospheric VKVSs is presented and then compared to long-standing documentation of low-Reynolds number VKVS dynamical properties. Through the use of high-performance numerical modeling of realistic atmospheres, it is found that several of the same dynamical features associated with classical VKVSs can also be found in association with atmospheric VKVSs. These features include vortex street shedding frequency, multiple side-by-side vortex street interaction phasing regimes, and gap-jet behavior between two vortex streets. In developing this dynamical characterization, effective length and time scales are identified and used to specifically catalog certain properties. Based on this, the results presented here support the idea that atmospheric vortex streets do indeed behave as the atmospheric analog of classical von Kármán vortex streets. In conjunction with this finding, the results also provide evidence to support quasi-universal similarity relationships which describe vortex streets in a scale-invariant sense.

Building on the results described above, the implications of the identified dynamical nature of atmospheric VKVSs are explored in the context of their impact on the refraction of long-range optical rays. It is found that the realistic vortex shedding associated with bluff islands can result in temporal perturbations of simulated ray heights on the order of tens of meters, at distances around 50 km downstream. Moreover, the dynamical properties outlined in this work may have several other applications in areas such as scalar dispersion and air-sea interaction.
A Dynamical Characterization of Atmospheric von Kármán Vortex Streets Induced by Bluff Topography

by

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BIOGRAPHY

Growing up in Wilmington, NC, Chris Nunalee’s interest in atmospheric science was initially sparked by his passion for nature and being outdoors. In particular, he learned to respect the power of wind while spending time offshore fishing with his family from a very young age. In 2007, Chris began undergraduate study at North Carolina State University, majoring in Meteorology, where he later graduated with honors in 2011. During the summer of 2010, Chris was fortunate enough to engage in an internship with MESO Inc. where he was first introduced to wind power meteorology. At about the same time, Chris connected with professor Sukanta Basu who was coincidentally in the process of building a new boundary layer meteorology research group at NCSU and who would later become his graduate school adviser.

During the summer of 2011, Chris began graduate studies at NCSU after engaging in undergraduate research in the fields of offshore wind resource assessment and mesoscale modeling. At the conclusion of his first academic year as a graduate student, Chris broadened his scientific exposure by spending several months at the National Center for Atmospheric Research (NCAR) in Boulder, CO where he did research on dispersion modeling using large-eddy simulation within NCAR’s Research Applications Laboratory. Chris would also return to NCAR in 2013 to complete his project on dispersion modeling, and in doing so he was also recognized as the first annual recipient of the Warner Internship for Scientific Enrichment award for his community involvement efforts. Also in 2013, Chris engaged in a 3 month industrial internship with WindSim AS in Tønsberg, Norway which enabled him to gain a more balanced perspective of wind modeling and its value to the wind energy industry.

In January 2015, Chris assumed a permanent position with WindLogics, a subsidiary of NextEra Energy Inc., as a Senior Resource Modeling Analyst in Juno Beach, FL. In his free time, Chris enjoys doing outdoor activities such as fishing, skiing, and playing soccer as well as spending time with his friends, family, and fiancé Jennifer.
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Furthermore, the following individuals are acknowledged for directly providing data, constructive suggestions, and/or professional guidance: Paul E. Bieringer, Roger Brode, Caroline Draxl, Arne Gravdahl, Steve Hanna, Ákos Horváth, Gary Lackmann, Larry Mahrt, Catherine Messiner, Kevin Mueller, Richard Rotunno, Andrea Vignaroli, and Jeffery Weil. Also, this work would not have been possible without the teachings of the graduate faculty of the Department of Marine, Earth, and Atmospheric Sciences along with collaboration with countless members of the NCSU boundary layer meteorology group from 2010 through 2015.

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Chapter 1

Introduction

Coherent vortex street structures, such as those shown in Figure 1.1, are common atmospheric features found in visible satellite imagery [165, 244]. Vortex streets are essentially rows of counter-rotating vortices shed in the lee of bluff bodies immersed in fluid flows. Vortex streets are probably best known for their ability to maintain mechanically stable circulations well downstream of their vorticity production source [229]. The visually intriguing mesoscale vortex streets (e.g., Figure 1.1) are regularly referred to as atmospheric analogs of classical von Kármán vortex streets (VKVSs) which have been observed in laboratory fluid flows for over one hundred years [26, 145, 227]. The examples shown in Figure 1.1 contain marine stratocumulus cloud decks which essentially trace the underlying vortex street wind pattern. That being said, it is important to note that atmospheric vortex streets are not necessarily confined to remote island settings and, as will be detailed later in Chapter 2, the physics which support their development can potentially exist in virtually any location where a bluff topographical or man-made feature is present (e.g., a mountain range, a hill, or a tall building). Despite their visual similarity to classical, low Reynolds number VKVSs, atmospheric VKVSs have a number distinct differences. For one, they occur in a gaseous fluid and not in a liquid fluid which means that there is a large difference in viscosity. In addition, atmospheric VKVSs have an entirely different formation mechanism from the classical VKVSs. That is, classical VKVSs form through the separation of a viscous boundary layer, while atmospheric VKVSs form through the tilting of baroclinically generated horizontal vorticity into the vertical axis [192, 208]. This dependency on baroclinic forcing gives atmospheric VKVSs a unique distinction from the heavily studied low Reynolds number VKVSs [182, 191, 193]. Furthermore, atmospheric VKVSs have an additional non-homogeneous dimension (i.e., the vertical axis) along with an obvious scale separation. Collectively, these differences justifies the question of whether or not the sound knowledge-base developed around classical low Reynolds number VKVS behavior can be justly applied to atmospheric-scale vortex streets.
Figure 1.1: MODIS satellite imagery of von Kármán vortex streets imprinted in the marine stratocumulus clouds downstream of Guadalupe Island (left panel) and the Canary Islands (right panel). Note the coherent interaction of the Canary Island vortex streets.

A few studies have implied that isolated atmospheric VKVSs do indeed share similar non-dimensional static behavior with classical VKVSs [81, 244]. However, a quantitatively thorough assessment of corresponding dynamical similarities, such as vortex shedding frequency, eddy decay rate, and multi-vortex interaction, is needed. This last example mentioned is particularly interesting given what is known about the interaction of multiple side-by-side low Reynolds number VKVSs (details in Section 2.2). Essentially, low Reynolds number VKVS interaction can be classified into one of four categories: (1) as a single bluff-body wake, (2) as a flip-flopping wake, (3) as an in-phase synchronized wake, or (4) as an anti-phase synchronized wake [8, 107, 219]. The extension of these classes of quasi-idealized multiple vortex street interaction to the atmosphere has not been explored in the literature. However, multiple side-by-side VKVS patterns occur commonly as illustrated in Figure 1.1 (right panel). Consequently, these interactive wake patterns are perhaps very relevant to specific local and regional wind and temperature patterns since many of the locations where VKVSs have been observed are near rugged multi-island archipelagos (e.g. Aleutian Islands, Canary Islands). Therefore, in addition to supporting fundamental scientific research, characterizing the similarities between low and high Reynolds number VKVSs also has several practical implications. For example, the non-linear periodicity associated with vortex street phasing may help to explain some of...
the dynamics of orographic gap winds [25, 73, 172] while the self-stabilizing nature of certain vortex phasing regimes (i.e., anti-phase synchronization) may contribute to the extreme length and time scales of island wakes in the atmosphere [207, 242].

1.1 Objectives and Science Questions

The objective of this research is aimed at quantitatively evaluating the spatio-temporal similarities between low Reynolds number VKVSs and those that occur on the atmospheric scales. Given the vastness of this objective, this work will primarily focus on characterizing the wake structure and periodicity of isolated and interactive VKVSs; in parallel, this work will also investigate some of the implications of these characteristics on other physical processes. An emphasis is placed on cataloging the effective parameters which largely influence dynamical wind and temperature regimes near island archipelagos. These parameters include dominant length, time, and velocity scales. By investigating the dynamical relationships between physical parameters (e.g., island geometry, stability, and wind speed) and vortex street properties (e.g., shedding frequency, eddy lifetime, etc.), it is anticipated that a more complete theory of atmospheric vortex street dynamics can be developed. In conjunction, this research also has an element dedicated to investigating some of the practical implications of applying classical VKVS theory to the atmospheric scales. This latter component deals with specific applications such as wind predictability and the effects of VKVSs on long range optical ray propagation.

The primary science questions raised by this research are:

1) Can atmospheric vortex streets be accurately simulated using conventional mesoscale numerical modeling frameworks and if so what inherent limitations exists?
2) What length and time scales dominate vortex shedding frequency? Does this adhere to Strouhal-Reynolds number similarity theory for low Reynolds number flow past small bluff bodies?
3) Can multiple quasi-idealized atmospheric vortex streets demonstrate phasing regimes similar to those observed in low Reynolds number vortex streets? What mechanisms support three-dimensional VKVS synchronization at this scale?
4) In what capacity does vortex street phasing influence the spatio-temporal behavior of gap winds?
5) How is ambient turbulent diffusion, thermal stability, and momentum modified in atmospheric vortex streets?
6) What are the implications of VKVSs on long range optical ray propagation near mountainous islands?
In order to address these questions, a multifaceted hypothesis testing framework is utilized based on a rigorous mesoscale numerical modeling approach. Using strategic quasi-idealized and realistic numerical simulations, the behavior of atmospheric VKVSs will be isolated from external processes, analyzed, and compared to results reported from low Reynolds number experiments.

1.2 Dissertation Outline

The underlying theme of this research aims to use advanced mesoscale numerical modeling platforms to understand and characterize the (dis)similarities between classical low Reynolds number vortex streets and mesoscale vortex streets which are associated with atmospheric flow past mountainous islands. That being said, the chapters of this dissertation are outlined as follows:

Chapter 2 provides an in-depth literature review of previous research on the topic of atmospheric VKVSs along with a review of pertinent studies dealing with low Reynolds number two dimensional vortex streets. Then, in Chapter 3 we discuss the scientific methods used throughout this study which is centralized around the use of numerical atmospheric modeling.

In Chapter 4 we highlight the importance of using accurate and representative static topographic relief data when using a numerical mesoscale model to simulate atmospheric wake patterns. In Chapter 5 we qualitatively and quantitatively compare and contrast mesoscale model simulations to satellite-derived stereoscopic wind vectors at cloud top levels. In doing so, we evaluate the ability of the mesoscale model to realistically capture the spatial and quasi-temporal properties of atmospheric vortex streets. In addition, we also comment on the model sensitivity to planetary boundary layer (PBL) parameterization with respect to island wake simulations. The topic of PBL scheme sensitivity is then further discussed in Appendix A.

In Chapter 6 we explore the periodicity of isolated von Kármán vortex streets and compare our observations to the Reynolds number - Strouhal number (Re-Sr) similarity theory associated with low Reynolds number vortex streets. In addition, we characterize the effective length scale of mesoscale vortex streets and use it to describe the extension of the classical Re-Sr similarity theory to the geophysical scale.

Next, in Chapter 7 we investigate the behavior of multiple interacting vortex streets induced by two side-by-side identical islands. Here, we compare our results to those produced from similar direct numerical simulations of lower Reynolds number flows and observational studies. Furthermore, we analyze the features of the resulting gap-jet and discuss its functional relationship to the vortex street interaction.

Additionally, in Chapter 8 we simulate three realistic vortex streets and couple mesoscale model output to a geometric optics-based ray tracing code. Using this system, we document
the influence of atmospheric vortices on long-range optical ray trajectories and describe the physics of refractive anomalies embedded within the simulated vortex streets.

Finally, in Chapter 9 we provide a chapter-by-chapter succinct account of the results presented throughout this dissertation and recall intrinsic themes which arose throughout this research. Furthermore, we discuss future directions of this work which incorporates the use of large-eddy simulation (LES) modeling approaches. LES modeling of stratified flow around an isolated small hill as it pertains to passive dispersion is referenced to in Appendix B.
Chapter 2

Background

The majority of past research dealing with atmospheric VKVSs has been devoted to understanding the necessary ambient conditions and vorticity budget which supports their formation. A few decades ago, Etling [67] emphasized the importance of an elevated layer of stable stratification below the island height for vortex formation (see Figure 2.1). This requirement is based on the popular dividing streamline concept introduced by Snyder et al. [209]. The dividing streamline concept hypothesizes that in stably stratified flow past an obstacle, a critical height exists which separates flow over the crest of the obstacle from flow around the flanks of the obstacle.

For the conditions described in Figure 2.1, the stable free atmosphere above the atmospheric boundary layer capping thermal inversion supports the presence of a dividing streamline below the island crest. Below the dividing streamline, the flow is essentially two-dimensional, which is a critical component for forming the two-dimensional vortices downstream.

Mathematically, the dividing streamline \( H_c \) can be estimated by Eq. 2.1 where \( H \) is the height of the obstacle and \( Fr \) is the Froude number defined in Eq. 2.2. Here, \( U \) is the upstream wind velocity and \( f_{BV} \) is the Brunt-Väisälä frequency (Eq. 2.3) where \( g \) is the gravitational constant, \( \Theta_o \) is a reference potential temperature, and \( \theta \) is mean potential temperature as a function of height \( z \) above ground level. Essentially, \( Fr \) is a ratio of inertial forces to buoyancy forces and therefore it acts as an effective indicator of the expected flow behavior and wake regime associated with stratified flow around a bluff body. As shown in Figure 2.2, at low Fr (i.e., weak stability) the flow is free to pass over the crest of the island whereas with increasing Fr the flow tends to split and go around the island flanks.

\[
H_c = H(1 - Fr) \quad (2.1)
\]
\[ F_r = \frac{U}{Hf_{BV}} \]  

\[ f_{BV} = \sqrt{\frac{g}{\Theta_o}} \frac{df}{dz} \]  

Below the elevated capping inversion layer, environments conducive to VKVS formation are typically characterized by a well-mixed atmospheric boundary layer (ABL) [133]. Application of the dividing streamline concept to VKVSs implies lateral flow around the obstacle within the stratified layer between the ABL inversion and \( H_c \) [93]. If, for example, \( H_c \) of the island is beneath the elevated inversion then a fully over-obstacle flow may develop along with potential downstream gravity waves [40, 164]. This effect is clearly visible in Figure 2.3 as the shorter island more upstream produces a gravity wave and the higher obstacle downstream produces a VKVS.

From a large-scale dynamics perspective, mesoscale VKVSs can be conceptualized using a potential vorticity (PV) framework where positive and negative potential vorticity (PV) banners develop downstream of each side of the island. These banners are amplified through interaction with a gravity wave aloft, above the dividing streamline [6, 197, 208]. In this sense, the gravity
wave acts to generate and amplify vertical vorticity by tilting and stretching horizontal vorticity into the vertical axis on the lee slope of the island as illustrated in Figure 2.4 [66]. This concept has also put forth compelling evidence that the effect of surface friction is weak compared to baroclinicity when it comes to the production of vertical vorticity cores in the vortex streets [75].

Furthermore, mesoscale VKVSs have also been conceptualized using a PV non-conservation framework in which each vortex represents a positive or negative PV anomaly. In this case, PV is generated through two primary and interconnected mechanisms. First, enhanced horizontal vorticity is originally produced by dissipation of the mountain wake through the manifestation of a hydraulic jump [198]. Epifanio and Durran [64, 65] explored the properties of such hydraulic jumps in stratified flows past obstacles and found that PV generation was possible in nonlinear viscous wakes even in the absence of surface friction. In conjunction with energy dissipation, baroclinic tilting of horizontal vorticity into the vertical axis is the second dominant process in lee PV production [192].
Research dedicated to understanding the formation of organized vorticity anomalies in the wake of mountainous terrain has progressed significantly in recent decades; however, significantly less is known about the characteristics of these vortices in comparison to vortices produced in the laboratory. Therefore, the focus of this research lies in the fundamental behavior of VKVSs in the atmosphere, irrespective of forcing subtleties.

2.1 Vortex Street Structure and Dynamics

Due to the baroclinic processes involved in the formation of atmospheric vortex streets, the individual vortices that propagate downstream in vortex streets have unique thermal signatures. Essentially, each vortex manifests as a warm core potential temperature anomaly which is characterized by a sagging boundary layer inversion height [81]. As a vortex propagates downstream through the street, the stretching in the column acts partially to sustain the rotational momentum. This behavior can be captured by a mesoscale model as is reflected in Figure 2.5 (top panel).

Figure 2.5 (bottom panel) shows a vertical cross section of potential temperature through a simulated VKVS associated with the island of Madeira. Here the unsettled recirculation zone just downstream of the island is clearly distorting the thermal profile from the incoming upstream profile. In this figure, the simulated low-level thermal profile upstream of Madeira generally adheres to the quasi-idealized theory of VKVS formation of Etling [67] with a capping...
Figure 2.4: Evolution (a-d) of an initial value problem illustrating the tilting of horizontal vorticity into vertical vorticity. Blue (red) contours are negative (positive) buoyancy and blue (red) lines are negative (positive) vortex tubes. Figure cited from [66].

inversion above 600 m msl along with a well-mixed boundary layer beneath. Also, Figure 2.5 (bottom panel) illustrates the presence of a gravity wave signature embedded within the downstream VKVS in addition to a hydraulic jump above the crest of the island. These features are expected according to the theoretical mesoscale dynamics associated with atmospheric VKVSs [197, 198, 208].

Over the years, a few quantitative metrics have been used to spatially characterize VKVSs, some of which were developed by von Kármán himself. Two such metrics are the aspect ratio,

\[ \frac{h}{a} \]  \hspace{1cm} (2.4)

and the dimensionless width,

\[ \frac{h}{d} \]  \hspace{1cm} (2.5)

where \( h \) is the distance between the two counter-rotating vortex trains, \( a \) is the distance between two vortices in the same vortex train and \( d \) is the diameter of the obstacle. Following von Kármán’s theoretical framework for inviscid flow around a cylinder, the aspect ratio yields a
Figure 2.5: Simulated vertical cross-section of instantaneous potential temperature (K) associated with a VKVS near the island of Madeira on July 5th, 2002 (upper panel). The cross-section is oriented streamwise with the upstream to the left and downstream to the right. Simulated vertical cross-section of instantaneous potential temperature (K) collocated with an individual von Kármán vortex (bottom panel). Collectively, in both panels warm core vortex, gravity wave, and hydraulic jump signatures are identifiable along with a distinct atmospheric boundary layer capping inversion above 600 m msl upstream of the island. The details of this simulation will be discussed in later Chapters.
theoretically stable value of 0.28 while the dimensionless width ratio has been observed in the laboratory to be 1.2 [225]. The importance of these dimensionless parameters lies in the fact that they are generally considered to be the universal properties of VKVSs with no dependence on exterior factors. Knowledge of these typical values for laboratory flows has provided some sense of predictability for the turbulent wake of bluff bodies for engineering applications such as aerodynamically-induced vibration [149].

With regards to atmospheric VKVSs, Young and Zawislak [244] analyzed 30 individual VKVS events and determined typical values of $\approx 0.42$ for the aspect ratio and $\approx 1.61$ for the dimensionless width, albeit with a certain margin of error. These differences from laboratory VKVS parameters motivates the speculation that perhaps the universality of these concepts is not applicable to mesoscale atmospheric flows. After all, classical VKVSs have generally been studied in a purely two-dimensional sense; however, in the ABL an additional non-homogeneous dimension is introduced. This extra dimensionality adds gravity to the system and also introduces variability through momentum, heat, and moisture. In addition, the bluff body responsible for the VKVSs varies vertically and can not be idealized as a cylinder. To the best of our knowledge, studies devoted to understanding the influence of these changes on the characteristics of atmospheric VKVSs have not been reported in the scientific literature. This means that dependencies of universal metrics (e.g., aspect ratio) on environmental conditions (i.e., free-stream wind speed, length scale, viscosity, sea surface temperature, etc.) may be present given the significant differences between laboratory and atmospheric VKVSs.

One key to understanding the aspect ratio, and other qualities of the vortex street geometry, is grasping the behavior of the vortex shedding frequency ($N$) or period ($P = 1/N$). This is because, intuitively, $N \approx U_e/a$ where $U_e$ is defined as the downstream eddy propagation speed and $a$ is the distance between two vortices of the same sign of rotation used to determine the aspect ratio. Soon after the discovery of mesoscale VKVSs, Chopra and Hubert [45] studied a particular vortex street induced by the island of Madeira and estimated a period of a little over 7 hours. This estimate was based on the assumption that $U_e$ was approximately 85% of $U_o$ which was well justified based on laboratory observations [31]. As satellite technology advanced, the ability to directly measure $N$ also became more feasible. In 1977, Thomson et al. [223] studied wakes generated by the Aleutian Islands and determined, using satellite pictures and limited meteorological data, vortex shedding periods ranging from 3.8 to 6.2 hours. Even more recently Li et al. [133] used a numerical weather prediction model (i.e., MM5) to simulate a specific VKVS in the wake of the Aleutian Islands and found $P$ to be about 3 hours and 40 minutes for that particular event.

While there is not a deterministic universal model for $N$, much work has been dedicated to characterizing its relationship to the Reynolds number ($Re$). Since the pioneering work of
Roshko [191], many researchers have observed a similarity relationship between the dimensionless Strouhal number \( \text{Sr} \) and \( \text{Re} \), where \( \text{Sr} = NL/V \) and \( \text{Re} = VL/\nu \). In this case, \( \nu, V, \) and \( L \) represent viscosity, a relevant velocity scale, and a relevant length scale (i.e., obstacle diameter), respectively. Essentially, \( \text{Sr} \) can be thought of as a normalized shedding frequency value. This similarity relationship has illustrated the fact that \( \text{Sr} \) behaves uniquely for different ranges of \( \text{Re} \), the asymptotic approach of \( \text{Sr} \) to 0.21 with increasing \( \text{Re} \) is one such characteristic of the \( 10^2 < \text{Re} < 10^4 \) regime [191]. Nevertheless, radically different behavior has been observed at higher \( \text{Re} \) values such as a \( \text{Sr} \) ‘jump’ around \( 10^5 < \text{Re} < 10^6 \) [5]. The formulation of empirical relationships between \( \text{Sr} \) and \( \text{Re} \) has been pursued by many scholars over the past six decades and a few empirical classifications for \( \text{Sr} \) and \( N \) do exist [182]. While beneficial for certain applications, these relationships are theoretically only valid for certain \( \text{Re} \) ranges and only up to \( \text{Re} \approx O(10^9) \). However, \( \text{Re} \) can commonly range from \( O(10^8) < \text{Re} < O(10^{11}) \) in the ABL\(^1\) and yet coherent VKVSs are still clearly visible from satellite imagery. In other words, the presence of naturally occurring VKVSs in the ABL provides the opportunity to document the \( \text{Sr} – \text{Re} \) relationship in the extremely high \( \text{Re} \) regime.

In this research, we aim to document \( N \) for mesoscale atmospheric flows in an effort to gain insight into the temporal behavior of mesoscale atmospheric VKVSs. In addition, we also aim to put our findings into the context of historical laboratory observations by commenting on the natural \( \text{Sr} – \text{Re} \) relationship in the atmosphere. Therefore, the value of this research is twofold: first and primarily, we determine the relationship of \( N \) with \( L \) and secondly, we compare atmospheric VKVSs to classical VKVSs with regards to their inherent \( \text{Sr} – \text{Re} \) relationship.

### 2.2 Low Reynolds Number Vortex Street Interaction

In low Reynolds number, two dimensional flow past multiple side-by-side bluff bodies, a suite of vortex street interaction classes have been cataloged. These classes can be divided into four major categories which are illustrated in Figure 2.6.

The classes of multiple vortex street interaction are largely dictated by the gap ratio, which is the length between the center of two bodies divided by the diameter of each of the bodies. At small gap ratio (i.e., when the two bodies are close to each other) the wake pattern essentially behaves as if the two bodies are a single bluff body. This vortex street wake pattern is known as a single bluff-body wake. As the gap ratio increases, the wake pattern enters a flip-flopping phase where the wake vortex streets transition back and forth between a single bluff-body wake and that of two distinct wake patterns. Then, at even higher gap ratios, two unique vortex street synchronization regimes can develop. These include in-phase and anti-phase synchronization

---

\(^1\)We arrive at these values for \( \text{Re} \) based on magnitudinal analysis assuming \( \nu \approx 10^{-5} \text{m}^2 \text{s}^{-1}, O(1) < V < O(10) \text{m} \text{s}^{-1}, \) and \( O(1) < L < O(100) \text{km}. \)
phases. Experimental studies by Williamson [240] show that in the synchronization phases the two vortex streets merge far downstream at a conversion point and that the location of the conversion point is dependent upon gap-ratio. In other words, the synchronized side-by-side vortex streets could potentially be an elongated version of a single bluff body wake with the assumption that in the single bluff body wake the convergence point is very near to the bodies. Various studies have debated the predominance of in-phase over anti-phase synchronization, and vice versa, over the years. However, direct numerical simulations (e.g., Kang [107]) have shown that both types of wake regimes can exist at the same gap-ratio and that Reynolds number can be influential.

With each gap ratio, and phase of vortex street interaction, the flow between two bluff bodies exhibits substantially different behavior. At small gap ratios, when single bluff-body wakes prevail, the downstream wake behind one body tends to dominate over that of the other body [8, 107]. In these cases, the gap jet between two bodies may have a skewed, or biased, orientation off of the large scale stream flow (2.7). This orientation bias is known to fluctuate over time [8]. Also, at larger gap ratios in the flip-flopping wake regime, the gap jet may have pulsing behavior both in magnitude and orientation [41]. It is expected that if similar vortex street interaction regimes exist in the atmosphere then similar gap jet behavior can also be expected [175]. Consequently, gap jet fluctuations on atmospheric scales can have implications for orographic gap winds [172]; for example, in natural atmospheric channels between two islands (e.g., Alenuihaha Channel).

In this research, we catalog the interaction of multiple side-by-side vortex streets for mesoscale atmospheric cases of flow past Gaussian hills with varying gap-ratios. In parallel, we explore the behavior of the inevitable gap-jet and comment on its implications for wind predictability.
Figure 2.6: Simulated instantaneous vorticity (left panels) and streamlines (right panels) downstream of two identical bluff bodies immersed in low Reynolds number 2 dimensional flow. Each panel demonstrates a different vortex street phasing regime: (a) anti-phase synchronization, (b) in-phase synchronization, (c) flip-flopping, and (d) single bluff body [107].
Figure 2.7: Time averaged particle image velocimetry depicting variations in gap jet axis with respect to gap ratio. Figure cited from [8].
Chapter 3

Numerical Modeling Methods

Numerical modeling will be used as the primary tool to address the science questions introduced in Chapter 1 due to its ability to capture many scale-specific atmospheric dynamics relevant to atmospheric VKVSs. In this research, the public domain Weather Research & Forecasting (WRF) model will be used to carry out all numerical simulations in hindcast mode using boundary conditions provided by various global reanalysis datasets. The WRF model is a finite difference, non-hydrostatic community model and has been used to model complex island wake flows in the past, including VKVSs [49, 40, 170]. The WRF model solves the Reynolds Averaged Navier-Stokes (RANS) equations of motion (Eq. 3.1 - 3.4) on an Arakawa-C grid (See Figure 3.1) and uses a 3rd-order Runge-Kutta method to integrate forward in time.

\[
\frac{\partial \pi}{\partial t} + \bar{u} \frac{\partial \pi}{\partial x} + \bar{v} \frac{\partial \pi}{\partial y} + \bar{w} \frac{\partial \pi}{\partial z} = -f \bar{v} - \frac{1}{\bar{\rho}} \frac{\partial \bar{P}}{\partial x} - \frac{\partial \bar{u}' \bar{w}'}{\partial z} \tag{3.1}
\]

\[
\frac{\partial \bar{v}}{\partial t} + \bar{u} \frac{\partial \bar{v}}{\partial x} + \bar{v} \frac{\partial \bar{v}}{\partial y} + \bar{w} \frac{\partial \bar{v}}{\partial z} = -f \bar{u} - \frac{1}{\bar{\rho}} \frac{\partial \bar{P}}{\partial y} - \frac{\partial \bar{v}' \bar{w}'}{\partial z} \tag{3.2}
\]

\[
\frac{\partial \bar{q}}{\partial t} + \bar{u} \frac{\partial \bar{q}}{\partial x} + \bar{v} \frac{\partial \bar{q}}{\partial y} + \bar{w} \frac{\partial \bar{q}}{\partial z} = -\frac{S_q}{\bar{\rho}} - \frac{\partial \bar{q}' \bar{w}'}{\partial z} \tag{3.3}
\]

\[
\frac{\partial \bar{\theta}}{\partial t} + \bar{u} \frac{\partial \bar{\theta}}{\partial x} + \bar{v} \frac{\partial \bar{\theta}}{\partial y} + \bar{w} \frac{\partial \bar{\theta}}{\partial z} = -\frac{1}{\bar{\rho}c_p} \frac{\partial F}{\partial z} - \frac{L_v}{\bar{\rho}c_p} - \frac{\partial \bar{\theta}' \bar{w}'}{\partial z} \tag{3.4}
\]
In addition, the WRF model uses several physics parameterization schemes to account for sub-grid scale processes. These schemes include representations of radiation (both shortwave and longwave), turbulent diffusion, convection, moist processes (i.e., microphysics), and surface interactions. In this work, the parameterization configuration is primarily static; however, sensitivity of model solutions as it pertains to mesoscale phenomena is described in Appendix A and also in Chapter 5. In all of the results presented here, turbulence in the planetary boundary layer (PBL) was parameterized using K-theory. That is, it is assumed that vertical turbulent fluxes of quantities were proportional to the corresponding vertical gradients of quantities multiplied by some coefficient, or $K$. For PBL modeling, the $K$ coefficient essentially represents eddy diffusivity and its relationship to turbulent fluxes is shown Eq. 3.5.

$$
\frac{u'w'}{\nu} = -K_m \frac{\partial \pi}{\partial z}; \quad \frac{v'w'}{\nu} = -K_m \frac{\partial \pi}{\partial z}; \quad \frac{\theta'w'}{\nu} = -K_h \frac{\partial \theta}{\partial z}; \quad \frac{q'w'}{\nu} = -K_q \frac{\partial q}{\partial z}
$$

(3.5)
Prescription of $K_m$, $K_h$, and $K_q$ is open to many approaches but in this research we primarily utilize the Mellor-Yamada-Janjic (MYJ) scheme [99]. The MYJ scheme is a 1.5 order closure scheme which uses an additional prognostic equation for turbulent kinetic energy (Eq. 3.6) in order to prescribe the eddy diffusivity coefficients.

$$\frac{\partial E}{\partial t} = u'w' \frac{\partial u}{\partial z} - v'w' \frac{\partial v}{\partial z} - \frac{g}{\rho_o} \frac{\partial \rho'}{\partial z} + \frac{\partial}{\partial z} \left( E'w' + \frac{\rho'w'}{\rho} \right) - \epsilon$$ (3.6)

Here, $E$ represents turbulent kinetic energy and $\epsilon$ represents turbulent energy dissipation. Using Eq. 3.6, the eddy diffusivities can be computed based on the relationships shown in Eq 3.7, where $q = \sqrt{2E}$.

$$K_m = l_M S_m q; \quad K_h = l_M S_h q; \quad (3.7)$$

In Eq. 3.7, $l_M$ is a master mixing length scale and $S_m$ and $S_h$ are coupled coefficients that modify $l_M$ based on wind shear and buoyancy. The formulation for $l_M$ is shown in Eq. 3.8 where $k$ is the von Kármán constant (0.4), $\alpha$ is a constant set to 0.25, and $H_{pbl}$ is the height of the PBL.

$$l_M = l_o \frac{k z}{k z + l_o}; \quad l_o = \frac{\int_0^{H_{pbl}} |z| q \, dz}{\int_0^{H_{pbl}} q \, dz} \quad (3.8)$$

$S_m$ and $S_h$ are computed by satisfying Eqs 3.9 and 3.10 where $A_1$, $A_2$, $B_2$, and $C_1$ are constants derived from experimental data. In addition, the coefficients $G_m$ and $G_h$ essentially represent shear turbulent energy production and buoyancy production respectively (Eq. 3.11).

$$S_m(6A_1 A_2 G_m) + S_h(1 - 3A_2 B_2 G_h - 12A_1 A_2 G_h) = A_2$$ (3.9)

$$S_m(1 + 6A_1^2 G_m - 9A_1 A_2 G_h) - S_h(12A_1^2 G_h + 9A_1 A_2 G_h) = A_1(1 - 3C_1)$$ (3.10)
For the purposes of this research, we focus only on mesoscale atmospheric simulations. That is, all modeled data and results are generated using parameterized atmospheric boundary layer turbulence as the horizontal resolutions were generally around 1 - 2 km (i.e., the meso-gamma scale). Alternatively, one could increase the horizontal resolution and move into the microscale atmospheric regime by employing the technique of large-eddy simulation (LES). However, LES requires a horizontal resolution within the inertial subrange and therefore demands much higher computational resource than mesoscale simulations. This is an exciting new area of research and is potentially very complimentary to the physics topic explored here but it is beyond the current scope. Nevertheless meso-gamma scale simulations such as those presented in this work are suitable to capture the important dynamics of mesoscale island wakes and vortex streets.

Through the mesoscale modeling used in this research we are able to gain insight into the parameters that dictate realistic large-scale (i.e., on the order of 10 to 100 km) VKVSs invoked by gradually sloping islands (i.e., low height to width aspect ratio). In conjunction, this enables documentation of the dynamical relationships between simulated mesoscale VKVS behavior and the transient regional atmosphere.

A major theme of this research involves the use of quasi-idealized Gaussian hills to produce VKVSs using realistic atmospheric boundary conditions. Essentially, this allowed for experimentation with the functional relationships between island configurations (i.e., height, width, and, in some cases, gap ratio) and VKVS properties (frequency, size, etc.) and environmental parameters. Furthermore, these relationships were compared and contrasted to those observed in low Reynolds number flows.

Despite the obvious advantages and utility of using a numerical modeling approach for this research, there was nevertheless model solution uncertainty largely attributed to factors such as physics parameterization, numerics, and boundary conditions. Therefore, in Chapter 5 we carry out a model comparison (i.e., validation) against observational data provided by satellite-based wind retrievals at cloud height. Additionally, in Chapter 4 we highlight the importance of using representative static topographic relief data when performing realistic numerical simulations of flow near and around bluff islands.
Chapter 4

High Fidelity Numerical Modeling of Mesoscale Island Wakes and Sensitivity to Static Topographic Relief Data

4.1 Introduction

Massively-parallel computing platforms now enable regional-scale numerical weather prediction (NWP) models\(^1\) to be easily integrated with fine-scale grid spacings, down to approximately 1 km horizontally. A valuable benefit of such high-resolution models is their capability to simulate orographically induced flow phenomena. Examples of such phenomena include gap-winds \cite{147}, lee-rotors \cite{7}, and wake vortices \cite{133}. The accuracy of model simulations of orographic flows has been verified against a suite of observational data including, but not limited to, ground-based instruments e.g., lidar \cite{132}, mesonets \cite{30}; satellite-based remote sensing instruments e.g., SAR \cite{152}; and airborne measurement platforms e.g., aircraft \cite{74}, radiosonde \cite{169}. Despite the increased resolvability, and overall fidelity, offered by finer resolution models as it pertains to orographic flows, mesoscale NWP models are still constrained by multiple factors \cite{55} such as necessary physics parameterizations \cite{57}. The treatment of sub-grid scale (i.e., sub-mesoscale) processes such as turbulence, radiative transfer, moisture phase change, etc. collectively contributes to the uncertainty of model solutions \cite{47}. At the same time, it has also been demonstrated that model uncertainty can be increased through the prescription of inaccurate, or unrepresentative, time-dependent atmospheric boundary conditions \cite{119,177}. In

\(^{1}\)In the context of this article, NWP models refer to models that may run in forecast or hindcast modes.
the past decade, advanced data assimilation techniques, coupled with improved remote sensing capabilities, have been shown to reduce simulation uncertainty \[10, 30\] and increase forecast skill \[185\]. While great efforts have been expended to identify sources of NWP error with respect to model configuration (i.e., physics parameterizations) and dynamic (meteorological) boundary conditions, often overlooked is the sensitivity of model solutions to static boundary conditions, namely topographic relief.

Presently, there exists several global terrain height datasets which can be used by regional-scale NWP models. One of the most used surface relief datasets, named GTOPO30, was developed by the United States Geographic Survey and comprised through a synthesis of numerous international digital elevation models. GTOPO30 contains maximum spatial resolution of 30 arc seconds and is the default dataset for many community models such as the Weather Research and Forecasting (WRF) model. Aside from GTOPO30 data, other satellite-derived global terrain height datasets also exist such as the Shuttle Radar Topography Mission (SRTM) \[69\], and the Advanced Spaceborne Thermal Emission and Reflection Radiometer (ASTER) \[4\]. These datasets offer higher spatial resolutions globally of 3 arc seconds and 1 arc second, respectively. The construction of surface terrain height grids in NWP models from source datasets, such as GTOPO30 or SRTM, typically involves sub-grid scale averaging of the source data, grid-scale spatial interpolation during data ingestion, and/or preprocessing smoothing effects \[e.g., see the WRF model Preprocessing System Documentation; 167\]. Although in many circumstances these activities are necessary, they can effectively result in under-resolved topographic relief. Under-resolved terrain height implies that the NWP model generated terrain height does not fully capture the relevant features of the natural topography described by the source data \[101\] and can result in terrain height discrepancies on the order of tens to hundreds of meters \[102\]. Such discrepancies have been shown to result in significant error in simulated low-level wind fields \[189, 103, 196\]. Aside from under-resolved terrain height in modeled grids, which is essentially an oversimplification of the source terrain height data, we show in this chapter that uncertainty in source terrain height datasets themselves can be significant enough to result in fundamental differences in simulated orographic flow mechanics. This result illustrates that the sensitivity of NWP models can be more complex than 1st-order biases recently documented by \[222\].

In this chapter, we simulate two realistic cases of atmospheric flow past mountainous islands; for each case, we run the WRF model simulations using GTOPO30 and SRTM source terrain height data while keeping all other model configurations identical. From the results, we comment on the fundamental differences in simulated atmospheric wake patterns associated with the two terrain height fields. At the same time we compare the simulated flow features to those expected from visible satellite imagery. Our results will demonstrate that selection of ter-
rain height source data can, in some cases, be critical to successfully capturing the fundamental mechanics of mesoscale orographic wakes.

## 4.2 Case Studies and Modeling Details

Two historical atmospheric events were considered in this chapter, both corresponding to cases of flow past mountainous islands. Since the islands were far from any upstream surface heterogeneity, only the local terrain features associated with the islands acted to perturb the local winds and consequent cloud structures. For these events, the wind wake characteristics associated with each island were indicated by distinct cloud structures captured by visible satellite imagery provided by the Moderate Resolution Imaging Spectroradiometer (MODIS) instrument. The modeled wind wake patterns of the events were compared to one another and the differences were documented in the context of the inferred wake patterns shown in satellite imagery.

The first, and primary, case study involved the Spanish island of Gran Canaria (GC) off the west coast of Northern Africa on 30 April 2007. MODIS visible satellite imagery from this day (Fig. 4.1a) revealed a coherent pattern of dipole vortices (i.e., von Kármán vortices) being shed downstream of GC around 10:30 UTC. GC has a diameter of approximately 50 km at sea level and has a peak elevation of 1948 m a.m.s.l. GC’s SRTM-based topography is shown in Fig. 4.1 for reference.

The second case study presented here involves flow past several islands which collectively comprise the Lesser Antilles (LA) in the Eastern Caribbean. On 31 July 2013, MODIS visible satellite imagery of the Lesser Antilles region (Fig. 4.1c) illustrated distinct wakes behind all of the major islands of the LA. Contrary to the GC case which had a coherent vortex shedding wake regime, the LA case had weak wind wakes where the rotation behind each island was not strong enough to counter the background wind flow. Furthermore, the wakes were correlated with a reduction in cumulus cloudiness and darker sea surface color, a phenomenon investigated by Smith [207]. The windward islands of the LA are generally lower than GC but, nonetheless, are predominately mountainous with peak elevations near 1 km for each island (see Table in Fig. 4.1d).

The numerical simulations performed in this study used the Weather Research and Forecasting (WRF) model which was initialized by ERA-Interim reanalysis data (physics configurations are shown in Table 4.1). The simulations used a nested four domain configuration centered on the islands of interest. Of note, a horizontal grid spacing of 1 km was chosen in the inner-most domain (d04) while the parent domains (d03–d01) used grid spacings of 3, 9, and 27 km, respectively. Additionally, in d04 the control simulations used GTOPO30 terrain height while the experimental simulations used terrain height data interpolated from SRTM 3.
<table>
<thead>
<tr>
<th>Symbol</th>
<th>Island</th>
<th>Peak Elevation</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>Guadeloupe</td>
<td>1467 m</td>
</tr>
<tr>
<td>B</td>
<td>Dominica</td>
<td>1447 m</td>
</tr>
<tr>
<td>C</td>
<td>Martinique</td>
<td>1397 m</td>
</tr>
<tr>
<td>D</td>
<td>St. Lucia</td>
<td>950 m</td>
</tr>
<tr>
<td>E</td>
<td>St. Vincent</td>
<td>1234 m</td>
</tr>
<tr>
<td>F</td>
<td>Grenada</td>
<td>840 m</td>
</tr>
</tbody>
</table>

Figure 4.1: MODIS-TERRA visible satellite imagery of Canary Islands on April 30th, 2007 with Gran Canaria in the upper right (a) and SRTM terrain height profile of Gran Canaria (b). MODIS-TERRA image of the Lesser Antilles on August 1st, 2013 (c). SRTM-derived peak elevation for each of the major windward islands in the Lesser Antilles (d).
data. d01, d02, and d03 used 10, 5, and 2 min GTOPO30 terrain height, respectively. All other modeling variables were held constant between the control simulations and experimental simulations. For both the original GTOPO30 and SRTM terrain height fields, the default smoothing and interpolation methods were selected. That is, 1 pass of the built-in WRF Preprocessing System (WPS) smoother-desmoother and 4 point averaging interpolation, respectively.

Table 4.1: Model Physics Configurations.

<table>
<thead>
<tr>
<th>Parameterization</th>
<th>Name</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Microphysics</td>
<td>WRF Single-Moment 5-class</td>
<td>Hong et al. [86]</td>
</tr>
<tr>
<td>Longwave Radiation</td>
<td>RRTM Longwave</td>
<td>Mlawer et al. [157]</td>
</tr>
<tr>
<td>Shortwave Radiation</td>
<td>Dudhia Shortwave Radiation</td>
<td>Dudhia [59]</td>
</tr>
<tr>
<td>Convection</td>
<td>Kain-Fritsch (d01 and d02)</td>
<td>Kain [106]</td>
</tr>
<tr>
<td>Land Surface</td>
<td>Noah Land Surface Model</td>
<td>Chen and Dudhia [43]</td>
</tr>
<tr>
<td>Planetary Boundary Layer</td>
<td>Mellor-Yamada-Janjić</td>
<td>Janjic [99]</td>
</tr>
<tr>
<td>Surface Layer</td>
<td>Monin-Obukhov Similarity Theory</td>
<td>Monin and Obukov [160]</td>
</tr>
</tbody>
</table>

4.3 Gran Canaria Case Study

In this section, we analyze the atmospheric flow patterns downstream of GC as simulated by the WRF model with GTOPO30 terrain and SRTM terrain. Before beginning the analysis, we compare the discrepancies between the two terrain height data fields. Figure 4.2 presents a southern view of the model terrain height for GC as generated by GTOPO30 and SRTM. Notice that aside from increased ruggedness in the SRTM-based terrain height, there is also a significant increase in peak terrain height of GC island of nearly 1 km. Additionally, Fig. 4.2 also illustrates the upstream mean potential temperature cross-section in the lower troposphere on the day of 30 April 2007. Within the potential temperature cross-section, a well-mixed planetary boundary layer (PBL) can be identified by the nearly constant potential temperature in the lowest 800 m of the atmosphere. Above this layer, in the free atmosphere, a thermal capping inversion was present. Most importantly for the purposes of this chapter, is that the increase in peak elevation of GC with the SRTM data makes the modeled GC island penetrate into the stably stratified free atmosphere.

As the original GTOPO30-based elevation of GC was predominately within the well mixed PBL, the simulated flow around it was mostly 3 dimensional. That is, the impinging air
parcels were able to rise and cross the crest of the island barrier and then descend on the lee slope without significant buoyant restriction. This effect acted to produce the unorganized wake pattern shown in the lower-left panel of Fig. 4.2. Alternatively, with the SRTM-based elevation, the increased topographic steepness along with the layer of stable stratification beneath the maximum height of the island caused much of the flow to split and pass around the lateral flanks of GC. This flow behavior generated coherent lee vortices (i.e., von Kármán vortices) which were
shed downstream of the island, similar to what was observed by the MODIS satellite imagery shown in Fig. 4.1.

In addition to invoking differences in the simulated wake pattern of GC, the SRTM-based and GTOPO30-based simulations also produced substantial variability to the wind regime on GC itself. In Fig. 4.3, an instantaneous streamwise wind speed cross section is presented for both simulations. Of particular note is the wind speed extrema (greater than 17 m/s) on the crest of GC in the GTOPO30-based simulation. This zone of high wind speed was a result of the confluence caused by compression of the air column as it passed over the crest of the island. Alternatively, in the SRTM-based simulation this zone of strong wind speed was not simulated due to the lack of significant air column compression over GC. Instead, the lateral flow around GC produces a zone of weak wind speed along the island centerline with respect to the flow direction.

Figure 4.3: Instantaneous wind speed cross-sections for the GTOPO30-based (top panel) and SRTM-based (bottom panel) simulations at 10:30 UTC on 30 April 2007. Cross sections are oriented in the streamwise axis with inflow to the left.
4.4 Lesser Antilles Case Study

The second case study presented here deals with boundary layer flow impinging on the Eastern slopes of the Lesser Antilles (LA) island archipelago. As can be seen in Fig. 4.1, the wake signatures from all of the major islands in this region persisted for up to approximately 300 km downstream. Contrary to the GC case, the wake patterns in the LA case did not contain strong enough vorticity to counter the ambient wind speed and therefore coherent wake vortices did not form. This type of wake pattern has been called a weak wake pattern by Smith [207], and forms in conditions of slower wind speed and lower island height in comparison with the vortex shedding patterns with the GC case.

In the upper panels of Fig. 4.4, the regional topographic relief is shown for the GTOPO30-based simulation vs. the SRTM-based simulation. Of particular note is the fact that the island of Dominica, one of the more prominent of the islands in the LA in the SRTM-based simulation, is represented as flat (1 m m.s.l.) in the GTOPO30-based model elevation. At the same time, other neighboring islands (e.g., St. Vincent) appear relatively similar, despite them being slightly smaller in size in the GTOPO30-based model. The differences in the depiction of Dominica’s relief in the two simulations manifested in substantial differences in regards to the simulated 6 h mean surface wind speeds. The lower panels in Fig. 4.4 show the mean surface wind fields simulated by the two model runs. Most notably, the weak wind wake associated with Dominica is nearly non-existent in the GTOPO30-based simulation while it extends hundreds of km in the SRTM-based simulation. In addition, the zone of enhanced wind speed associated with funneling between Dominica and its northern neighbor of Guadeloupe is increased in the SRTM-based simulation. Lastly, the unique shapes of the individual island wakes showed signs of variability between the GTOPO30 and SRTM based simulations.

4.5 Summary

In this work, we have simulated two realistic cases of atmospheric flow past mountainous islands using the WRF model. For each case, we explored the sensitivity of the simulated wake patterns with respect to two different terrain height source datasets (i.e., GTOPO30 and SRTM). Our results show cases where the differences in modeled terrain height corresponded to fundamental differences in simulated wake mechanics. For the GC case, the simulation which used GTOPO30 terrain height had a peak island elevation which was nearly 1 km lower than that in the SRTM-based simulation. For this case, the GTOPO30-based terrain did not reach the stably stratified thermal inversion above the planetary boundary layer while the SRTM-based terrain extended hundreds of meters into the free atmosphere. This difference resulted in substantially less vertical vorticity downstream of GC island, along with an area of wind speed
Figure 4.4: The WRF model’s GTOPO30-based terrain height (upper left) and interpolated SRTM terrain height (upper right). Lower left and right panels depict averaged wind speed in the boundary layer from 06:00–12:00 UTC as simulated by the GTOPO30-based run and SRTM-based run, respectively.

Extrema on the crest of the island in the GTOPO30-based simulation. In other words, the SRTM-based simulation produced more significant lateral flow around the island and downstream von Kármán vortices, in agreement with MODIS visible satellite imagery, while the GTOPO30-based simulation facilitated anomalous wind acceleration on the crest of the island and incoherent downstream vortices.

For the LA case, the GTOPO30-based model terrain represented the island of Dominica to be essentially flat and near sea level (i.e., 1 m.m.s.l.) and consequently resulted in no surface wind wake pattern. At the same time, the SRTM-based simulation resulted in a weak wind
wake field which extended hundreds of km downstream of Dominica. The latter was similar to what was illustrated in visible satellite imagery.

This work explored the value of using representative terrain height source data for high resolution mesoscale modeling activities. Moreover, it was highlighted that considerable care should be taken while selecting orographic relief input data when simulating atmospheric flow over, around, and downstream of remote mountainous islands (e.g., Gran Canaria and Dominica). That being said, future studies should evaluate the accuracy global terrain datasets for other locations and their representativeness for mesoscale modeling.
Chapter 5

Evaluation of Mesoscale Model Simulated Vortex Street Winds with MISR Stereoscopic Wind Vectors

5.1 Introduction

Orographically-induced atmospheric vortex streets are common geophysical flow features which regularly form in association with remote island archipelagos. While these large atmospheric features are capable of perturbing the ambient environment in which they exist, for example through air-sea interactions [40, 242], there are limited observational instruments which enable us to comprehensively study them. This limitation is largely due to the facts that (1) vortex streets are a highly dynamic phenomenon in nature, with time scales on the order of a few hours [170, 81], (2) they exhibit extreme spatial heterogeneity, with length scales on the order of hundreds of km, and (3) as previously mentioned, they predominately occur in locations with sparse observational data (i.e., offshore regions). In lieu of the aforementioned factors, numerical modeling is perhaps one of the most effective tools available for studying the natural characteristics and behaviors of atmospheric vortex streets. That is, mesoscale numerical models (i.e., the WRF model) can easily provide high-fidelity atmospheric flow fields with resolutions as high as the km-scale for any location and time. As such, several mesoscale modeling studies have been carried out in the past dealing with the phenomenon of vortex shedding in the lee of mountainous islands (e.g., [49, 133]).

Nevertheless, in order to gain confidence in the ability of mesoscale models to realistically resolve the important attributes of VKVSs, it is essential that comprehensive model validations be performed which compare model-generated vortex street properties to objective (observa-
tional) data. Inevitably, such validations cannot be easily conducted using solely conventional meteorological measurement systems (e.g., wind profilers, radiosondes). Fortunately, advanced satellite-based technology is now capable of providing unique data derivatives which offers a more complete picture of atmospheric vortex streets.

In the early 1960s, at the dawn of the satellite era, Hubert and Krueger [91] first documented atmospheric VKVSs using coarse resolution visible satellite imagery. Since then, sophisticated low-altitude sun-synchronous satellites have begun generating an extensive suite of remote sensing products at resolutions down to about 300 m. At the forefront of this endeavor, the Multi-angle Imaging SpectroRadiometer (MISR) instrument on the TERRA satellite has the ability to provide stereo-derived cloud wind vectors down to a resolution of 4.4 km horizontally. This technology has been applied to atmospheric vortex streets as small as the ones produced by the Norwegian island of Jan Mayen [87]. In this work, two VKVS case studies are presented which were captured by MISR imagery. For each of these cases, we perform high-fidelity WRF model simulations (i.e., simulations with high resolution and the most accurate boundary conditions) and then qualitatively and quantitatively compare the WRF model-generated vortex street winds to satellite-derived MISR stereoscopic wind vectors.

5.2 Data and Methods

The MISR instrument has an orbit that covers the entire earth’s surface in approximately 9 days. In other words, it only provides instantaneous atmospheric information for any one region (approximately 10:30 AM local time). Contained within the MISR instrument are a total of 9 cameras each associated with a different view angle with respect to nadir (i.e., vertically downward towards the earth). Aside from the one nadir camera, the other cameras (i.e., 4 each forward and aftward) maintain angles of 26.1, 45.6, 60.0, and 70.5 degrees with respect to nadir. During each polar orbit trip, MISR provides an image viewing swath of about 360 km which extends the entire circumference of the Earth, roughly 20,000 km. From the high resolution images captured by the MISR cameras (i.e., greater than 300 m), stereoscopic wind vectors and corresponding heights can be extracted based on the orbital velocity of MISR [88], albeit with a coarser resolution of about 4.4 km. Furthermore, the algorithm used to derive the stereoscopic wind vectors includes rigorous quality control operations, yet localized erroneous data can still be present.

Based on the visible satellite imagery provided by MISR, two well-captured vortex street events were identified: one induced by the island of Madeira on July 9th, 2010 (M10), and another induced by Gran Canaria on April 30th, 2007 (C07). These two events, shown in Figure 5.1, were also simulated using the WRF model as previously mentioned. The simulations used ERA-Interim reanalysis data for initialization and began at 0000 UTC the day before the
event was documented by MISR; this allowed for sufficient spin-up time.

The stereoscopic wind vectors measured by MISR are associated with the location-dependent cloud top height (CH). Unfortunately, the precision of the instantaneous MISR CH measurements has an uncertainty range on the order of ±300 m. Figure 5.2 illustrates the estimated CH field for each the M10 and C07 case studies. Clearly, for the Madeira event the CH values varied by hundreds of meters of horizontal distances of just a few kilometers while similar behavior was observed for the C07 case. As will be shown in the following sections, significant spatial variability of wind was associated with this variability of CH. Furthermore, this height uncertainty produced a challenge when comparing MISR winds to the WRF model simulated wind fields. To mitigate for the uncertainty of MISR CH values, we compare WRF wind fields by assuming a similar level of uncertainty. That is, we first compute the mean CH for each case
Figure 5.2: MISR observed cloud top height over Madeira island on July 9th, 2010 (left panel) and Gran Canaria on April 30th, 2007 (right panel).

(672 m msl for M10 and 697 m msl for C07) and then determine the closest WRF model vertical grid level closest to that height. Finally, we allow a ±3 grid level range for comparison. That is, the closest WRF-simulated wind speed to the MISR wind speed, at each location, within the allowed vertical range is used for comparison. In the following sections of this chapter we will compare the WRF model simulated wind fields to those observed by the MISR stereoscopic wind vector algorithm in order to explore the agreement with respect to the two realistic VKVSs.

5.3 Madeira 2010 Case Study

The first comparison case, M10, had good stratocumulus cloud coverage (Figure 5.1) which led to a horizontally comprehensive observed MISR wind field. The measured horizontal wind components, U and V, are shown in Figure 5.3, along with the WRF model simulated wind components. Several interesting features of the vortex street can be observed in Figure 5.3. The zonal U wind speed approximately represented the lateral, or transverse, flow component and had an upstream value of -6 to -4 m/s. In the simulation, a very similar range of values were found; this, in combination with the similarity of the upstream V winds (i.e., approximately -2 to -5 m/s), indicates that the axis of flow impingement onto Madeira was well captured.

Aside from the ambient upstream wind velocity, many flow features inside of the vortex street captured by MISR were also resolved by the WRF model simulation. Foremost, the near-island U wind speed anomalies, present in the MISR field, illustrate a region of relatively
Figure 5.3: MISR-derived stereoscopic $U$ and $V$ component wind speed (top left and bottom left panels respectively) and WRF model-derived $U$ and $V$ component wind speed (top right and bottom right panels respectively) associated with the Madeira case study which occurred on July 9th, 2010 at 1030 UTC. Black boxes (A1-A4) represent areas of wind speed anomalies embedded within the vortex street and the areas in which magnitude distributions are shown in Figure 5.4.
strong westerly flow (e.g., 0 to 2 m/s) both upstream and downstream of Madeira island. In addition, just to the south of this region of Westerly flow is a sharp wind shear zone with U wind speeds dipping to roughly 8 m/s. While these two features were also depicted in the WRF model simulation, they were somewhat less prominent; that is, the length of the shear zone was shorter and less pronounced and the region of relatively strong Westerly wind was less extensive spatially. Farther downstream, between 31 and 32 degrees North, another zone of relatively high U wind speed was observed by MISR. This area was associated with the clearly defined transverse jet shown in block 065 of Figure 5.1 (left panel). Essentially, this jet was separating, while at the same time fueling, the cyclonic and anti-cyclonic vortices to the North and South, respectively. While the exact position and geographic coverage of the aforementioned jet differed slightly in the WRF model simulation with respect to the MISR observation, both datasets agreed well on its relative peak magnitude of 0 - 1 m/s. Additionally, two other (older) jets can be seen, albeit with less clarity, to the Southwest. With regards to these jets, again their precise structures differ between the WRF simulation and MISR observations even more so than the younger jet features to the Northwest. This suggests that the vortex street is perhaps best resolved in the WRF model closer to the island of Madeira.

Moving on, we next inspect the V wind speed comparison between MISR and the WRF model (Figure 5.3). Similar features to the U wind speed comparison are found; that is, there are repeating patterns within the vortex street region. Between 31.5 and 32.5 degrees North, MISR captures a hook-shaped positive V wind speed anomaly with peak values in the range of 0 - 5 m/s. Similarly, the WRF model also produced a hook-shaped V wind speed feature with roughly the same values; however, it was located slightly to the North of the MISR feature. In addition, another positive V wind speed anomaly can be seen farther to the Southwest (between -20 and -18.5 degrees West) which resembles a triangular shape. While the WRF model did produce a similar triangular-shaped V wind speed anomaly, its magnitude was not as extreme and its position was slightly to the North when compared to the MISR anomaly.

The U and V wind speed magnitude distributions associated with the WRF model simulation and the MISR stereoscopic winds for the Madeira Case are shown in Figure 5.4. In all areas (A1-A4), there appeared to be consistent agreement between the two wind fields with respect to the distributions. That is, the minimum and maximum magnitudes of the range of wind speeds agreed to within ± 2 m/s respectively. Furthermore, the shapes of the distributions varied with area; A1 had a narrow range of wind speeds with a peak around -4 m/s which which was consistent between both the WRF model and the MISR measurements. It should also be noted that in A1 the WRF model produced a distribution weighted towards the more positive magnitudes which MISR produced the opposite. In A2, the WRF simulated range of U wind speeds was approximately 2 m/s greater than the MISR range; however, the WRF magnitude
Figure 5.4: Frequency distributions of the U (top panels) and V (bottom panels) component wind speed values bounded within the areas A1 - A4 (See Figure 5.3) at a bin size of .25 m/s. The MISR horizontal grid was interpolated to the WRF model horizontal grid in order to have a consistent number of samples for comparison.
peak was -4 m/s while the MISR peak was -2 m/s. For A3, the V wind speed distribution peak was about -2 m/s in both the WRF and MISR wind fields. Lastly, in A4 the range of distributions demonstrated close agreement between WRF and MISR; however the WRF wind field had a peak shifted about 2 m/s to the more positive side of the the MISR wind field. These differences could be due to several inherent differences between the WRF and MISR results (e.g., a difference in aspect ratio), or a difference in instantaneous phase. The latter of which does not necessarily reflect a fundamental bias between the WRF model winds and the MISR winds.

We also compare the full horizontal wind vectors simulated by the WRF model to those measured by MISR for the M10 case. From Figure 5.5 (upper panel) it can be seen that the ambient upstream wind velocity observed by MISR fell between roughly 4 and 8 m/s from the Northeast. Downstream, but still outside of the vortex street, higher wind speeds, upwards of 12 m/s, were recorded. In regards to the orographically modified flow field (i.e., the vortex street), the flow around the flanks of the island of Madeira had localized wind speed maxima roughly twice the magnitude of the upstream flow. Inside the vortex street, strong rotational and shear vorticity is evident by the changing directions of the vectors. Essentially, each vortex had a total wind direction change of about 90 degrees with wind speed differences up to about 10 m/s across a couple of kilometers. In Figure 5.5 (lower panel), it can be seen that the WRF model simulated vortex street depicts many of the anomalous flow features documented by the MISR stereoscopic winds (e.g., 90 degree wind turning). Nevertheless, the WRF model simulated wind speed extrema associated with individual vortex perimeters appears under-resolved.

Finally, Figure 5.6 illustrates three quantitative agreement metrics between the WRF model and the MISR stereoscopic wind vectors. Since the MISR stereoscopic wind fields are instantaneous (i.e., approximately 1030 UTC) and the WRF model generates a continuous spatio-temporal wind field, we compare each time in the WRF simulations to the instantaneous MISR wind field. Among other things, this allows us to investigate whether or not there is a time shift in the WRF model. In addition to the control WRF simulation already discussed, Figure 5.6 also shows results from two identical simulations but using different PBL schemes. Obviously, the zonal (U) wind speed produces the most distinct vortex street frequency signal with a period of approximately 8 hours. The frequency signal in Figure 5.6 indicates that when the simulated wind vortex street is in-phase with the MISR vortex street that the agreement is higher. In the meridional (V) wind speed and full horizontal wind speed panels the frequency oscillation is less distinct; however, a slight increase in agreement can be seen across all three metrics. As for the variability associated with PBL parameterization, generally there is little difference with regards to the three agreement metrics presented here. The only notable difference is the fact that the YSU and QNSE schemes had slightly longer and shorter (respectively) vortex street
Figure 5.5: MISR and WRF horizontal wind speed (top and bottom panels respectively) at approximately 715 m msl valid for July 9th, 2010 at 1030 UTC.
Figure 5.6: Time series starting at 00 Z on July 9th, 2010 of statistical comparison metrics associated with the WRF and MISR wind fields for Madeira’s wake. Upper left, upper right, and lower center panels are associated with zonal, meridional, and horizontal wind components respectively. Red, green, and blue traces correspond to bias, root mean squared deviation (RMSD), and correlation respectively. Solid lines, thick dashed lines, and thin dashed lines represent WRF model simulations performed using MYJ, YSU, and QNSE PBL parameterizations respectively. Statistics were computed by horizontally interpolating the 1 km WRF grid points to the MISR 4.4 km grid point within blocks 64 through 67.
oscillation periods on the order of about 1 hour. This can be seen in the upper left panel of Figure 5.6 as the second major peak in RMSD and correlation occur at different times with respect to the MYJ peak. Aside from this difference, the relative ranges of RMSD and correlation fluctuations associated with the shedding of vortices is fairly consistent. This adds confidence in the control modeling system which uses the MYJ scheme as it shows that the uncertainty contributed by PBL parameterization is relatively low. For more information on the sensitivity of mesoscale model solutions with respect to PBL scheme, please see Appendix A.

5.4 Gran Canaria 2007 Case Study

Compared to the M10 case, the C07 case had less stratocumulus cloud coverage, especially near the Canary Islands. Nevertheless, MISR still observed a large portion of the vortex street wind signature as shown in Figure 5.7. While many similar features can be identified between M10 and C07, some notable differences are also present. Foremost, due to the orientation of the vortex street with respect to the MISR orbital path, a longer portion of the vortex street was captured and up to 4 individual transverse jets can be seen. As can be seen, the simulated U wind magnitudes inside the transverse jet cores agree the best with MISR closer to the initiation zone (i.e., near Gran Canaria). Further downstream, the U wind anomalies decay faster than what was observed by MISR. This is an indication that the length of the vortex street is underestimated by the WRF model, possibly due to overactive diffusion common in finite difference models such as the WRF model or PBL parameterization. On the other hand, the spacing between vortices was well-represented by the WRF model.

Looking next at the full horizontal wind vectors (Figure 5.8), it can be noted that the ambient upstream simulated windspeed is about 2-4 m/s greater than the MISR wind speeds. This speed overestimation is also evident farther downstream outside of the vortex street. Considering that the wind speed difference is prominent along the upstream (i.e., Northern) domain boundary, it is possible that the ERA-Interim boundary conditions used in this simulation were slightly in error. In order to investigate this possibility, another simulation was run for the C07 case but using GFS-FNL (not shown). However, the differences between the GFS-FNL based simulation and the ERA-Interim based simulation were negligible.

In addition, the quantitative comparison statistics already introduced in Section 5.3 are again presented for the C07 case in Figure 5.9. Similar to the M10 case, a regular fluctuation of quantitative agreement can be seen in regards to the U wind component which corresponds to the simulated vortex street frequency. Furthermore, this frequency is also identifiable in both the V wind component statistics and the full horizontal wind statistics. In this C07 case, the vortex street frequency was higher than in the M10 case at approximately 4 hours. The stronger impinging wind speed in the C07 case would support the increased recorded frequency.
Also, note that the relative magnitude of bias, RMSD, and correlation in this C07 case was comparable to the M10 case for the U wind component. That is, the correlation and RMSD ranged from about 0.5 to 0.9 and 0.5 to 2.5 respectively, depending on the instantaneous phase of the simulated vortex street, and the bias ranged from about -0.5 to 0.0. On the other hand, the V wind component demonstrated much lower agreement as a result of the previously mentioned ambient V wind speed bias, which statistically hovered around 1.0. Despite this, the expected oscillation pattern is still distinguishable in both the V wind component and full horizontal wind magnitude agreement statistics.
Figure 5.8: MISR and WRF horizontal wind speed (top and bottom panels respectively) at approximately 715 m msl valid for April 30th, 2007 at 1030 UTC.
Figure 5.9: Time series starting at 00 Z on April 30th, 2007 of statistical comparison metrics associated with the WRF and MISR wind fields for Gran Canaria’s wake. Upper left, upper right, and lower center panels are associated with zonal, meridional, and horizontal wind components respectively. Red, green, and blue traces correspond to bias, root mean squared deviation (RMSD), and correlation respectively. Statistics were computed by horizontally interpolating the 1 km WRF grid points to the MISR 4.4 km grid point within blocks 68 through 72.
5.5 Summary

We have evaluated the agreement between WRF model simulated winds and MISR stereoscopic winds associated with two realistic VKVS events. It was found that the WRF model generally captured the characteristic wind features inherent to the vortex streets such as flow speed-up around the island flanks and on the outer edges of the vortices, transverse jets, and vortex spacing. Nonetheless, the more subtle features of the vortex street observed by MISR, such as the precise shape of lateral velocity anomalies, were not represented in all locations. Generally, the simulated vortex street flow features agreed with MISR observed flow features closer to the island, Madeira for the M10 case and Gran Canaria for the C07 case. Especially in the C07 case, the older vortices far downstream showed less agreement between the WRF simulation and the MISR stereoscopic wind fields. This supported the possibility that excessive diffusion, intrinsic to the WRF model, was perhaps contributing to the premature dissipation of von Kármán vortices as they propagated downstream. Nevertheless, the quantitative comparison statistics shown in Figures 5.6 and 5.9 indicate that the WRF model is capable of simulating vortex street wind fields that agree with MISR observations and that sensitivity to PBL scheme is possible but not likely a major contributor of uncertainty. However, a more comprehensive intercomparison of the sensitivity of mesoscale modeling to PBL parameterization is contained in Appendix A. All in all, the results presented here support the idea that a mesoscale model such as the WRF model can be used to skillfully simulate the wind fields associated with realistic large island-induced vortex streets in the atmosphere.
Chapter 6

On the Periodicity of Mesoscale von Kármán Vortex Streets

6.1 Introduction

While VKVSs have been heavily studied in the laboratory [182, 191, 193], our scientific understanding of the behavior and characteristics of their atmospheric cousins is generally confined to what has been gleaned from visible satellite imageries [45, 91, 223, 244], idealized numerical simulations [81, 192, 197, 208], and a handful of realistic numerical modeling studies [40, 49, 133]. Based on the facts that observational satellite images simply use marine stratocumulus clouds as tracers for the underlying wind pattern and only provide 2-dimensional plan views with coarse resolution, we have learned little about the quantitative differences between atmospheric and classical vortex street properties. The chronic challenges associated with sparse observational data, and the inherent shortcomings of numerical modeling, have translated into a vague understanding of the properties of mesoscale atmospheric VKVSs. This in turn provokes one to question whether the sound knowledge-base concerning classical VKVSs can be justly applied to atmospheric-scale vortex streets.

While it was once thought that atmospheric VKVSs form in the same way as classical VKVSs (i.e., through boundary layer separation), it has been shown that atmospheric VKVSs are inherently different as they may form through the tilting of baroclinically generated horizontal vorticity into the vertical axis [192, 208] in the absence of a frictional boundary layer. On the other hand, others have advocated the importance of turbulent diffusion and potential vorticity non-conservation [197, 198] for atmospheric VKVS formation. The popularity of

\footnote{The material presented in this chapter appears in the following publication: C.G. Nunalee and S. Basu (2014) On the Periodicity of Mesoscale von Kármán Vortex Streets, Environmental Fluid Mechanics, DOI: 10.1007/s10652-014-9340-9.}
multiple lines of thought such as these gives credence to the elusive nature of mesoscale wake vortices resembling von Kármán vortices. Irrespective of their generation mechanisms, it is well justified to speculate that structures as large as mesoscale VKVSs (i.e., hundreds of kilometers) may have significant implications on other meso/micro scale atmospheric processes (i.e., vertical and horizontal wind shear, thermal stability, turbulent kinetic energy dissipation, etc.). Such interactions, in addition to their implications for societal operations (i.e., astronomy, dispersion, aviation, etc.), drives the need to understand mesoscale VKVS characteristics and behaviors.

In this Chapter, we present new information pertaining to the temporal behavior of atmospheric VKVSs (i.e., vortex shedding period) and compare our findings to classical VKVSs. Through a series of numerical simulations, we highlight some of the factors which govern the vortex shedding frequency \( N \) of mesoscale VKVSs and lay the foundation for empirical relationships between \( N \) and these factors. This study is organized as follows: Section 6.2 presents background information on the atmospheric VKVS problem; in Section 6.3 we outline the scientific questions addressed in this work and the experiments conducted. Results from all modeling experiments are presented in Sections 6.4 and 6.5; and finally in Sections 6.6 and 6.7 we provide a discussion of the results and directions for future work, respectively.

6.2 Background

In 1912, von Kármán and Rubach made the remarkable discovery that the local circulations within VKVS-type wakes are theoretically stable when a certain vortex spacing is observed [229]. This condition of stability means that VKVSs have the potential to concentrate vorticity in narrow wake zones for extremely long distances. Later in the early 1960’s, when satellite observations first became readily accessible, Hubert and Krueger [91] reinforced this stability concept by observing atmospheric VKVSs existing hundreds of kilometers downstream of the bluff body that invoked them. In their report, they compared individual island wakes to laboratory VKVSs and emphasized the fact that these mesoscale eddies seem to maintain energy for very long periods of time, despite being separated from their mechanical sources by very large distances (i.e., > 100 km).

One key to understanding the aspect ratio, and other qualities of the vortex street geometry, is grasping the behavior of the vortex shedding frequency \( N \) or period \( P = 1/N \). This is because, intuitively, \( N \approx U_e/a \) where \( U_e \) is defined as the downstream eddy propagation speed and \( a \) is the distance between two vortices of the same sign of rotation used to determine the aspect ratio. Soon after the discovery of mesoscale VKVSs, Chopra and Hubert [45] studied a particular vortex street induced by the island of Madeira and estimated a period of a little over 7 hours. This estimate was based on the assumption that \( U_e \) was approximately 85% of \( U_o \) which was well justified based on laboratory observations [31]. As satellite technology
advanced, the ability to directly measure $N$ also became more feasible. In 1977, Thomson et al. [223] studied wakes generated by the Aleutian Islands and determined, using satellite pictures and limited meteorological data, vortex shedding periods ranging from 3.8 to 6.2 hours. Even more recently Li et al. [133] used a numerical weather prediction model (i.e., MM5) to simulate a specific VKVS in the wake of the Aleutian Islands and found $P$ to be about 3 hours and 40 minutes for that particular event.

While there is not a deterministic universal model for $N$, much work has been dedicated to characterizing its relationship to the Reynolds number ($Re$). Since the pioneering work of Roshko [191], many researchers have observed a similarity relationship between the dimensionless Strouhal number ($Sr$) and $Re$, where $Sr = NL/V$ and $Re = VL/\nu$. In this case, $\nu$, $V$, and $L$ represent viscosity, a relevant velocity scale, and a relevant length scale (i.e., obstacle diameter), respectively. Essentially, $Sr$ can be thought of as a normalized shedding frequency value. This similarity relationship has illustrated the fact that $Sr$ behaves uniquely for different ranges of $Re$, the asymptotic approach of $Sr$ to 0.21 with increasing $Re$ is one such characteristic of the $10^2 < Re < 10^4$ regime [191]. Nevertheless, radically different behavior has been observed at higher $Re$ values such as a $Sr$ ‘jump’ around $10^5 < Re < 10^6$ [5]. The formulation of empirical relationships between $Sr$ and $Re$ has been pursued by many scholars over the past six decades and a few empirical classifications for $Sr$ and $N$ do exist [182]. While beneficial for certain applications, these relationships are theoretically only valid for certain $Re$ ranges and only up to $Re \approx O(10^8)$. However, $Re$ can commonly range from $O(10^8) < Re < O(10^{11})$ in the ABL\(^2\) and yet coherent VKVSs are still clearly visible from satellite imagery. In other words, the presence of naturally occurring VKVSs in the ABL provides the opportunity to document the $Sr – Re$ relationship in the extremely high $Re$ regime.

In this chapter, we have employed a rigorous technique to document $N$ in an effort to gain insight into the temporal behavior of mesoscale atmospheric VKVSs. In addition, we also aim to put our findings into the context of historical laboratory observations by commenting on the natural $Sr – Re$ relationship in the atmosphere. Therefore, the value of this research is twofold: first and primarily, we determine the relationship of $N$ with $L$ and secondly, we compare atmospheric VKVSs to classical VKVSs with regards to their inherent $Sr – Re$ relationship.

### 6.3 Methodology and Simulation Details

In the present study, we have used one of the most advanced mesoscale numerical weather prediction models, the Weather Research and Forecasting (WRF) model [205], to simulate and study the intrinsic properties of atmospheric VKVSs. Recently, several studies have demonstrated

\(^2\)We arrive at these values for $Re$ based on magnitudinal analysis assuming $\nu \approx 10^{-6}$ m\(^2\) s\(^{-1}\), $O(1) < V < O(10)$ m s\(^{-1}\), and $O(1) < L < O(100)$ km.
that mesoscale numerical models can accurately simulate atmospheric VKVSs [40, 49, 133]. Particularly, Couvelard et al. [49] found the WRF model to realistically capture atmospheric VKVSs with a spatial resolution of only 6 km. However, they noted the need for increased resolution in future studies; thus we have selected a resolution of 2 km in this study (Table 6.1).

After validating the model’s numerical representation of a specific VKVS event, we measure the vortex shedding frequency based on the timing of the characteristic transverse jets which develop between the discharge of two counter rotating vortices [193]. Combining this estimate for the vortex shedding frequency with approximate length and velocity scales, we put forth deterministic values for $Sr$ in the atmosphere. Next, we adjust the length scale of the system by varying the diameter and height of a quasi-idealized island surrogate; this effectively modifies $Re$. While keeping all other conditions constant, we document the $Sr – Re$ relationship for atmospheric VKVS events. In parallel, we also explore the dominance of the horizontal and vertical length scales (i.e., island diameter and height) of the system with respect to the observed $Sr – Re$ relationship. Finally, we test the universality of the observed relationship by simulating additional VKVS events invoked under diverse settings in two other locations across the globe.

The Portuguese island of Madeira was selected as the primary island studied in this work due to its rugged terrain perpendicular to the regional Northeastern trade winds and popularity among previous VKVS researchers [27, 40, 45, 49, 67]. The island of Madeira is fairly mountainous and covers an area of about 823 km$^2$ (Figure 6.1) with a maximum elevation of 1,862 m above mean sea level (msl). It is a very rugged landmass with a central chain of mountains
that essentially runs west/east, maintaining a high elevation throughout. Madeira is one of the most suitable islands for numerical studies considering it is among the largest islands known to produce a clearly discernible VKVS pattern in the marine stratocumulus clouds. This is important for this study because even the geometry of Madeira is near the lower resolvable limit of modern mesoscale models such as the WRF model [204]. From NASA’s Moderate Resolution Imaging Spectroradiometer (MODIS; [111], [112], [179]) image gallery a control VKVS event was identified in the wake of the Portuguese island of Madeira on July 5th, 2002 (left panel of Figure 6.2).

Figure 6.2: MODIS satellite imagery of multiple von Kármán vortex streets in the wake of Madeira Island and the Canary Islands (Left) and WRF model simulated cloud cover (Right) near Madeira for July 5, 2002. Left image is a courtesy of NASA’s MODIS website [165].

In this study, we have used a similar modeling configuration to that used in the successful Madeira simulations of Couvelard et al. [49]. However, in the present simulations we have used higher horizontal resolution of 2 km by 2 km (See Table 6.1). Two domains were used, an outer and an inner nested domain; and in both domains there were 61 pressure-based, terrain following vertical coordinates between the surface and 100 mb with approximately 28 levels below 1 km above ground level (AGL). This vertical resolution is similar to typical operational
numerical weather prediction (NWP) models; however, the horizontal resolution is significantly higher than most NWP models [1]. The simulations conducted for this study were run for a total of 48 hours with the first 24 hours being used as spin-up time for the model. The final 24 hours of the simulations were used for analysis and enabled us to study the temporal evolution of the vortex street, specifically the shedding rate. The physics parameterizations used in these simulations included: Mellor-Yamada-Janjić PBL scheme [99]; WRF Single-Moment 5-class Microphysics [86], RRTM Longwave Radiation [157], Dudhia Shortwave Radiation [59], Kain-Fritsch Cumulus in the outer domain [106], and Noah Land Surface Model [43]. For initialization and lateral boundary conditions, the ERA-Interim reanalysis dataset produced by the European Centre for Medium Range Weather Forecasts [29] was used. It contains environmental data records from 1979 through the present at 6-hour intervals, and includes a horizontal resolution of approximately 79 km with 60 vertical coordinate levels. In this particular study, we have not explored the sensitivity of our numerical VKVS simulations to physics parameterization schemes or model configuration, but this would be an interesting subject for future research.

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<td>451</td>
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<td>902 km × 902 km</td>
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6.4 Real-Terrain Case Study

For this particular event (Figure 6.2), the wind in the lower troposphere was impinging on Madeira from the north-northeast as shown by the orientation of the lee vortices in Figure 6.2; these vortices are traced by the marine stratocumulus cloud deck present at the top of the boundary layer. Most importantly, the wind direction was nearly perfectly perpendicular to Madeira’s mountain range. This caused maximal horizontal flow blockage sufficient to invoke the observed VKVS. Also in Figure 6.2, the VKVS is observed to maintain a coherent pattern for a very long distance downstream of Madeira (i.e., hundreds of kilometers). Additionally, smaller von Kármán vortices can be seen shedding from the nearby Canary Islands to the southeast of Madeira.

The WRF model qualitatively simulated the atmospheric VKVS associated with the real Madeira terrain fairly well (right panel of Figure 6.2). The diameter of each simulated vortex
closely resembled the actual vortex size (i.e., roughly the cross-stream diameter of Madeira) immediately downstream of Madeira. Further downstream, the two vortex trains steadily diverged from one another and each vortex became larger and more diffused. From the simulated cloud pattern (Figure 6.2) the distance between the vortices of the same rotation was similar to the satellite observations; however, the overall length of the vortex street was chronically underestimated by the WRF model. This could perhaps be a result of numerical diffusion or the PBL parameterization used.

As with most numerical modeling efforts, quantitative inaccuracies were present in the WRF simulation. Radiosonde profiles of wind speed and potential temperature at Madeira’s capital of Fuchal are shown in Figure 6.3 in comparison with simulated profiles. On this day, Fuchal was on the leeward side of the island which contributed to the strong decrease in wind speed beneath about 1000 m msl. From the observations, a thermal inversion was apparent around 800 m msl which was an important environmental component for the development of the vortex street; below the inversion the radiosonde recorded a well-mixed boundary layer. The WRF model struggled to capture these features and instead produced a fairly stratified ABL at the location of Fuchal. Also, the simulated and observed free-atmosphere thermal profile was noticeably different above 2000 m with the WRF producing a warm bias of about 5 K. However, the thermal profile depicted by the WRF model just upstream of Madeira was much more similar to that recorded by the radiosonde with a well mixed boundary layer beneath a strong inversion. Although the simulated inversion was a few hundred meters lower than what was observed in the radiosonde, the radiosonde temperature measurements may have been altered by the thermal effects induced by the island itself while the upstream modeled profile corresponded to an entirely marine ABL.

Nonetheless, it is important to note that the highly-dynamic nature of island wakes makes it difficult to compare model solutions to instantaneous observations. Figure 6.4 displays a vertical cross-section through the island of Madeira in the streamwise direction; here the unsettled recirculation zone just downstream of the island is clearly distorting the thermal profile from the incoming upstream profile. In this figure, the WRF-simulated low-level thermal profile upstream of Madeira generally matches what was observed by the radiosonde with a capping inversion above 600 m msl along with a well-mixed ABL beneath. Also, Figure 6.4 illustrates the presence of a gravity wave signature embedded within the downstream VKVS in addition to a hydraulic jump above the crest of the island. These features are expected according to the mesoscale dynamics known to be associated with atmospheric VKVSs [197, 198, 208].

The stratified layers above the inversion acted to divert the flow around the flanks of the island. Since most of the flow traversed the island laterally, we assumed that the flow speed-up over the crest of the island was minimal. Therefore, we took the wind speed at around
1500 m msl to be roughly the free-stream, pressure-driven (i.e., geostrophic) wind speed and characteristic velocity scale of the flow ($U_o$). In this case, we will assume $U_o$ to be on the order of 10 m s$^{-1}$ which is consistent between both the numerical simulation and the observation. Using this velocity scale, and the thermal profile previously discussed, we arrived at a Brunt-Väisälä frequency of approximately 0.017 Hz and a Froude number of approximately 0.23 for the simulated 0 - 1800 m msl layer. From here, we estimated the dividing streamline height for this real case to be approximately 1212 m msl.

The WRF model simulated the vortex street as having a lateral spacing ($h$) of about 54 km and a longitudinal spacing ($a$) of about 187 km. This produced an aspect ratio of 0.29. This value was considerably smaller than the average aspect ratio observed by Young and Zawaislak [244] for atmospheric VKVSs of 0.42, but closer to the theoretically derived stable aspect ratio of 0.28 [229]. Using the assumed values of the upstream wind speed ($U_o$) and the vortex spacing ($a$), we can estimate $1/N \approx 6$ h using the assumption made by Chopra and Hubert [45] that the eddy propagation speed ($U_e$) is approximately 85% of ($U_o$) [31]. Furthermore, this yields $Sr$ approximately equal to 0.20 assuming a length scale equal to the cross-stream length of the island ($\approx 45$ km) at its base. However, it should be noted that the validity of the eddy propagation speed assumption is questionable in the atmosphere as it was derived based on laboratory VKVSs; in our simulations we observed $U_e$ to be as low as 60% of $U_o$.

The wind perturbations associated with the VKVS propagated downstream in a regular pattern. Figure 6.5 shows that the presence of Madeira’s steep topography primarily redirects upstream momentum around the island laterally. This increases wind speeds on the flanks of

Figure 6.3: (Left) Observed and simulated wind speed vertical profiles at Funchal Madeira. (Right) Observed and simulated potential temperature vertical profiles at Fuchal, Madeira and at a point just upstream of Madeira. Both plots valid for July 5th, 2002.
the island up to nearly 20 m s$^{-1}$, a 100% increase from the upstream wind flow which was about 10 m s$^{-1}$. These two high wind speed zones were separated by somewhat of a recirculation zone where wind speeds were generally less than 1.5 m s$^{-1}$. Although the recirculation zone meanders with the shedding of each counter-rotating vortex, its edges may easily have a wind difference of 15 m s$^{-1}$ over 6 kilometers or less. Additionally, the enhancement of wind shear in this region directly leads to enhanced turbulence and narrow ribbons of anomalously high turbulent kinetic energy (not shown).

High momentum transverse jets are also visible in the lee of Madeira and can be observed penetrating across the wake from alternating sides. Generally, these jets changed in their orientation by about 90$^\circ$ from one another but in some cases up to 180$^\circ$ from one another. Essentially, the momentum withheld in these jets evolve to form the outer rings of each propagating vortex. This is what initially fuels the high tangential wind speeds in the rim of each vortex. These wind speed maxima can be seen in Figure 6.5 maintaining themselves for hundreds of kilometers downstream, nearly all the way south to the Canary Island archipelago. The alternating jets occur at regular intervals and correspond to the vortex shedding frequency. The two snapshots in Figure 6.5 illustrate nearly identical vortex streets separated in time by 5 hours. During these 5 hours, the vortex pattern has completed one entire period and has returned back to its original phase. The passage of two transverse jets across one particular location represents the shedding frequency of the vortex street, and will thus be used in Section 6.5 as a measurable predictor of $S_r$. 

Figure 6.4: Streamwise vertical cross-section of simulated potential temperature (K) through the island of Madeira (left is upstream and right is downstream). Gravity wave and hydraulic jump signatures are identifiable along with a distinct ABL capping inversion above 600 m msl upstream of the island.
Figure 6.5: Wind speed contours (colored) and horizontal wind velocity vectors (black) at ≈110 m msl at 4 UTC (top) and at 9 UTC (bottom). Notice that the vortex street appears nearly identical at these two times despite being separated in time by 5 hours; this highlights the roughly 5-hour period of the street.
6.5 Quasi-Idealized-Terrain Case Study

Next, additional simulations were ran for the same VKVS event described in Section 6.4. But, in each new simulation the island of Madeira was substituted with an idealized bell-shaped island of predefined diameter and height (See Eq. 6.1). Efforts to systematically vary these two aspects of the island exposed the natural sensitivity of the VKVS shedding frequency to the fundamental length scale of the system. The experimentation presented in this section has two primary components: (1) the height of the island is kept constant (i.e., similar height to the real island of Madeira; 1800 m msl) and the diameter is varied and (2) the height of the island is varied while the diameter is held constant for a few cases and varied for others. Next, in Section 6.5.2 two unrelated VKVS events were simulated in other parts of the world under different conditions but with the control, idealized Madeira island (i.e., $H = 1800$ m and base diameter = 40 km) replacing the actual islands.

The geometry of the idealized islands were defined by

$$Z_{xy} = \frac{H}{4} \left[ 1 + \cos \left( \frac{2\pi(x - d)}{d} + \pi \right) \right] \left[ 1 + \cos \left( \frac{2\pi(y - d)}{d} + \pi \right) \right]$$

where $Z_{xy}$ is the local terrain height of the island as a function of zonal ($x$) and meridional ($y$) position, $H$ is the maximum terrain height at the center of the island, and $d$ is the diameter of the base of the island. These new islands used the modal land-use classification of central Madeira and were symmetrical in both horizontal axes but had varying diameters and maximum heights. The control quasi-idealized simulation essentially duplicated the major geometry of Madeira with $H = 1800$ m and base diameter of 40 km. The new idealized islands had base diameters of 20, 60, 80, 100, 140, and 200 km and at the height of the upstream simulated thermal inversion ($\approx 800$ m msl; see Section 6.4) the islands were roughly 60% of their base diameter for the control height for the Madeira event. As will be subsequently shown in Section 6.5.1, the two cases with island diameters of 140 and 200 km did not produce any form of VKVS signature. This was important because it provides some of the first evidence of the upper $Re$ range for atmospheric VKVSs. On the other hand, we tested the smallest length scales possible given our mesoscale modeling approach. Since this study is concentrated on mesoscale VKVSs we essentially captured the entire range of possible mesoscale VKVSs, albeit that covers only about one order of magnitude of $Re$.

In addition, the maximum height of the islands were also adjusted to new heights of 1400, 2200, and 2800 m in order to test the sensitivity of the $Sr - Re$ relationship to this (vertical) length scale. It is critically important to note that the formulation for idealized terrain height used here (Eq. 6.1) only weakly adjusted island diameter at the inversion height when $H$ is varied.
and base diameter is held constant. That is, the difference in island diameter at inversion height for the $H = 1400$ m and $H = 2800$ m cases is only about 7 km (Figure 6.6). This behavior is desirable because it demonstrates if and how $Sr$ responds to varying $H$ when the largest horizontal length scale above the inversion is held nearly constant.

Figure 6.6: Vertical profiles of terrain height through the center-line of each idealized island. Note that at the height of the ABL inversion the diameters of all islands are very similar regardless of differing $H$.

Additionally, two other VKVS events were identified from the MODIS satellite imageries. These events were associated with the rugged Pacific Ocean islands of Guadalupe off of the west coast of Baja Mexico, and the westernmost island in the Juan Fernández archipelago, Alejandro Selkirk Island. For both cases, the actual islands were replaced with the control idealized island used in the Madeira case and the shedding frequency of the simulated VKVSs were documented. Idealized islands were used instead of real terrain for these cases in order to avoid introducing terrain-induced bias into our periodicity sensitivity tests and also because Guadalupe and Alejandro Selkirk Islands are considerably smaller than Madeira which would have required adjusting the resolution of our numerical grid in order to simulate them with high fidelity. By examining the characteristics of these two additional events we were able to assess whether background meteorological and/or geographical conditions (e.g., global positioning, boundary layer height, upstream wind speed, etc.) affect $Sr$ for atmospheric VKVSs.
6.5.1 Madeira Event

In order to quantitatively document the shedding frequency in each simulation, $U$ and $V$ wind speed time series were taken from locations downstream of the islands which experienced the passage of each transverse jet. The downstream locations where the time series were taken correspond to roughly one island diameter downstream and at heights from the surface up to 2000 m msl. From these time series, Fourier spectral analysis and autocorrelation functional analysis were used to compute the VKVS period, and eventually $Sr$, at each grid level within the lower troposphere. This approach produced approximately 50 $P$ (i.e., $Sr$) samples per simulation or case; the mean and standard deviation of these values were computed for analysis and are eventually presented in Table 6.2.

Assuming that the islands’ diameters at the height of the inversion represented an effective length scale, the respective Reynolds numbers for each Madeira simulation (in increasing order) were $1.2 \times 10^{10}$, $2.4 \times 10^{10}$, $3.6 \times 10^{10}$, $4.8 \times 10^{10}$, $6.0 \times 10^{10}$, $8.4 \times 10^{10}$, and $1.2 \times 10^{11}$. Please note that the WRF model neglects molecular viscosity in the governing momentum equations; however, the effects of molecular dissipation are parameterized within the TKE equation. By assuming the standard molecular viscosity of air, we were able to estimate $Re$ based on the simulated wind speed. On the other hand, the WRF model computes eddy viscosity ($K_M$) which provides the opportunity to estimate modified Reynolds numbers ($Re_K$) using an eddy viscosity approach instead of the traditionally molecular viscosity approach. The computed $Re_K$ values are presented in Table 6.2 for comparison to $Re$ and the values of $K_M$ are presented later in this section along with a brief discussion.

Figure 6.7 displays how $N$ varied with island diameter for the quasi-idealized Madeira events. The passage of the transverse jets act as proxies for the periods of the VKVSs and are demarcated by the times with the strong wind speeds shown by the hot colors and longer vectors. Additionally, the jets alternate in their orientation by 90° to 180° (refer to Section 6.4) with the larger vortex streets favoring the 90° end of the spectrum and vice versa for the smaller vortex streets. The time span between the passage of two jets signifies one full period of the vortex street.

In Figure 6.8, the autocorrelation analyses for the control case are shown. It is clear to see that for this case the periods estimated by the two techniques yield very similar values of around 4 hours and also for all heights plotted. In addition, the periods for the $U$ and $V$ winds agree fairly well. To be concise, similar analysis for the other Madeira cases are not shown here, but their results are summarized by cases $M1 – M13$ in Table 6.2.

From Table 6.2, it is evident that increasing the island diameter resulted in essentially a monotonic increase in VKVS period, up to a particular threshold that is. Here $L$ represents the effective diameter of the islands (assumed to be the cross-stream diameter at the ABL
Figure 6.7: Time series of wind speed and direction corresponding to a location approximately one diameter downstream from the island and at 100 m msl. From top to bottom, the time series represent cases $M1 - M7$ shown in Table 6.2. The vertical lines represent 2 hour time intervals and the magnitude of the wind speed is denoted by the relative length of the vectors and their colors (i.e., cool colors represent weak wind and hot colors represent strong wind).

inversion), $P$ is the measured vortex period, and $Re$ and $Sr$ are the dimensionless Reynolds and Strouhal numbers. For cases $M1 - M5$, VKVS signatures were very obvious as the wind direction made regular oscillations, and 2-dimensional plan view images of the systems (not shown here) clearly showed repeating patterns of vortices downwind of the islands. On the other hand, for cases $M6$ and $M7$ no von Kármán type vortex streets were ever observed. In these cases, the vortices shed behind each flank of the island (i.e., vortex train) never appeared to
Figure 6.8: Fourier spectral analysis of the $U$ (upper left) and $V$ (upper right) wind components for the control case and autocorrelation function analysis (bottom). Multiple lines of the same color and symbol represent analysis for different heights above the surface. The green lines represent peaks in spectral energy and the analyzed energy frequency.

grow large enough to interact with the opposite train. This leads one to believe that perhaps this is near the upper $Re$ limit for VKVS formation in the atmosphere. Nevertheless, for cases $M1 - M5$ VKVS periods of 2-9 hours were observed which led to Strouhal numbers in the range of 0.15 to 0.22 for all cases. Interestingly, this is very similar to what has been documented for lower Reynolds number flows containing VKVSs. For case $M5$, $N$ was difficult to identify as this Reynolds number appeared to fall within a sort of transition zone; in this case the island wake appeared to alternate between a definitive von Kármán vortex street and a non-coherent wake structure leading one to question the validity of the $Sr$ estimate for this case.

Additionally, the heights of the idealized islands were varied both greater than and less than the original Madeira maximum height of 1800 m msl while the base diameter was kept 40 km (cases $M8 - M10$ in Table 6.2). These new heights reached 1400, 2200, and 2600 m msl; however, interestingly we observed little differences in periodicity between these simulations and the corresponding simulation of 40 km diameter, $H = 1800$ m idealized Madeira case. For a more in-depth perspective of this observation, we also conducted additional simulations.
Table 6.2: Simulation diagnostics for quasi-idealized cases assuming a length scale defined by the island diameter at the ABL inversion (\(L\)) and constant wind velocity scale of 10 m s\(^{-1}\). \(L^*\), \(H^*\), and \(P^*\) represent the control simulation island diameter, height, and observed period, respectively (i.e., \(L^* \approx 24\) km, \(H^* = 1800\) m, and \(P^* = 3.93\) h). \(Re_K\) is the modified Reynolds number computed by substituting eddy viscosity for molecular viscosity. Finally, \(M\), \(G\), and \(JF\) represent simulations of Madeira, Guadalupe, and Juan Fernández events, respectively.

<table>
<thead>
<tr>
<th>Case</th>
<th>(H_c) (m)</th>
<th>(L/L^*)</th>
<th>(H/H^*)</th>
<th>(Re)</th>
<th>(Re_K)</th>
<th>(P) (h)</th>
<th>(P/P^*)</th>
<th>(Sr)</th>
</tr>
</thead>
<tbody>
<tr>
<td>M1</td>
<td>1212</td>
<td>0.5</td>
<td>1.0</td>
<td>1.2(\times)10(^{10})</td>
<td>48900</td>
<td>2.20±.00</td>
<td>0.56±.00</td>
<td>0.150±.000</td>
</tr>
<tr>
<td>M2</td>
<td>1212</td>
<td>1.0</td>
<td>1.0</td>
<td>2.4(\times)10(^{10})</td>
<td>97800</td>
<td>3.93±.10</td>
<td>1.01±.03</td>
<td>0.185±.005</td>
</tr>
<tr>
<td>M3</td>
<td>1212</td>
<td>1.5</td>
<td>1.0</td>
<td>3.6(\times)10(^{10})</td>
<td>146700</td>
<td>5.85±.34</td>
<td>1.49±.09</td>
<td>0.170±.010</td>
</tr>
<tr>
<td>M4</td>
<td>1212</td>
<td>2.0</td>
<td>1.0</td>
<td>4.8(\times)10(^{10})</td>
<td>195600</td>
<td>7.45±.62</td>
<td>1.89±.16</td>
<td>0.185±.015</td>
</tr>
<tr>
<td>M5</td>
<td>1212</td>
<td>2.5</td>
<td>1.0</td>
<td>6.0(\times)10(^{10})</td>
<td>244500</td>
<td>8.25±.75</td>
<td>2.10±.19</td>
<td>0.205±.015</td>
</tr>
<tr>
<td>M6</td>
<td>1212</td>
<td>3.5</td>
<td>1.0</td>
<td>8.4(\times)10(^{10})</td>
<td>342300</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>M7</td>
<td>1212</td>
<td>5.0</td>
<td>1.0</td>
<td>1.2(\times)10(^{11})</td>
<td>489000</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>M8</td>
<td>812</td>
<td>0.9</td>
<td>0.8</td>
<td>2.2(\times)10(^{10})</td>
<td>88020</td>
<td>3.65±.32</td>
<td>0.93±.08</td>
<td>0.167±.015</td>
</tr>
<tr>
<td>M9</td>
<td>1612</td>
<td>1.1</td>
<td>1.2</td>
<td>2.6(\times)10(^{10})</td>
<td>107580</td>
<td>4.17±.17</td>
<td>1.06±.04</td>
<td>0.176±.010</td>
</tr>
<tr>
<td>M10</td>
<td>2012</td>
<td>1.1</td>
<td>1.4</td>
<td>2.6(\times)10(^{10})</td>
<td>107580</td>
<td>4.15±.18</td>
<td>1.06±.05</td>
<td>0.193±.025</td>
</tr>
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<td>M11</td>
<td>812</td>
<td>1.8</td>
<td>0.8</td>
<td>4.3(\times)10(^{10})</td>
<td>2152080</td>
<td>7.37±.70</td>
<td>1.88±.18</td>
<td>0.167±.015</td>
</tr>
<tr>
<td>M12</td>
<td>1612</td>
<td>2.2</td>
<td>1.2</td>
<td>5.3(\times)10(^{10})</td>
<td>2630320</td>
<td>8.08±.25</td>
<td>2.06±.07</td>
<td>0.182±.005</td>
</tr>
<tr>
<td>M13</td>
<td>2012</td>
<td>2.3</td>
<td>1.4</td>
<td>5.5(\times)10(^{10})</td>
<td>2749880</td>
<td>8.34±.84</td>
<td>2.12±.21</td>
<td>0.190±.015</td>
</tr>
<tr>
<td>G1</td>
<td>1274</td>
<td>1.0</td>
<td>1.0</td>
<td>4.0(\times)10(^{10})</td>
<td>98700</td>
<td>4.50±.33</td>
<td>1.15±.09</td>
<td>0.160±.010</td>
</tr>
<tr>
<td>JF1</td>
<td>1244</td>
<td>0.5</td>
<td>1.0</td>
<td>1.2(\times)10(^{10})</td>
<td>417391</td>
<td>1.92±.25</td>
<td>0.50±.05</td>
<td>0.175±.025</td>
</tr>
</tbody>
</table>

of all three new island heights but with 80 km diameters (cases \(M11 – M13\), Table 6.2). In these cases, we also observed the shedding frequency to demonstrate little variability between each simulation of varying height and to be very similar to case \(M4\). Essentially, these results indicate that the island diameter represents the dominant, intrinsic length scale of the system (more so than the height) and thus primarily drives the \(Sr - Re\) relationship. Fundamentally, this is expected as the mesoscale VKVS phenomenon is generally assumed to be the atmospheric analog of classical 2-D VKVSs. In essence, this assumption of two-dimensionality is supported
by our findings here. The two-dimensional nature of VKVSs is also evident in the velocity spectra (Figure 6.8 - top panels). The -3 spectral scaling is a hallmark for 2-D turbulence \[134, 204\]. A detailed spectral characterization of VKVSs is beyond the scope of this chapter. However, we would like to point out that a -5/3 regime (i.e., inertial range) was also evident in some of the simulations in addition to (sometimes in lieu of) the -3 spectral regime. These results will be reported in a future publication.

Based on the concept of two-dimensionality, the thermally stable layer between the ABL inversion and the island dividing streamline acts as the critical zone harboring lateral flow around the island, thus fueling the VKVSs. From this point, one may consider that the maximum horizontal length scale of the island, within this critical zone, is related to the largest and most dominant scales of motion in the VKVS. Additionally, islands which typically instigate VKVSs are, roughly speaking, cone-shaped which leaves the largest diameter within the critical zone to be at the ABL inversion height. This island diameter can be thought of as the effective length scale of the VKVS system which dictates the \( Sr - Re \) relationship. This hypothesis is supported by the results found here as the formulation used here for creating idealized island terrain height lead to little change in island diameter at the ABL height when changing maximum island height.

![Figure 6.9](image)

Figure 6.9: Upstream planar averaged vertical profiles of eddy viscosity for the Madeira event (left), the Guadalupe event (middle), and the Juan Fernández event (right). Times are valid for July 5th, 2002.

Table 6.2 also introduces a modified Reynolds number \( Re_K \) which is formulated the same as the traditional \( Re \) except that eddy viscosity \( K_M \) is used in lieu of molecular viscosity. For atmospheric applications, the use of \( Re_K \) is perhaps more appropriate than the use of \( Re \) considering the fact that atmospheric flows are dominated by turbulent diffusion over molecular diffusion by many orders of magnitude. With regards to mesoscale VKVSs, the use of \( Re_K \) has
been occasionally used in previous studies [81, 133, 223]. In the present work, eddy viscosity within the ABL was volume averaged throughout the upstream portion of the domain. Then, these $K_M$ values were averaged in time throughout the simulation period and incorporated into Table 6.2. Figure 6.9 illustrates the mean profiles of $K_M$ for each event. Using mean $K_M$ values we find Reynolds numbers ranging from $\approx 5 \times 10^4$ to $2.7 \times 10^6$; comparing our $Sr$ measurements with laboratory measurements of VKVSs in this $Re$ range we find close similarities. This implies that the periodicity of atmospheric VKVSs behave similar to classical VKVSs when the concept of eddy viscosity is employed.

6.5.2 Guadalupe and Juan Fernández Events

The conclusions from cases M1 – M13 in Table 6.2 directly support the hypothesis that the dimensionless period of mesoscale VKVSs is essentially not a function of a vertical length scale and instead primarily a function of a horizontal length scale and a velocity scale. While the concept of a universal $Sr$ number for atmospheric VKVSs, virtually irrespective of $Re$, is an intriguing conclusion; unfortunately, the data presented here are too imprecise to make such a bold conclusion. Nevertheless, the information displayed in Table 6.2 does seem to indicate that the dimensionless periods of mesoscale VKVSs do fall in a relatively compact range (i.e., 0.15-0.22). In order to support this finding, we considered the Guadalupe and Juan Fernández events previously described and compared their simulated shedding frequency to that of the quasi-idealized Madeira cases (G1 and JF1).

Although some free-stream conditions were similar between all three events, as typical for VKVS formation (i.e., a thermal inversion below the crest of the island, steady free-stream wind flow from a constant direction, etc.), these simulations had entirely different initial and boundary conditions and thus acted as acceptable trial candidates for testing the universality of the atmospheric $Sr – Re$ relationship. In Figure 6.10, time-height plots of potential temperature profiles can be seen for cases G1 and JF1. Comparing these figures to what was illustrated in Sec. 6.4, it can be noted that the Guadalupe event had a slightly lower thermal inversion from the Madeira event. This increased the effective length scale to about 65% of the base diameter of the idealized island, compared to 60% for the Madeira cases. Alternatively, the Juan Fernández event had a much higher thermal inversion height than the control Madeira event. For this case the effective length scale was about 30% of the island base diameter which was about half that of the control Madeira case. With these length scales in mind, we would expect case G1 to have a slightly longer period than the control case (M2) and case JF1 to have a period of about half that of the control case in order to keep $Sr$ generally within the same range as found for the Madeira case. This is indeed what was found. It should be pointed out that these additional simulations (cases G1 and JF1) had free-stream wind speeds similar
to the Madeira case of near 10 m s\(^{-1}\) at the boundary layer height. This means that we did not test scenarios where the velocity scale was significantly higher or lower than 10 m s\(^{-1}\); but on the other hand, we were not aware of any VKVS events which occurred in extremely higher or lower wind circumstances.

Figure 6.10: Upstream potential temperature profiles for the Guadalupe (left panel) and Juan Fernández (right panel) events corresponding to hours 24 - 36 of the simulations.

6.6 Discussion

In this chapter, we have used the WRF model as a tool to understand the behavior of quasi-idealized VKVSs induced by mountainous islands within the marine atmospheric boundary layer. While the WRF model underestimated the length of these particular streets, other characteristics of the VKVSs were qualitatively, and somewhat quantitatively, similar to observations; namely the vortex spacing and consequently shedding frequency appeared realistic. The period was estimated for the control case to be about 4 hours yielding a Strouhal number of approximately 0.185 when assuming a velocity scale of 10 m s\(^{-1}\) and a length scale defined by the island diameter at the height of the thermal inversion (i.e., about 24 km). Furthermore, subsequent experimentation conducted by systematically adjusting the length scale (i.e., \(Re\)) of the islands produced little identifiable change in this value of \(Sr\) while within a \(Re\) regime conducive to VKVS formation. This indicates that the period of mesoscale atmospheric VKVSs is a direct and linear function of \(L\). Also, it was found that the island diameter, specifically at the height of the ABL inversion, was the dominant length scale (\(L\)) associated with VKVS frequency modification. This conclusion was supported by the fact that (based on the island
shape equation used here; Eq. 6.1) changing the maximum height of the island resulted in little change to the island diameter at the ABL inversion and therefore the frequency of the VKVS remained generally constant. In addition, numerical analysis of two other, very unique cases conformed to the $Sr - Re$ relationship found for the control event. This supports the quasi-universality of the range of $Sr$ put forth in Table 6.2 for mesoscale atmospheric flows.

Although somewhat broad, the range of Strouhal numbers put forth here for mesoscale VKVSs generally resembles what has been observed in laboratory flows of much lower Reynolds number. The weakly fluctuating values of $Sr$ with respect to $Re$ leads the authors to believe that atmospheric VKVSs behave similarly to VKVSs in the range of $10^2 < Re < 10^4$ studied by Roshko [191]. If this is indeed the case, one might conclude that the use of eddy viscosity, as opposed to molecular viscosity, could be used as an instrument to conceptualize similarity theories gleaned from laboratory VKVSs in the context of atmospheric VKVSs.

It should be noted that while the $Sr$ measurement method used here provided a reasonable assessment of the general $Sr - Re$ relationship for atmospheric VKVSs, it was limited by the fact that the selection of the reference point was a somewhat subjective process. Identifying a downstream point that is perfectly in the center of the wake is a difficult task and additionally the wake itself may not be symmetrical nor entirely steady throughout the day. These slight asymmetries and time dependencies may arise from changes in wind direction or ABL height and eventually contribute to the uncertainty of the values presented in Table 6.2.

On a final note, understanding the behavior of vortex shedding frequency in atmospheric flows is another step forward in identifying and understanding the universal $Sr - Re$ similarity relationship. From this new knowledge, we can better understand the placement of embedded wake vortices as a function of time and the spacing between atmospheric vortices. From a practicality perspective, the ability to estimate the vortex shedding frequency, and vortex spacing, from commonly known environmental parameters may prove advantageous through a number of applications varying from optical wave propagation [230] to long-range pollutant dispersion [84].

### 6.7 Future Directions

Moving forward, future studies should be carried out to explicitly assess mesoscale model sensitivity to physical parameterization schemes (surface layer, planetary boundary layer, etc.) and validity when simulating atmospheric VKVSs. This will help to identify any model bias associated with defining the $Sr - Re$ relationship in the turbulent atmosphere. Such assessments will be challenged by limited observational data; however, evolving remote sensing technology such as stereo satellite-derived winds at cloud height may offer unique opportunities to address this issue [87]. Additionally, more advanced numerical modeling techniques, such as large-eddy
simulation (LES), can be used to study atmospheric VKVSs on the meteorological microscale. Perhaps studies of this nature could provide better insight into the bridge between atmospheric and classical vortex streets. To the knowledge of the authors, only one such study ([81]) has been published which was devoted to investigating the structure of von Kármán vortices produced by somewhat smaller obstacles than those in this study. In that work, the authors calculated slightly lower $Sr$ values than we have here, near 0.12 to 0.13; this provokes the question that perhaps the dynamics of atmospheric VKVSs varies throughout the meso-micro scale transition regime.

Also, LES should produce a more realistic turbulence regime in the lee of the island and possibly provide insight into interaction of turbulence and mesoscale dynamics in this region. In addition, projects should be undertaken to more accurately measure the vortex shedding frequency, possibly by using meteorological measurements as benchmarks [87]. The current study is limited by the fact that the only validation of simulated $Sr$ against realistic $Sr$ was possible by comparing the spatial distances between eddies in the satellite imagery to the simulated eddy patterns. This weakness, combined with the unavoidable limitations of numerical modeling, introduces an important source of uncertainty which will need to be addressed in future studies before an atmospheric $Sr – Re$ similarity relationship can be solidified.
Chapter 7

Spectral Properties of Low-Froude Number Flow Past Multi-Island Archipelagos and Effects on Gap-Jet

7.1 Introduction and Background

Within the field of fluid mechanics, a large body of literature is evolving which focuses on the interaction of side-by-side vortex streets at low Reynolds numbers (see Zdravkovich [246] and Sumner [218] for comprehensive reviews). In the mid 1980s, Williamson [240] examined distinct classes of harmonic modes associated with varying gap-ratio (i.e., the gap between the centers of two identical cylinders divided by the diameter of one of the cylinders). In connection with these harmonics, various vortex street synchronization regimes have been documented in experiments using both particle image velocimetry [8] and direct numerical simulation [107]. Furthermore, laboratory studies have demonstrated that under conditions of flow past two side-by-side identical bluff bodies, that multiple vortex shedding frequencies can co-exist and that these frequencies are a function of the dominant length scales of the system [219]. In essence, the existence of more than one shedding frequency is associated with the effective length scales of each of the single bodies in addition to the effective length scale of both bluff bodies acting as one. Based on this understanding, the prevalence of individual frequencies is intuitively a function of gap-ratio (GR). That is, smaller gap-ratios tend to induce a stronger low frequency signal while the higher gap-ratios are correlated with a stronger high frequency signal [219]. This correlation exists due to the direct relationship between GR and the size of the gap between the two bodies [232]. With a large gap comes greater disruption of the large amplitude vortex street and vise versa.
In addition to the relationship between GR and the spectral properties of the wake region, as previously mentioned this relationship is also connected with varying vortex street interaction regimes. At small gap ratio (i.e., when the two bodies are close to each other) the wake pattern essentially behaves as if the two bodies are a single bluff body [44]. This vortex street wake pattern is known as a single bluff-body wake. As the gap ratio increases, the wake pattern enters a flip-flopping phase where the wake vortex streets transition back and forth between a single bluff-body wake and that of two distinct wake patterns [136]. Then, at even higher gap ratios, two unique vortex street synchronization regimes can develop. These include in-phase and anti-phase synchronization. Various studies have debated the predominance of in-phase over anti-phase synchronization, and vice versa, over the years. However, much is still unknown about what actually dictates the development of one of the synchronization phases over the other.

Additionally, with each gap ratio, and phase of vortex street interaction, the flow between two bluff bodies can exhibit substantially different behavior. At small gap ratios, when single bluff-body wakes prevail, the downstream wake behind one body tends to dominate over that of the other body [8, 107]. In these cases, the gap jet between two bodies may have a skewed, or biased, orientation off of the large scale stream flow (see Figure 2.6 right panel). This orientation bias is known to fluctuate over time [8]. Also, at larger gap ratios in the flip-flopping wake regime, the gap jet may have pulsing behavior both in magnitude and orientation [41]. It is expected that if similar vortex street interaction regimes exist in the atmosphere then similar gap jet behavior can also be expected [175]. Consequently, gap jet fluctuations on atmospheric scales can have implications for orographic gap winds [172]; for example, in natural atmospheric channels between two islands (e.g., Alenuihaha Channel).

Despite this extensive literature base, the topic of high Reynolds number, geophysical scale interactive vortex streets has been virtually unexplored. However, it has been shown that the relationship between GR and vortex street phase can be a function of Reynolds number [243]. In addition, the atmospheric analog of the problem of multiple vortex street interaction induced by two side-by-side bluff bodies has important relevance to both science as well as society in many forms. For example, the wind forecasting community could benefit from a deeper understanding of the meandering of gap-jets between multiple islands and the physical oceanography community could benefit from a deeper understanding of the lifetime of far-field von Kármán vortices as they pertain to sea-surface stresses. Furthermore, there are several island archipelagos that exist around the world which produce this phenomenon (e.g., Aleutian Islands, Canary Islands). At the same time, aside from differences in Reynolds number, atmospheric vortex streets have additional complexity over those previously studied in laboratory settings as they also have an additional dimension (i.e., the vertical dimension). That being said, it is unknown whether or
not the observations and theories gleaned from classical multi-vortex street interaction can be extended to atmospheric events.

In this work, we present results from numerical simulations of flow past multiple side-by-side islands in order to explore the properties of interactive vortex streets. Through a suite of experiments, we document the characteristics of atmospheric vortex street synchronization and its implications on low-level winds. More specifically, we focus on the spectral features of the simulated vortex streets by cataloging the relationship between gap-ratio and the frequencies of spectral peaks. In addition, we comment on the dynamical properties of the associated gap-jets and their (dis)similarities to 2-dimensional, low Reynolds number studies.

7.2 Methodology

In order to investigate the interactive behavior of atmospheric vortex streets and compare them to low-Reynolds number observations, we perform a suite of experimental simulations. Primarily, the simulations are intended to test the relationship between GR and the phase of multi-vortex street interaction. Here we evaluate 4 different gap-ratios (i.e., 1.0, 1.5, 2.0, and 3.0), and two different island diameters (i.e., 50 km, and 75 km). Additionally, we do tests for two different sets of boundary conditions to investigate whether our results hold for variable boundary conditions. Table 7.1 outlines the simulations discussed in this work. For each test, the simulations employed quasi-stationary boundary conditions in order to ensure that the angle of approaching wind was consistently (approximately) perpendicular to the axis of the islands alignment with one another. That being said, the simulations were run for 48 hours in total with the final 36 hours being used for analysis. Identical boundary conditions were prescribed at 00, 24, and 48 hours into the simulations and therefore the only transient forcing throughout the simulations were variations due to diurnal effects. The two different sets of boundary conditions which were used here corresponded to the Canary Islands on July 5th, 2002 (i.e., C02) and Madeira Island on July 10th, 2010 (i.e., M10). These particular events have been previously discussed in other chapters. Similar to Chapter 6, here we prescribe an idealized pair of islands side-by-side with peak heights of 2 km msl.

The numerical simulations performed here employ similar configurations to those used in the previous chapters. That is, the WRF model was ran using ERA-Interim boundary conditions and identical physics parameterizations to those used throughout this research (i.e., shown in Table 4.1). The grid mesh used 4 domains with a highest resolution of 1 km horizontally and one-way nesting. The orientation of the grid was aligned roughly with the angle of approach of boundary layer flow and subsequent vortex propagation axis (Figure 7.1). d04 had horizontal dimensions of 799 by 799 grid cells and the center point of the prescribed islands lied along the 525th grid row from the lower boundary. The center point between the two islands was at
Table 7.1: Simulation experiments

<table>
<thead>
<tr>
<th>Test</th>
<th>D</th>
<th>GR</th>
<th>Date</th>
</tr>
</thead>
<tbody>
<tr>
<td>C02 GR1.0 50km</td>
<td>50 km</td>
<td>1.0</td>
<td>July 5th, 2002</td>
</tr>
<tr>
<td>C02 GR1.5 50km</td>
<td>50 km</td>
<td>1.5</td>
<td>July 5th, 2002</td>
</tr>
<tr>
<td>C02 GR2.0 50km</td>
<td>50 km</td>
<td>2.0</td>
<td>July 5th, 2002</td>
</tr>
<tr>
<td>C02 GR3.0 50km</td>
<td>50 km</td>
<td>3.0</td>
<td>July 5th, 2002</td>
</tr>
<tr>
<td>C02 GR1.0 75km</td>
<td>75 km</td>
<td>1.0</td>
<td>July 5th, 2002</td>
</tr>
<tr>
<td>C02 GR1.5 75km</td>
<td>75 km</td>
<td>1.5</td>
<td>July 5th, 2002</td>
</tr>
<tr>
<td>C02 GR2.0 75km</td>
<td>75 km</td>
<td>2.0</td>
<td>July 5th, 2002</td>
</tr>
<tr>
<td>C02 GR3.0 75km</td>
<td>75 km</td>
<td>3.0</td>
<td>July 5th, 2002</td>
</tr>
<tr>
<td>M10 GR1.0 50km</td>
<td>50 km</td>
<td>1.0</td>
<td>July 9th, 2010</td>
</tr>
<tr>
<td>M10 GR1.5 50km</td>
<td>50 km</td>
<td>1.5</td>
<td>July 9th, 2010</td>
</tr>
<tr>
<td>M10 GR2.0 50km</td>
<td>50 km</td>
<td>2.0</td>
<td>July 9th, 2010</td>
</tr>
<tr>
<td>M10 GR3.0 50km</td>
<td>50 km</td>
<td>3.0</td>
<td>July 9th, 2010</td>
</tr>
</tbody>
</table>

approximately the 450th grid column from the left.

7.3 Vortex Street Phasing Regimes

To study the phasing regimes associated with each gap-ratio, we first examine instantaneous U wind speed horizontal cross-sections at the lowest vertical grid level for the C02 boundary conditions (i.e., the C02 cases will be the control cases in this chapter). In Figure 7.2, the U wind speed (in this case lateral wind speed) is shown 24 hours after model initialization and the differences in simulated wake patterns can be clearly seen. In the GR1.0 case, a faintly identifiable single bluff-body wake can be seen as denoted by the broad regions of transverse momentum. On the other hand, in the GR1.5 case variability in the lateral wind component with respect to downstream location in the vortex street exemplified a flip-flopping phase (i.e., temporal fluctuation between a single bluff-body wake and a synchronized wake). Finally, in the GR2.0 and GR3.0 cases anti-phase and in-phase synchronization appears most prominent (respectively) as the lateral wind speed anomalies have an inverse phase at any given point between the two side-by-side vortex streets. While the instantaneous patterns presented in
Figure 7.1: Domain configuration and schematic illustrating the adjustments made to d04. \( L \) represents the distance between the centers of the two identical islands and \( D \) represents the diameter of one island. The red arrow indicates the approximate direction of advection within the ABL.

Figure 7.2 generally agree with the PIV measurements taken by Akilli et al. [8] for flow around 2 side-by-side cylinders at low Reynolds number, the wake regimes can be much better visualized in this case by examining Hovmoller diagrams.

A Hovmoller diagram enables one to compare the spatio-temporal properties of a transient phenomena. Here, we select a specific cross-section that is approximately 1 island diameter downstream of the island pairs and oriented laterally across the background flow as shown in Figure 7.3. This cross-section, or trajectory, is taken for total horizontal windspeed at the lowest grid level height and then plotted in time for each of the GR1.0 through GR3.0 cases (Figure 7.4).

Figure 7.4 presents Hovmoller diagrams of horizontal wind speed for each case. In the GR1.0 case it can be seen that the zone of low wind speed along the Hovmoller transect is significantly more disorganized than the three other cases. That is, there is a lack of a coherent vortex street or pair of parallel vortex streets. However, moving to GR1.5 it can be seen that two individual vortex streets begin to appear. Yet still, at all times there co-exists both dominant and less dominant vortex streets as is evidenced by the differences in the size and shape of the weak wind speed zone trailing behind each island. Furthermore, case GR1.5 appears to fall
Figure 7.2: Instantaneous plan view of surface zonal wind velocity for C02 GR1.0 D50km (upper left), C02 GR1.5 D50km (upper right), C02 GR2.0 D50km (lower left), and C02 GR3.0 D50km (lower right) experiments.

in the flip-flopping wake regime given the hand-off of wake dominance from one vortex street to the other (e.g., around hour 12). On the other hand, cases GR2.0 and GR3.0 demonstrate clear synchronization which does not vary with time. As further confirmation of the features identified in the full horizontal wind speed Hovmoller diagrams, the corresponding U wind speed Hovmoller diagrams are shown in Figure 7.5.

Thus far in this chapter we have used qualitative techniques to classify the vortex street phasing regimes. However, because we have taken a numerical modeling approach we have the capability to also quantitatively catalog the vortex street phasing associated with each gap-ratio. To this end, we use cross-correlations associated with two points along the Hovmoller transect.
directly behind the center of each island. In Figure 7.6 the cross-correlations corresponding to the U wind speed in each GR case are shown. Starting first with the GR1.0 case, it can be seen that a positive, or in-phase, correlation exists as denoted by the initial positive cross-correlation at 0 h lag time. Furthermore, the lowest frequency is around 5 hours. However, for the GR1.5 case the correlation was slightly negative with a non-decaying frequency magnitude with respect to lag time. This is quite possibly an artifact of the apparent flip-flopping behavior present in the GR1.5 case. Looking next at the GR2.0 case, it can be seen that again there is an anti-phase cross-correlation yet it is a much cleaner distribution. That is, there appears to be one singular frequency contributor as the correlation decreases steadily with increasing lag time. For comparison, in the GR1.5 case there appears to be multiple frequency contributors as there is a surge in correlation energy at both 5 hrs and about 10 hrs lag times which seemingly corresponds to the flip-flopping behavior previously mentioned. Finally, the GR3.0 case is very similar to the GR2.0 case; however, the phase is in-phase as opposed to anti-phase.

Collectively, the results presented by the cross-correlation analysis corroborate what was alluded to in the Hovmoller diagrams previously introduced. However, more importantly the phasing regimes found here for the most part represent the phasing regimes documented for low Reynolds number 2-dimensional flows [218]. To be more specific, at $\text{GR} \leq 1.4$, $1.4 \leq \text{GR} \leq 1.5$,
Figure 7.4: Horizontal wind speed Hovmoller diagram taken from a surface transect perpendicular to the ambient flow approximately 1 island diameter downstream for the C02 GR1.0 D50km (upper left), C02 GR1.5 D50km (upper right), C02 GR2.0 D50km (lower left), and C02 GR3.0 D50km (lower right) experiments.
Figure 7.5: U wind component wind speed Hovmoller diagram taken from a surface transect perpendicular to the ambient flow approximately 1 island diameter downstream for the C02 GR1.0 D50km (upper left), C02 GR1.5 D50km (upper right), C02 GR2.0 D50km (lower left), and C02 GR3.0 D50km (lower right) experiments.
GE ≥ 1.5, and GR ≥ 3.0; Kang [107] observed single bluff body, flip-flopping, in-phase, and anti-phase vortex street interaction phases, respectively. While the bifurcation regions found by Kang [107] generally agree with what is presented here; it should be noted that the present simulations tended to produce in-phase synchronization at GR3.0 and anti-phase synchronization at GR2.0 where other studies have found the reverse behavior. At the same time, in the GR2.0 case presented here we observed both in-phase and anti-phase synchronization when testing two different island diameters. This conforms to what has been found for many other low Reynolds number experiments when the in-phase and anti-phase synchronization regimes lacked a clear bifurcation and had overlapping GR regions.

Figure 7.6: U wind speed cross-correlation between two points directly behind both islands for C02 GR1.0 (upper left), C02 GR1.5 (upper right), C02 GR2.0 (lower left), and C02 GR3.0 (lower right) experiments. Black lines represents D50km cases while red lines represent D75km cases.
7.4 Energy Spectra

To better understand the implications of the aforementioned phasing regimes on the inherent dynamical nature of the vortex streets, the use of spectral analysis is warranted. In this section, we use wind speed time series from within the boundary layer at specific monitor locations located in the wake region to characterize the spectral properties of each GR case.

As mentioned early in this chapter, past experiments of multi-vortex streets in low-Reynolds number flows have yielded a dual spectral energy peak. Of note, Williamson [240] found dual spectral peaks for gap-ratios ranging from about 1.2 to 2.4 for flow of Reynolds number = 200 while Sumner et al. [219] found dual peaks for a slightly broader range of at least 1.1 to 2.5 for Reynolds numbers between 1200 and 2200. Both of these results were gleaned from laboratory experiments. The interpretation of these spectral features is that the lower frequency is associated with flow around both obstacles while the higher frequencies are associated with flow around individual obstacles. Obviously, the strength of the spectral energy is a function of GR with the smaller gap-ratios having a defined two-peak signal. Furthermore, the relative strength of one spectral peak over the other is a function of the relative dominance of large-scale vortex shedding versus small-scale vortex shedding. That being said, with increasing GR the lower frequency peak dissolves. In Sumner et al. [219], the Sr combinations found were 0.09 and 0.18, 0.10 and 0.40, 0.14 and 0.32, 0.17 and 0.25, and 0.21 for GR = 1.0, 1.5, 2.0, 2.5, and 3.0 respectively. Correspondingly, Williamson [240] recorded Sr on the same magnitude (i.e., about 0.09 to 0.2).

In the simulations presented here we identify several similar features to those mentioned above. First, looking at the U wind speed spectral density plot for the C02 D50km case (Figure 7.7), it can be seen that variability in the position of the peak spectral power density demonstrates a correlation with GR. For the C02 D50km experiments, generally with increasing GR comes increasing Sr (i.e., shorter periods). This is due to the fact that the effective length scale decreased as the distance between the islands increases. Also, the respective energy associated with the spectral peak tends to encompass a larger range of frequencies with lower GR. In other words, as GR increases the range of dominant frequencies becomes more narrower and the spectral slope becomes steeper. Moreover, the Sr-GR combinations were in the range of 0.21 to 0.24 which was very similar but slightly higher than what was found by Williamson [240] and Sumner et al. [219]. In this case, and in those to follow, Sr was computed by normalizing frequency by the diameter of the base of one of the islands divided by the upstream mean ABL wind speed (i.e., approximately 10 m/s for all cases).

Next, we examine the identical case to the C02 D50km case but with 50% wider islands which yields island base diameters of 75 km (Figure 7.8). Here a similar relationship between Sr and GR is visible; however, due to the fact that the island diameters have increased, the periods
Figure 7.7: U wind speed spectra from monitor points ranging from 2 - 3.5 island diameters downstream and 1-3 island diameters off of the centerline. C02 GR1.0 D50km (upper left), C02 GR1.5 D50km (upper right), C02 GR2.0 D50km (lower left), and C02 GR3.0 D50km (lower right) experiments.
Figure 7.8: U wind speed spectra from monitor points ranging from 2 - 3.5 island diameters downstream and 1-3 island diameters off of the centerline. C02 GR1.0 D75km (upper left), C02 GR1.5 D75km (upper right), C02 GR2.0 D75km (lower left), and C02 GR3.0 D75km (lower right) experiments.
have also increased relative to the D50km cases. In addition, the dual spectral peak previously observed in the GR1.5 case is not apparent. This could possibly be due to the definition of gap-ratio used here. That is, we have defined gap-ratio as the distance between island centers at mean sea level. However, based on the evidence provided in the previous chapter this ratio could also be defined based on the gap between islands at the inversion height. If the latter was done, then GR would slightly differ from what is shown here. Nevertheless, many of the same features commented on in the C02 D50km cases are also visible in the C02 D75km cases.

Figure 7.9: U wind speed spectra from monitor points ranging from 2 - 3.5 island diameters downstream and 1-3 island diameters off of the centerline. M10 GR1.0 D50km (upper left), M10 GR1.5 D50km (upper right), M10 GR2.0 D50km (lower left), and M10 GR3.0 D50km (lower right) experiments.
Figure 7.9 shows the same spectral analysis for the identical experiments but with the M10 simulation boundary conditions (i.e., July 9th, 2010). In the M10 cases the spectral peaks in each GR case can be easily identified; however, they are somewhat less pronounced compared to the C02 cases. This is likely due to the fact that the upstream wind speed was weaker in the M10 boundary conditions than in the C02 boundary conditions. In connection, the weaker upstream wind speed produced longer periods for all gap-ratios when compared to the C02 D50km cases. Despite this difference, the increasing Sr with increasing GR persists in the M10 D50km case which supports the Sr-GR relationship identified in the previous C02 D50km and C02 D75km cases. Essentially, the fact that this behavior exists with two different forcing boundary conditions provides evidence that perhaps the Sr-GR relationship is not unique to a particularly limited wind regime. Furthermore, it should be noted that in this case the range of Sr associated with the difference GR simulations is higher than in the C02 cases (i.e., 0.22 to 0.28). This difference could represent a combination of few different factors such as a Reynolds number dependency in the Sr-GR relationship or the level of uncertainty inherent to our simulations.

Finally, Figure 7.10 was generated identically to Figure 7.9 but instead of using quasi-stationary boundary conditions, these simulations used realistic time-dependent boundary conditions. Although the mean upstream wind speed fluctuated during the 48 hour simulation and produced variable vortex street periods. It can generally be seen that the same Sr-GR relationship behaves similarly to what was observed for the simulations which used quasi-stationary boundary conditions. This helps to strengthen confidence in the effectiveness of the quasi-idealized boundary condition methodology used in this chapter.

A summary of the results presented in this section are shown in Table 7.2. In this table, only the Sr associated with the highest magnitude power spectral density peak is presented; however, as shown throughout this section there were many cases in which dual peaks were identified. For the most part, the range of dimensionless vortex street shedding frequencies was aligned with what has been found in laboratory experiments of Williamson [240] and Sumner et al. [219] for low Reynolds number (i.e., Re = 200 - 2200) vortex street pairs (i.e., 0.09 ≤ Sr ≤ 0.22).

7.5 The Gap Jet

Building on the previous two sections, the behavior of the gap-jet that forms between two bluff bodies immersed in turbulent flow is closely associated with the interaction of the two trailing vortex streets. At the low gap-ratios the gap-jet axis has been shown to favor one side of the wake region with respect to the ambient upstream flow direction [8]. This asymmetry, which occurs in cases of flow past two of more side-by-side obstacles, causes one side of the vortex street to dominate over the other. Under this situation, the skewed gap-jet’s orientation can
Figure 7.10: U wind speed spectra from monitor points ranging from 2 - 3.5 island diameters downstream and 1-3 island diameters off of the centerline using realistic time-dependent boundary conditions as opposed to quasi-stationary boundary conditions. M10 GR1.0 D50km (upper left), M10 GR1.5 D50km (upper right), M10 GR2.0 D50km (lower left), and M10 GR3.0 D50km (lower right) experiments.
Table 7.2: Simulation experiments and observed frequencies.

<table>
<thead>
<tr>
<th>Test</th>
<th>Phase</th>
<th>Sr</th>
<th>P</th>
</tr>
</thead>
<tbody>
<tr>
<td>C02 GR1.0 50km</td>
<td>Single Wake</td>
<td>0.21</td>
<td>5.2 h</td>
</tr>
<tr>
<td>C02 GR1.5 50km</td>
<td>Flip-Flopping</td>
<td>0.21</td>
<td>5.2 h</td>
</tr>
<tr>
<td>C02 GR2.0 50km</td>
<td>Anti-Phase</td>
<td>0.24</td>
<td>4.5 h</td>
</tr>
<tr>
<td>C02 GR3.0 50km</td>
<td>In-Phase</td>
<td>0.24</td>
<td>4.5 h</td>
</tr>
<tr>
<td>C02 GR1.0 75km</td>
<td>Single Wake</td>
<td>0.23</td>
<td>7.3 h</td>
</tr>
<tr>
<td>C02 GR1.5 75km</td>
<td>Flip-Flopping</td>
<td>0.23</td>
<td>7.3 h</td>
</tr>
<tr>
<td>C02 GR2.0 75km</td>
<td>Anti-Phase</td>
<td>0.28</td>
<td>6.1 h</td>
</tr>
<tr>
<td>C02 GR3.0 75km</td>
<td>In-Phase</td>
<td>0.28</td>
<td>6.1 h</td>
</tr>
<tr>
<td>M10 GR1.0 50km</td>
<td>Single Wake</td>
<td>0.22</td>
<td>6.3 h</td>
</tr>
<tr>
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<td>5.6 h</td>
</tr>
<tr>
<td>M10 GR2.0 50km</td>
<td>Anti-Phase</td>
<td>0.28</td>
<td>5.0 h</td>
</tr>
<tr>
<td>M10 GR3.0 50km</td>
<td>In-Phase</td>
<td>0.28</td>
<td>5.0 h</td>
</tr>
</tbody>
</table>

meander across the wake region which manifests as the flip-flopping phase previously described in the low Reynolds number literature.

To study the phenomenon of meandering gap-jets in an atmospheric context, we first examine a plan view depiction of the temporal evolution of the near surface wind speeds for the C02 D50km GR1.0 case (Figure 7.11). As can be seen in the top left panel, the 36 hour mean gap-jet orientation is approximately aligned with the background wind direction. However, looking at three individual 12 hour mean wind plots taken from the full 36 hour analysis period, the non-stationarity of the gap-jet with time can be clearly observed. This non-stationarity is evidence of the gap-jet meandering found in laboratory studies (e.g., [8]); however, in this case the meandering frequency is so low relative to the vortex street shedding frequency that a mostly single bluff body wake is observed (see Figure 7.4).

For a more in-depth look at the gap-jet meandering found in the GR1.0 case, we look to instantaneous wind speed fields. In Figure 7.12 the instantaneous horizontal wind speed fields can be seen for the GR1.0 case at an interval of 4.5 hrs. From this figure, not only can the meandering of the gap-jet be seen, but also the correlation between the gap-jet orientation and
Figure 7.11: 36 hour mean gap winds (upper left) and first (upper right), middle (lower left), and last (lower right) 12 hour mean gap winds. Plot corresponds to winds at the lowest vertical grid level in the C02 GR1.0 D50km case.
the low wind speed asymmetry in the wake region. In other words, as the gap-jet shifts its axis to one side of the wake region the opposite side grows a broader recirculation zone. While this builds the pressure perturbations associated with vortex street initiation, the constant gap-jet presence and variability effectively prevents the development of a vortex street pattern.

Moving forward, we now investigate the properties of the simulated gap-jets with respect to the gap-ratio in the C02 D50km cases. To do so we again utilize Hovmoller diagrams, but this time oriented along the streamwise direction as shown in Figure 7.13. In Figure 7.14, Hovmoller diagrams from the gap between the islands is shown. As expected, the magnitude of the gap-jet in the GR1.0 case was consistently greater than each of the other cases (i.e., GR1.5 and GR2.0) as a result of the more potent funneling effect induced by the narrow gap between the two islands. Also, the wind shear across the gap-jet entrance region is significantly higher in the GR1.0 case than in the other 2 cases. The gap-jet signature was not present in the GR3.0 case and so that case has been neglected from the analysis presented here.

Also in Figure 7.14, the intensity of the gap-jet along the Hovmoller transect exhibits a pulsing behavior with the frequency of the pulses being correlated with the gap-ratio. That is, the gap-jet pulsing has a period on the order of tens of hours for the GR1.0 case while it is approximately 5 hours for the GR1.5 case. The higher frequency pulsing behavior present in the GR1.5 case is essentially what produces the flip-flopping behavior shown in Figure 7.4. That being said, the GR1.0 and GR1.5 cases are very similar except for the fact that the gap-jet meandering frequency is much higher in the GR1.5 case.

Collectively, the meandering nature of the gap-jets found the simulations presented here reflect what has been observed in low Reynolds number flow around 2-dimensional cylinders. Furthermore, the gap-jet behavior agrees with what was observed for the vortex street phasing as a function of GR shown in Section 7.3. Nevertheless, as is the case in low Reynolds number experiments the primary physics which dictate the frequency of the gap-jet meandering is mostly elusive. However, a new paper by Carini et al. [41] provides evidence that the secondary frequency associated with the flip-flop vortex street phase, and similarly the gap-jet meandering, originates as a result of an inherent instability associated with the in-phase regime which materializes at low GR.

### 7.6 Summary

In this work, we have evaluated the dynamical characteristics of side-by-side vortex streets associated with two identical Gaussian-shaped islands separated by a gap perpendicular to the ambient atmospheric flow. Through 12 different simulations we tested the four different gap-ratios, two different island diameters, and two different sets of boundary conditions. Largely similar behavior was observed in these atmospheric simulations to what has been documented
Figure 7.12: Gap jet meandering with time in the C02 GR1.0 D50km case. Panels represent instantaneous wind speed vectors and magnitudes at the lowest vertical grid level in the C02 GR1.0 D50km case at an interval of 4.5 hours going left to right and top to bottom.
in lower Reynolds number 2-dimensional cases of flow past two side-by-side cylinders.

First, it was found that the interactive behavior of the two side-by-side vortex streets demonstrates similar phasing regimes with respect to gap-ratio as has been recorded in low-Re flows. These regimes were revealed using 36 hr Hovmoller diagrams and cross-correlation analysis and were identified as: single bluff-body wake, flip-flopping wake, anti-phase synchronization, and in-phase synchronization in order with respect to increasing GR. To reiterate, these regimes have been found in both observational studies using particle image velocimetry [8] and direct numerical simulation [107].

In addition, we explored the spectral properties of the multi-vortex street wake region by looking at various time series within the atmospheric boundary layer. This enabled detection of single and multiple spectral peaks with frequencies that fluctuated with respect to changing GR. This indicated the existence of, in some cases, two vortex street frequencies one associated with the length scale of one island and the other associated with the length scale of both islands effectively acting as one. This behavior has been cataloged by several other studies for low Reynolds number flows [e.g., 219, 240].

Another characteristic of the vortex street interaction which we investigated here was the behavior of the gap-jet. It was documented that the orientation, and to some degree the magnitude, of the high wind speed gap-jet fluctuated with time at the low gap-ratios. On the other hand, with increasing gap-ratios, the temporal variability of the gap-jet orientation and its magnitude decreased. The meandering nature of the lower GR gap-jet is a feature associated with the flip-flopping vortex street interaction phase and essentially dictates which side of the wake region will have the instantaneous dominant vortex shedding. The dominance of one obstacle’s vortex street over the other has been clearly shown using time averaged laboratory observations [8] and the hand-off in time of this dominance has been theorized to be connected to an instability of the in-phase wake regime [41].

In conclusion, this chapter has presented numerical simulations and analysis which support the concept that the dynamical characteristics of atmospheric side-by-side interactive vortex streets are similar in nature to low Reynolds number, 2-D vortex streets. Moving forward, this work can be extended and reinforced by performing additional simulation of atmospheric vortex street interaction under different environmental conditions (i.e., geographical location, island height, island configurations).
Figure 7.13: Surface transect location (red line) used to construct Hovmoller diagrams in Figure 7.14.
Figure 7.14: Horizontal wind speed Hovmoller diagram taken from a surface transect parallel to the ambient flow in the center of the gap for the C02 GR1.0 D50km (upper left), C02 GR1.5 D50km (upper right), and C02 GR2.0 D50km (lower center) experiments. The two black lines represent the northern and southern extent of the islands to the east and west of the Hovmoller transect.
Chapter 8

Mapping Optical Ray Trajectories through Island Wake Vortices

8.1 Introduction

Optical refraction is a phenomenon by which the propagation trajectory of a wave bends towards a more dense medium (interchangeably, also referred to as optical ray bending in this work). As historically documented by Lehn et al. [129], the subject of optical refraction has been known to mankind for millennia and is largely responsible for several optical phenomena in the atmosphere. Examples include, the premature return of the Sun in the polar spring (i.e., the Novaya Zemllya effect) [130], rapidly evolving figures on the horizon (i.e., Fata Morgana) [188, 127], and the classic desert inferior mirage effects [98]. Although, these examples represent exotic manifestations of atmospheric refraction effects, some degree of optical refraction is expected whenever there are spatial gradients of temperature, humidity, or pressure along the propagation path. To an observer, the optical refraction caused by these gradients essentially results in displacement of the perceived location of objects [128, 71]. That being said, refraction effects become especially prominent when considering long-range (i.e., 10s to 100s of km) propagation paths [127]. At these distances, super-refraction conditions can enable optical waves to propagate beyond the horizon [52], while sub-refraction conditions can effectively reduce maximum visibility ranges for specific target and sensor height combinations. In addition, various mesoscale meteorological phenomena can also lead to complex spatio-temporal refractive patterns thereby influencing optical ray trajectories in the atmosphere. These variabilities significantly contribute to the variations in angle of arrivals of optical rays as path

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1 The material presented in this chapter appears in the following publication:
length increases.

Owing to the dynamical nature of the atmosphere, the meteorological conditions which dictate optical refraction, are neither stationary in time nor homogeneous in space. This is evidenced by experimental studies which have observed, for example, vertical laser beam drift of 4 m at a distance of 10 km over a 30 minute time period [35] and beam movement at a rate of 7.5 \( \mu \text{rad/sec} \) at a distance of 15 km [171]. Moreover, dynamic bending of laser beam trajectories, as observed experimentally, can be associated with large-scale self-organized atmospheric phenomena [234]. One such example is mesoscale land-sea breeze circulation which has been found to influence optical refraction and, in some cases, contribute to acoustic or electro-magnetic (radio frequency and optical frequency) wave ducting [15]. The ducting effect is associated with a specific refractive condition under which a layer of strongly stratified air traps electromagnetic or acoustic waves [187]. In ducting conditions, the trapping layer is commonly characterized by the thermal inversion at the top of the atmospheric boundary layer [118]. Capturing the position and evolution of the inversion layer, and the processes affecting it, is important for describing atmospheric refraction [77]. Even when atmospheric wave trapping is not present, the atmospheric boundary layer inversion can still be a dominant feature impacting refraction, especially in marine environments [39]. In this study, we analyze the optical refractive effects which are invoked in conjunction with mesoscale island wake vortex shedding. These vortices are capable of forcing significant perturbation, or even destruction, of marine boundary layer inversions downstream of islands [81, 170]. That being said, it should be noted that in this work we study km-scale refractive effects, which is not to be confused with microscale refractive effects associated with processes such as turbulence and gravity wave-breaking.

Island wake vortices, such as those embedded within von Kármán vortex streets (e.g., Figure 8.1), have commonly been observed near mountainous island archipelagos for several decades [45, 223, 27, 244]. These mesoscale atmospheric von Kármán vortex streets (VKVSs) differ from the classical VKVSs long observed in the laboratory since their formation does not primarily rely upon frictional boundary layer separation [208]. Instead, the vertical vorticity cores embedded within atmospheric VKVS patterns are formed through the tilting of baroclinically generated horizontal vorticity into the vertical axis immediately downstream of the island. This theory has been supported by recent studies demonstrating the association of von Kármán vortices with warm potential temperature anomalies, shed rhythmically downstream from their host islands [81, 170]. In addition, several other studies dedicated to quantifying the contribution of various physical mechanisms on the vertical vorticity generation and maintenance of VKVSs, have generally agreed that tilting of the thermal inversion is a dominant process [198, 197, 192].

The baroclinic tilting, and accompanying warm potential temperature anomalies, associated with mesoscale VKVSs are interesting from an optical wave propagation view point for several
reasons. For one, the baroclinic tilting can force strong temperature gradients, typically found in the vertical axis, to take on non-conventional horizontal components. In conjunction, the refractive anomalies collocated with the warm-core vortices may exhibit periodicity similar to that of the vortex shedding frequency, on the order of hours [170]. This phenomenon may result in the temporal variability of optical wave trajectories impinging on the VKVSs and thus impact the performance of imaging and laser communication systems operating between mountainous islands (e.g., Aleutian Islands, Canary Islands, Hawaiian Islands) [230, 76].

The current study provides analysis of the complex impact of island wake vortices on the trajectories of optical rays. This analysis is performed using a conventional ray tracing technique based on a geometric optics approximation. Furthermore, a non-hydrostatic mesoscale model is used to simulate the complex optical refractivity profiles associated with island wake vortices and used as input for the ray tracing studies. Through mesoscale modeling, realistic time-dependent refractivity fields which are heterogeneous in all three spatial dimensions are generated [14, 77, 231]. This approach differs from more basic analyses of optical refraction in the atmosphere which typically assume a stationary and horizontally homogeneous refractive index profile that is generalized to be valid for the entire region of interest [e.g., 121, 70].

### 8.2 Methodology and Simulation Details

To study the impact of island wake vortices on optical ray trajectories, three case studies were considered: Madeira island, July 5th, 2002; Guadalupe island, June 11th, 2000; and Maui island (Hawaii), May 27th, 2003 (Figures 8.1 and 8.2). The first two island wake vortices were identified based on low-altitude MODIS [165] and MISR [123] satellite imagery and were selected as case studies because they represent distinct wake vortex shedding at two different geographical locations. Madeira, in particular, is a popular island for VKVS research due to its rugged terrain perpendicular to the regional Northeastern trade winds; thus, it has been the subject of several other island wake studies in past literature [45, 27, 67, 49, 40]. Madeira has a prominent central chain of mountains that runs approximately West/East with a maximum elevation of 1,862 m above mean sea level (msl). At its widest point, Madeira island is about 45 km in diameter at sea level and covers an area of about 823 km$^2$. Guadalupe island, on the other hand, has a maximum elevation of about 1300 m msl and is significantly smaller at about 243 km$^2$. Since the wake vortices near Madeira island are large, well-organized, and extensively studied in the past, they will be our primary interest in this chapter.

The third (Hawaii) case study (Figure 8.2) was included because it lacked the well-defined vortices typically found within VKVSs, but consisted of wake vortices nonetheless. Within the sparse cloud cover in Figure 8.2, coherent wavy patterns and swirls are noticeable downstream (i.e., to the West) of the islands. Additionally, non-uniform solar reflection off of the sea
surface (i.e., sun glint) indicates varying surface wind speeds [108] which are present in the wake region. Clearly, the wake vortices associated with the Hawaiian islands in Figure 8.2 do not lend themselves to the classical von Kármán-type vortex street as did the Madeira and Guadalupe island-induced vortices. Instead, these wake vortices lack distinct closed circulations and are interactive with neighboring vortices produced by adjacent islands. For this case, we focus our analysis on the wake vortices associated with the Hawaiian island of Maui which has an area of 1883 km$^2$ and a maximum elevation of 3055 m msl.

Despite the added complexity of the Hawaii case, this location has significant practical importance as the well-known Mauna Kea and Haleakalā astronomical observatories are located in this region. While the high altitude locations of these observatories are ideal for reducing turbulence-induced scintillation [9] and scattering due to aerosols [54] in the planetary boundary layer, refractive anomalies introduced by vortices produced by the mountainous Hawaiian islands themselves may cause optical ray bending and enhanced intensity scintillations as observed by Vorontsov et al. [230] in a recent field experiment focusing on quasi-horizontal optical wave propagation.

### 8.2.1 Mesoscale Numerical Modeling

In this study, we have used the well-documented Weather Research and Forecasting (WRF) model to simulate the atmospheric flow and associated refractivity fields present in each of
the three cases. All the simulations were run for a total of 48 hours with the first 24 hours being used as spin-up time for the model. The final 24 hours of the simulations encompassed the days shown in Figures 8.1 and 8.2 and were used for analysis, thereby enabling us to study the structure and temporal evolution of the vortex patterns (e.g., periodicity). The simulations used a nested modeling domain with a highest resolution of 1 km horizontally to model the island wake vortices; the innermost domain was approximately 800 km by 800 km in size. In all of the domains, there were 51 pressure-based, terrain following, vertical coordinates between the surface and 100 mb with approximately 18 levels below 1 km above ground level. For initialization and boundary conditions, the ERA-Interim reanalysis dataset was used, which is derived through an extensive process of data assimilation and analysis and produced by the European Centre for Medium Range Weather Forecasts [29]. ERA-Interim contains environmental data records from 1979 through the present at 6-hour intervals and includes a horizontal resolution of approximately 79 km with 60 vertical coordinate levels.

The physics parameterizations used in these simulations included WRF Single-Moment 5-class Microphysics [86], RRTM Longwave Radiation [157], Dudhia Shortwave Radiation [59], Kain-Fritsch convection in the outer domains [106], Noah Land Surface Model [43] and the
Mellor-Yamada-Janjić planetary boundary layer scheme [99]. It should be noted that these model configuration options are not necessarily standard in mesoscale modeling and that other configuration options are also available. Therefore, it is important to consider model uncertainty with respect to configuration (e.g., spatial resolution, physics parameterization) when interpreting simulation results. To explore this issue, please see Appendix 8.8 for a comparison of model variability with respect to one configuration option (i.e., vertical resolution). Further research into this topic will be addressed in a future work aimed at quantifying the uncertainty of coupled mesoscale and ray tracing simulations. Nevertheless, this modeling system is very similar to that used in the successful Madeira VKVS simulations of Couvelard et al. [49] and Nunalee and Basu [170], albeit with higher horizontal resolution.

8.2.2 Optical Ray Tracing

The numerical results yielded from the WRF simulations were used to compute refractive index \( n \) as a function of wavelength \( \lambda \) at all locations within the innermost computational domain using Eq. 8.1.

\[
n_\lambda - 1 = \frac{1}{T}[A_d(\lambda)P_d + A_w(\lambda)P_w].
\] (8.1)

Here, \( T \) is temperature (K); \( P_d \) and \( P_w \) represent the partial pressures (hPa) of dry air and water, respectively. The corresponding reduced refractivities are \( A_d \) and \( A_w \). We define \( A_d \) and \( A_w \) using the equations suggested by Ciddor [46]:

\[
A_d(\lambda) = 10^{-8} \left( \frac{d_1}{d_0 - \frac{1}{\lambda^2}} + \frac{d_3}{d_2 - \frac{1}{\lambda^2}} \right) \frac{T_0^d}{P_0^d},
\] (8.2)

\[
A_w(\lambda) = 1.022 \times 10^{-8} \left( w_0 + \frac{w_1}{\lambda^2} + \frac{w_2}{\lambda^4} + \frac{w_3}{\lambda^6} \right) \frac{T_0^w}{P_0^w},
\] (8.3)

where \( T_0^d = 288.15 \) K, \( P_0^d = 1013.25 \) hPa, \( T_0^w = 293.15 \) K, and \( P_0^w = 13.33 \) hPa. Also, the coefficients \( d_0, d_1, d_2, d_3, w_0, w_1, w_2, \) and \( w_3 \) are as follows:

\[
d_0 = 238.0185 \text{ \( \mu m \)}^{-2}; d_1 = 5792105 \text{ \( \mu m \)}^{-2}; d_2 = 57.362 \text{ \( \mu m \)}^{-2}; d_3 = 167917 \text{ \( \mu m \)}^{-2};
\]

\[
w_0 = 295.235; w_1 = 2.6422 \text{ \( \mu m \)}^2; w_2 = -0.032380 \text{ \( \mu m \)}^4; w_3 = 0.004028 \text{ \( \mu m \)}^6.
\]

Based on the time-dependent 3D refractivity fields generated from the WRF model simulation, a series of ray tracing simulations were performed. The focus of the ray tracing simulations was to model the trajectories of optical rays as they propagated through the island wake vortices. That being said, the origin of the rays was defined by a height above mean sea level \( (h_o) \) with a horizontal propagation axis approximately aligned with the wake axis. In other words, this axis represents the azimuth of the vertical plane into which the simulated rays propagate.
In addition, the rays began propagation at an angle \( \theta_i \) with respect to the surface tangential plane. A schematic of the ray tracing physical configuration is provided in Figure 8.3. In Figure 8.3, \( R_j \) represents the surface relative range between ray origin and the ray position being analyzed. \( h_{i,j,k} \) denotes the height above mean sea level of the refracted ray trajectory (i.e., a hypothetical ray propagating through the atmosphere) with respect to \( \theta_i, R_j \), and time index \( k \). In addition, \( H_{i,j,k} \) represents the height above mean sea level of the identical ray except traveling through a vacuum. In this chapter, we will discuss the temporal variability of \( h_{i,j,k} \) as simulated by a ray tracing code initialized by refractivity fields generated by the WRF model. The temporal variability of \( h_{i,j,k} \) will be quantified by the height perturbation with respect to time \( (h'_{i,j,k}) \), where \( h'_{i,j,k} \equiv h_{i,j,k} - \langle h_{i,j,k} \rangle \) with angular brackets denoting temporal averaging over the analysis times described in Section 8.2.1 (i.e., the final 24 hours of the simulations).

Figure 8.3: Schematic of ray tracing simulation (not to scale). \( h_o \) is the height above mean sea level of the ray origin and \( R_j \) is the range of the ray propagation path. \( \theta_i \) represents the angle between the surface tangential plane and the ray departure angle at the ray origin. The refracted ray trajectory is shown in solid grey with \( h_{i,j,k} \) denoting the height of the refracted ray trajectory with respect to \( \theta_i \), at range \( R_j \), and at time \( k \). For comparison, the trajectory of the same ray propagating through a vacuum is also shown in addition to its height \( (H_{i,j,k}) \) at a range \( (R_j) \) and time \( k \). Analyses in Sections 8.3 and 8.4 will discuss ray height perturbation \( (h'_{i,j,k}) \) where \( h'_{i,j,k} \equiv h_{i,j,k} - \langle h_{i,j,k} \rangle \).

In all, a total of six numerical experiments were conducted which are summarized in Table 8.1. Four of the numerical experiments had ray propagation axes oriented downstream of the islands through the wake vortices. For comparison, two numerical experiments were also performed where the rays propagated upstream from the islands in order to benchmark the
results found in the downstream cases. For each case, rays originated from evenly distributed angles between minimum and maximum $\theta$ angles in radians, which in most cases were -0.010 and +0.010 radians. In order to investigate optical effects pertinent to different perspectives on the island for the Madeira island case, the height of the ray origin was varied from a location on the downstream slope of the island (500 m msl) to near the crest of the modeled island (1250 m msl)\(^2\).

Table 8.1: Details of the numerical ray tracing simulations.

<table>
<thead>
<tr>
<th>Case</th>
<th>Propagation Axis</th>
<th>min($\theta$)</th>
<th>max($\theta$)</th>
<th>$h_o$</th>
<th>$R_j$</th>
<th>Number of Rays</th>
</tr>
</thead>
<tbody>
<tr>
<td>Madeira</td>
<td>Downstream</td>
<td>-0.010, 0.010</td>
<td>500 m</td>
<td>50 km</td>
<td>50</td>
<td>50</td>
</tr>
<tr>
<td>Madeira</td>
<td>Downstream</td>
<td>-0.020, 0.000</td>
<td>1250 m</td>
<td>150 km</td>
<td>200</td>
<td>200</td>
</tr>
<tr>
<td>Madeira</td>
<td>Upstream</td>
<td>-0.010, 0.010</td>
<td>500 m</td>
<td>50 km</td>
<td>50</td>
<td>50</td>
</tr>
<tr>
<td>Guadalupe</td>
<td>Downstream</td>
<td>-0.010, 0.010</td>
<td>500 m</td>
<td>50 km</td>
<td>50</td>
<td>50</td>
</tr>
<tr>
<td>Guadalupe</td>
<td>Upstream</td>
<td>-0.010, 0.010</td>
<td>500 m</td>
<td>50 km</td>
<td>50</td>
<td>50</td>
</tr>
<tr>
<td>Hawaii</td>
<td>Downstream</td>
<td>-0.016, 0.000</td>
<td>1000 m</td>
<td>200 km</td>
<td>200</td>
<td>200</td>
</tr>
</tbody>
</table>

The ray trajectories were computed by solving the Eikonal equation, or ray tracing equation (Eq. 8.4).

$$\frac{d}{ds} \left( n \frac{dr}{ds} \right) = \nabla n.$$ (8.4)

By decomposing Eq. 8.4 into a system of two first-order differential equations (Eqs. 8.5 and 8.6), the ray tracing equation was numerically integrated (see Wheelon [237] for reference).

$$\frac{dr}{ds} = \frac{t}{n},$$ (8.5)

$$\frac{dt}{ds} = \nabla n,$$ (8.6)

In Eqs. 8.5 and 8.6, $r$ and $t$ represent the position and direction vectors of a ray, while $s$ is the scalar distance along a ray path. Individual rays were then traced using a method similar to that proposed by Southwell [212]. That is, the position and direction vectors of individual rays

\(^2\)Note that several factors caused the peak height of Madeira in the model (≈1250 m) to be lower than the actual peak height of Madeira (≈1850 m). These included uncertainty in the terrain height boundary conditions provided through digital elevation models, and terrain smoothing during interpolation onto WRF modeling grid.
were iteratively updated using the following equations:

\[ r_{i+1} = r_i + \delta t/n, \]  
\[ t_{i+1} = t_i + \delta \nabla n. \]  

(8.7)  
(8.8)

Here, \( r_i = x_i \hat{i} + y_i \hat{j} \) and \( t_i = n_i \cos \phi_i \hat{i} + n_i \sin \phi_i \hat{j} \) where \( \phi_i \) is the ray-trajectory associated angle and \( \delta \) is the integration step size. The ray tracing calculations were performed using an Euler scheme with a step size of 50 m [186]. Additionally, the ray tracing calculations were performed on a 2-D plane bounded by the locations of the ray origin, the Earth’s center, and the ray propagation azimuth.

### 8.3 Vortex Street Refractivity

Before investigating the behavior of ray trajectories oriented through the wake vortices, we first explore the flow perturbations induced by the island of Madeira in the primary case study. As already discussed in Section 8.1, vertical baroclinicity is an intrinsic feature of atmospheric vortex streets. In this section, we show that the atmospheric stability regime in and around the wake vortices is also an important factor that governs refractivity within the wake region.

Figure 8.4 illustrates an example of the instantaneous simulated VKVS wind pattern associated with the island of Madeira. It is evident from this figure that a VKVS is induced by the island penetrating through the stably stratified free atmosphere above the planetary boundary layer [67]. This wake signature displayed zones of high wind shear along the flanks of the island and along the edges of each vortex downstream. In addition, it is clear from Figure 8.4 that the wake vortices are collocated with warm potential temperature anomalies along with a broad area of warming downstream of the island. The thermal anomalies originate from the stably stratified free atmosphere above the planetary boundary layer. As confirmation of the fact that the crest of Madeira Island was reaching the free atmosphere, the upper panels in Figure 8.5 show the temporal evolution of potential temperature upstream (17.05 W / 32.95 N) and downstream (17.29 W / 32.52 N) of the island. In the upstream regime, it can be seen that there is a well-mixed boundary layer within about the first 800 m msl while a moderately strong thermal capping inversion is present in the layer just above. This stability profile was sufficient to invoke a complex wake region containing a hybrid system dominated by both vertical and horizontal advection. The top right panel in Figure 8.5 shows the effects of the vertical motion within the wake region. The regular oscillations of boundary layer height are manifestations of wave-like motions (e.g., gravity waves) associated with the stratification aloft [197].

Looking next at a vertical cross-section of potential temperature through Madeira island area
Figure 8.4: Plan view of instantaneous horizontal winds and potential temperature at approximately 500 m msl for the Madeira event at 6 UTC on July 5th, 2002. Contours represent potential temperature at 500 m msl from 289-295 K (blue to red) at an interval of .5 K and vectors represent normalized horizontal wind speed and direction (blue is weaker and red is stronger wind).

and its wake (Figure 8.5 bottom panel), evidence of the vertical perturbation to the inversion layer can be clearly seen up to a distance of roughly 200 km downstream of the island. In some locations, the vertical motion within the wake disturbed the ambient boundary layer inversion to the point where it became oriented nearly vertically. In addition, the thermal gradient within the inversion underwent a series of compressions and decompressions in association with the vertical motions. Lastly, upstream of the island, isotherms rise up vertically in response to the upstream high pressure anomaly before dipping down slightly and intersecting with the topography. This effect is also related to the thermal stability and is indicative of a layer of horizontal flow around the island.

The vertical motions induced by the stratified flow over and around the island initially
invoked thermal perturbations which fueled refractivity anomalies; the anomalies were then primarily advected downstream by horizontal motions. Figure 8.6 illustrates the position and magnitude of refractivity anomalies embedded within the wake region at a specific time. Obviously, there is a layer of high refractivity gradient near the height of the capping inversion, which will induce amplified bending of optical ray trajectories. In addition, areas of significantly high (low) refractivity gradient can be seen in association with the locations of compression (decompression) of the thermal inversion. Very close to the island itself (less than 10 km), intense turbulent diffusion has destroyed much of the thermal inversion and, as a result, the refractivity gradient is less distinct there.

The wake of Madeira induced interesting refractivity features not only in the vertical axis,
but also in the horizontal axis. In Figure 8.7, similarities can be seen between the refractivity anomalies and the cloud, wind, and thermal patterns already presented in Figures 8.1 and 8.4. The region of warm potential temperature close to the island is a product of the downslope flow on the lee of Madeira and also the breaking gravity wave. Farther downstream, the warm region can then be seen shedding into the outer flow in the expected VKVS pattern.

The downward protruding potential temperature bubbles seen in Figure 8.7 are compressing the boundary layer beneath them; as these areas return to equilibrium, they stretch the column and in turn, act to increase vertical vorticity. Through this mechanism, the vortices within the vortex street are able to maintain their individual rotation for longer times and distances downstream. The implications of these refractivity anomalies on optical ray trajectories will be discussed in the following section.

### 8.4 Optical Wave Propagation through Vortex Streets

In this section, we study the propagation of a series of simulated optical rays through the coherent wake vortices associated with the Madeira and Guadalupe islands-induced VKVS events. Using the ray tracing technique described in Section 8.2.2, ray trajectories were simulated through the wake region of Madeira island. Rays originated from a location on the southwestern slope of Madeira island (approximately 17.11 W / 32.75 N) and with a propagation azimuth angle southwest through the wake vortices.

Figure 8.8 shows ray trajectories calculated for four different times overlaid onto cross sections of refractive index vertical gradient (RIVG). The times shown are at an interval of
2.5 hours, and during this 7.5 hour period several repeating patterns are visible. Note also that similar repetitious patterns were continuously found throughout the entire 24 h simulation although they are not shown here. The wave-like motion of RIVG visible in the wake region is highly correlated with the vertical temperature gradient previously illustrated. Therefore, the layers of high RIVG are, for the most part, coincident with the strongly stratified ABL capping inversion. The simulated optical rays propagating through the VKVS in the vicinity of Madeira island were modified by two primary manifestations of the RIVG perturbations. The first, and most obvious, effect which had implications for the simulated ray trajectories was the vertical displacement of the ABL inversion. This displacement effectively modified the RIVG along the ray paths which, in turn, altered ray trajectories. In addition, the tightening and loosening of the thermal temperature gradient within the vortex street further modified the RIVG along the ray paths, as will be shown later in this section.

Next, it is intuitive that we compare the RIVG profiles of upstream and downstream transects in order to understand the differences in refraction between the wake region and the background flow. In the upstream regime there appears to be at least slight evidence of terrain induced flow modification; that is, between 10 and 30 km upstream small amplitude waves are distinguishable (Figure 8.9). These wave signatures share similar characteristics with trapped internal gravity waves expected in supercritical flow [38]. Despite the presence of these more subtle features, the upstream refractive index regime is much more quiescent than the downstream regime. The same was also observed for the Guadalupe island case.
Figure 8.8: Vertical gradient of refractive index (contours; m$^{-1}$) for four different times at an interval of 2.5 hours going left to right and top to bottom. Panels represent vertical cross sections through Madeira’s VKVS and black lines demarcate 10 ray trajectories each separated by an angle of .002 radians.

To further emphasize the temporal variability of the refracted ray heights, attention is directed to Figure 8.10. Here, the 50 km ray height perturbation (i.e., the instantaneous ray height difference from the 24 hour mean height) is shown for 20 different rays associated with both the Madeira and Guadalupe events. For rays propagating upstream, no regular vertical height perturbation from the mean develops. Downstream, however, vertical ray trajectory perturbations, up to 25 m, can be observed with periods on the order of hours for the Madeira case. The fluctuations illustrated in Figure 8.10 have time scales very similar to the known shedding frequencies of mesoscale atmospheric VKVSs [170].

The periodicity of the ray heights was investigated by taking the autocorrelation function of
the height perturbations shown in Figure 8.10. In doing so, the true periodic nature of the ray trajectory fluctuations were revealed as shown in Figure 8.11. For both events, especially for the Madeira event, the period is clearly visible as denoted by the high correlation peaks around 5 hours. The periods were recorded to be slightly longer than 5 hours for the Madeira event and about 4.5 hours for the Guadalupe event. Nevertheless, it should be noted that the periodicity produced by the simulated Guadalupe event was significantly weaker than that produced by the Madeira event. This is likely due to several factors, but the smaller size of Guadalupe island compared to Madeira island may be an important difference between the two events. As previously mentioned, the downstream periods correspond to the known periods of mesoscale atmospheric VKVSs which are a function of wind speed and island geometry [45, 133, 170]. This connection is important as it demonstrates the fact that optical ray bending in the atmosphere can be highly correlated with distinct structures in the atmosphere which evolve on time scales of minutes to hours.

In addition, we investigated rays with origins near the crest of Madeira island propagating downstream between an angle of 0.00 and -0.02 radians with the surface parallel plane (see Figure 8.3). In this simulation, we increased the ray density to 200 rays in order to better observe ray behavior further downstream. As can be seen in Figure 8.12 (left), rays originating from the crest of the island are still susceptible to modulation associated with the VKVS when considering lower angles with respect to the horizon. This demonstrates the relevance of island wake vortices for optical communication between island crests and lower-level (e.g., near sea-surface) receivers. The right panel of Figure 8.12 shows the fluctuation of ray height with time at a distance of 150 km downstream from the ray origin. The large degree of variability, up
to 50 m within a 12 hour time span, is essentially attributed to amplifications of the behavior shown in Figure 8.10. Also, the color scheme in Figure 8.12 (right) emphasizes the fact that the wake refractive structures are most prominent within a certain layer beneath the maximum height of the island. In addition, several fluctuation frequencies appear in Figure 8.12 (right) and the daily variability is substantially greater than that shown in Figure 8.10 (top left panel).

8.5 Non-Vortex Street Optical Wave Propagation

The results mentioned thus far highlight the implications of self-organized coherent von Kármán-type vortices on optical ray trajectories. However, there exists a spectrum of island wake vortex
Figure 8.11: Autocorrelation of ray heights at 50 km from the observer location. Left and right panels represent Madeira and Guadalupe islands, respectively. Each trace corresponds to a different ray with a different emission angle (blue is lower and red is higher).

Figure 8.12: (Left panel) Vertical gradient of refractive index (contours; m$^{-1}$) and ray trajectories (black lines) with an origin near the crest of Madeira. (Right panel) Time series of ray height perturbation 150 km from the ray origin. Each trace represents a different ray with a different emission angle (blue is lower and red is higher).

patterns [141] which may result in a similar spectrum of implications for refraction [37]. While it is beyond the scope of this work to investigate each of these patterns individually, it is necessary to mention how (a)typical the results presented in Section 8.4 are in the context of atmospheric wake vortices in general.

Using the WRF model, we simulated the event shown in Figure 8.2 and, from the output, we performed a single long-range (i.e., 200 km) ray tracing simulation. In this numerical
experiment, 200 rays were released between an angle of 0.000 and -0.016 radians from a height of 1000 m msl on the southwestern slope of Haleakalā on the island of Maui (20.70 N / 156.26 W). The rays propagated southwest into the wake region associated with Maui and the Hawaii islands towards a location at approximately 19.39 N and 157.75 W. The variability of the height perturbation above the surface is shown in Figure 8.13 along with an instantaneous snapshot of the ray trajectories.

Figure 8.13: (Left panel) Vertical gradient of refractive index (contours; m⁻¹) and ray trajectories (black lines) with an origin on the southwestern slope of Haleakalā. (Right panel) Time series of height perturbation 200 km from the origin location. Each trace represents a different ray with a different emission angle (blue is lower and red is higher).

Figure 8.13 clearly shows that the height of the rays fluctuates over this 18 hour period, as was observed in the Madeira island and Guadalupe island cases, yet it was much less periodic. That is, the regular period of the vortex shedding was not prevalent. Also, the amplitude of the fluctuations are surprisingly much lower (i.e., 3 m or less), despite the fact that the propagation distance is longer. Among other factors, this may perhaps be due to counteracting positive and negative boundary layer inversion height perturbations produced by Maui and the Hawaii islands along the ray trajectory. Nevertheless, the lower ray angles still demonstrate the greatest temporal height variability as was observed in the previous two case studies. Also, these lower angles again correspond to optical waves propagating through the thermal inversion at the top of the planetary boundary layer. Together with the results presented in Section 8.4, this indicates that optical ray trajectories downstream of bluff islands are very sensitive to the dynamics of the capping inversion. In any case, a physical understanding of complex wake vortices (e.g., those associated with the Hawaiian islands) as they pertain to optical ray bending
is a worthy subject for future analysis.

8.6 Discussion

In this study, we have used mesoscale numerical simulations to investigate the impacts of both coherent (i.e., von Kármán vortices) and incoherent island wake vortices on optical ray trajectories in the atmosphere. Using the technique of optical ray tracing, we have identified connections between ray trajectory perturbations and VKVS behavior. Thereby, we have shown mesoscale island wake vortex shedding to be responsible for vertical height perturbations of over 20 meters, during a 24 hour period, at a range of 50 km. At the same time, it was also observed that the distinct influence of island wake vortices varies in periodicity with respect to events. For the Madeira island and Guadalupe island cases, the autocorrelation of ray height perturbations indicated dominant frequencies of four to five hours. These simulated ray propagation anomalies were connected to intricate vertical temperature gradient features collocated with wake vortices. At levels between 300 and 500 m msl, simulated vortices were found to have warm core centers approximately 3-4 degrees K warmer than their surrounding environments at the same level. These temperature anomalies consequently produced RIVG anomalies with magnitudes up to 2-3 times greater than the surrounding environment. These anomalies were essentially due to two individual processes which were (1) the vertical displacement of the thermal capping inversion and (2) the compression and decompression of the thermal inversion with the oscillation of the downstream gravity wave. Additionally, it was further observed that incoherent vortices associated with the Hawaiian islands produced a much more convoluted ray trajectory compared to the coherent vortex streets of the Madeira island and Guadalupe island cases. This behavior manifested in the trajectory height perturbations to be much smaller and with indistinct periodicity. This was true despite an increased ray propagation distance of 200 km.

8.7 Concluding Remarks and Future Perspectives

The influence of meteorological features on the spatio-temporal behavior of long-range optical ray trajectories is critical to increasing the utilization of optical communication technology across long-range atmospheric paths. As described here, atmospheric phenomena (e.g., wake vortices) can induce temporal variability of optical ray trajectories which is anticipated to have implications for optical systems which depend on precise laser guidance. This connection exemplifies the dependency of atmospheric refraction on mesoscale phenomena and that it can, at least qualitatively, be numerically modeled. We have shown here that the coupled mesoscale model - ray tracing framework detailed in this work is capable of capturing a vast spectrum
of atmospheric flows which have direct influence on refraction and optical ray trajectories. Nevertheless, specific model configuration (e.g., spatial resolution, physical parameterization schemes) can potentially be sources of model uncertainty when it comes to accurately simulating ray trajectories (See Appendix 8.8).

With the opportunities available through mesoscale numerical modeling, future studies will be capable of investigating additional atmospheric phenomena (e.g., land-sea breeze circulation, low-level jets) which may cause severe optical anomalies beyond the classical wake patterns primarily explored here. The Hawaii case presented here symbolizes the complexity of certain real-world scenarios and the optical ray trajectories which they may induce. An understanding of the long-range optical effects associated with island wake patterns, both quasi-idealized and complex, may offer insight into some of the unexpected optical effects observed experimentally [230, 76]. For example, Gurvich et al. [76] found evidence of isotropic turbulence in a layer 2.2 - 2.4 km above the surface when considering a 144 km laser beam, while Vorontsov et al. [230] found intensity scintillation associated with a 149 km laser beam to not adhere to traditional turbulence models. The nature of these contradictions indicates that perhaps conventional understanding of long-range optical wave propagation in the atmosphere does not account for the full spectrum of important physics. That being said, future research should have the goal of eventually delivering accurate 3D refractivity fields down to the $O(m)$ scales in near real-time and for long optical path lengths on the order of 100s of kilometers. While such a task is currently impractical, petascale computing may eventually facilitate such endeavors enabling not only comprehensive refraction modeling, but also diffraction modeling across all atmospheric scales.

8.8 Appendix: Sensitivity to Vertical Resolution

It is important to acknowledge the potential sensitivity of the mesoscale modeling results presented in this study to specific model configuration. In this context, specific model configuration includes components such as spatial resolution, physical parameterization schemes, and grid discretization method. Given this, we studied one possible element of sensitivity in order to emphasize the uncertainty of the aforementioned results. Here we performed an additional mesoscale model simulation of the Madeira case, with subsequent ray-tracing through the wake region, but with double the original vertical resolution of 51 vertical grid levels. The results of the high vertical resolution (i.e., 101 vertical grid levels) run are shown in Figure 8.14. As can be seen from the vertical cross-section plot, the qualitative features of the RIVG anomalies embedded within the von Kármán vortices are very similar to those produced by the coarser resolution simulation (i.e., Figure 8.6). However, the layers of high vertical refractivity gradient appear noticeably sharper and more intense as would be expected with higher vertical
resolution. In conjunction, ray tracing results indicate that the general oscillatory nature of ray height perturbations 50 km downstream from the origin is also present in the high-resolution simulation (Figure 8.14, lower left panel). Furthermore, the relative period of the height perturbation oscillation is essentially the same between the high-resolution simulation (Figure 8.14, lower right panel) and its original counterpart (Figure 8.11, left panel). However, there are quantitative differences between the coarse-resolution and high-resolution ray tracing results. That is, primarily the high vertical resolution simulation had a greater range of height perturbation fluctuation during the 24 simulation hours of interest. Undoubtedly, this is due to the stronger vertical gradients of refractivity captured by the higher vertical resolution. With this in mind, we speculate that specific selection of mesoscale model configuration can potentially be a source of uncertainty when performing ray tracing based on mesoscale model output.

Figure 8.14: A vertical cross-section of RIVG ($m^{-1}$) through Madeira island and its wake region from the high vertical resolution simulation (upper panel) and a time series of ray height perturbations 50 km downstream of Madeira origin location (lower left panel) and corresponding autocorrelations (lower right panel) using high vertical resolution. Each trace represents a different ray with a different emission angle (blue is lower and red is higher).
Chapter 9

Conclusion and Future Directions

The purpose of this research has been focused on characterizing the dynamical properties of orographically-induced mesoscale range von Kármán vortex streets in the atmosphere. Atmospheric vortex streets are inherently a high-Reynolds number fluid phenomenon and, up until this point, there has not been a thorough evaluation of their dynamical properties with respect to the well-documented dynamical properties of low-Reynolds classical vortex streets. In addition to differences in viscosity (i.e., Reynolds number) and fluid type, atmospheric VKVSs have an extra (vertical) dimensionality along with a different formation mechanism associated with baroclinicity as opposed to the separation of a frictional boundary layer. These differences collectively challenge the applicability of, supposedly universal, dynamical properties associated with classical VKVSs to geophysical circumstances.

In studying the dynamical behavior of mesoscale range vortex streets we have made extensive use of a high-fidelity mesoscale numerical modeling platform, namely the WRF model. Before performing numerical experiments with the WRF model, we evaluated its ability to simulate realistic vortex streets as observed by remotely-sensed satellite imagery. This included documenting model sensitivity to static topographic relief datasets in Chapter 4 and quantitative verification against cloud-level stereoscopic winds in Chapter 5. In Chapter 4 we demonstrated the importance of using accurately representative static topographic relief data when performing numerical simulations of island wakes. In particular, the peak height of the island appeared critically important to capturing the correct wake regime (e.g., von Kármán vortex shedding, weak wind wake). Furthermore, in Chapter 5 we qualitatively and quantitatively compared WRF model simulations of two realistic VKVS events to MISR stereoscopic cloud height wind vectors. It was found that the model accurately captured many of the same features recorded by MISR such as the vortex street spacing, wind magnitudes embedded within the vortex streets, and the position of directional shear zones. We also explored the sensitivity of these simulated results to PBL parameterization and concluded that minor sensitivity can be expected. For
further information on the topic of PBL scheme sensitivity in mesoscale models, the reader was directed to Appendix A.

Based on findings made in earlier chapters, we moved on to performing scientific experiments in Chapter 6 where we cataloged the periodicity of mesoscale range vortex streets with respect to a length scale (i.e., Reynolds number). It was found that the simulated atmospheric vortex streets adhered to the well-defined Reynolds number-Strouhal number similarity theory associated with low Reynolds number 2D vortex streets when eddy viscosity is used as a proxy for kinematic viscosity. This result was supported by the fact that that the island diameter, specifically at the height of the PBL capping thermal inversion, acted as the effective length scale of the system. This relationship was discovered after varying the vertical peak height of the island produced little to no effect on shedding frequency while modifying the horizontal diameter of the island produced a direct modification of the vortex street frequency.

In Chapter 7 we explored the dynamical characteristics of two interactive side-by-side atmospheric vortex streets. This case has special applicability to society as many multi-island archipelagos are known to support interactive vortex streets (e.g., Aleutian Islands, Canary Islands, and Hawaiian Islands). In this chapter it was found that the simulated side-by-side vortex streets demonstrated distinct phasing regimes: single bluff body wake, flip-flopping wake, anti-phase synchronization, and in-phase synchronization. These phases also produced their own unique spectral properties sometimes leading to two individual spectral peaks; one associated with the frequency of flow around a single island and the other associated with flow around both islands. Furthermore, it was found that at low gap-ratios the gap jet orientation meandered with time on a period that was not consistent with either vortex street period. As GR increased the gap-jet meandering became less pronounced and its magnitude also increased. Understanding of this behavior could have applicability for wind forecasting in regions such as the Alenuihaha Channel where volatile wind patterns often interfere with recreational and commercial maritime activities.

In Chapter 8 we examined the effects of atmospheric von Kármán vortex shedding on the propagation trajectories of long-range optical rays. By coupling mesoscale model output with a geometric optics-based ray tracing code we found that the regular vortex shedding induced temporal perturbations of ray heights on the order of tens of meters at distances around 50 km downstream from the islands. These height perturbations were a direct result of the ray propagation through the thermal anomalies collocated with von Kármán vortices. Alternatively, in cases where non-coherent vortex shedding was simulated the height perturbations were substantially less intense. These findings highlight the potential impacts that island wakes can have on optical communication and demonstrate the value of mesoscale models when considering long-range optical paths influenced by horizontally heterogeneous refractivity fields.
Moving forward, there are several ways in which this research can be expanded upon. First, a more comprehensive model validation should be carried out which uses high-fidelity in-situ observations. In this work, we have used radiosonde data and remotely-sensed data for validation; however, both of these measurement types have inherent limitations. Radiosonde data is very difficult to geographically reference in a model because the radiosonde itself drifts horizontally with altitude. In an area where a vortex street exists, the horizontal variability of meteorological variables can be large thereby increasing the uncertainty of data gathered from a radiosonde released at one position. Some potentially viable alternatives to using radiosonde data for model validation in the context of simulating atmospheric vortex streets include the use of mesonets and aircraft measurement campaigns.

Building on the use of advanced meteorological observation platforms, a more thorough model validation should include an evaluation of the precise structure of simulated VKVSs versus observed VKVSs. This effort will include evaluating the vortex street aspect ratio which is intrinsically connected to the vortex shedding frequency. Along these lines, one challenge to overcome will be the accurate identification of the center of rotation of the atmospheric vortices. This could include the use of traditional metrics such as directional and magnitudinal wind shear or more advanced metrics which possibly incorporate the thermal signatures collocated within individual vortices. Furthermore, spatial and temporal characteristics of VKVSs can be explored through the analysis of local energy dispersion assuming the flow fields are well-resolved.

In addition, it is important to note that this research has only investigated mesoscale range vortex streets in the atmosphere and has not begun to explore the properties of microscale range vortex streets. However, before a comprehensive theory of atmospheric VKVSs can be developed it is important to understand how they behave on spatial scales where turbulent eddies are similar in size to the von Kármán vortices. To do so, the use of advanced large-eddy simulation (LES) models will be essential to properly account for the effects of turbulent motions on the dynamical effects of atmospheric VKVSs. The reader is pointed to Appendix B for more information about the capabilities of WRF-LES as it pertains to simulating stratified flow around an isolated hill. In connection with Appendix B, the existence of not only offshore atmospheric vortex streets, but also onshore VKVSs deserves exploration. Based on the literature referenced in Chapter 2, there is no obvious reason why vortex streets cannot indeed occur in conjunction with onshore hills and mountains. That being said, it would be valuable to understand whether or not the physics surrounding offshore VKVSs is also relevant to onshore VKVSs. Once a complete picture of geographically independent atmospheric vortex streets is developed, it will be possible to move forward and understand the role they play on global energy budgets.
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Appendix A

Numerical Modeling of Mesoscale Phenomena and Sensitivity to Planetary Boundary Layer Parameterization

A.1 Introduction

Offshore territories offer a number of advantages for wind power production such as reduced surface friction, space for larger turbine rotors, and close proximity to large population centers [125, 124, 200]. In 2011, Europe installed over 800 MW of offshore wind energy bringing its total capacity to 3,812.6 MW, nearly a 400% increase from 2007 [239]. In Asia, China claims the largest national offshore wind energy infrastructure with 150 MW installed capacity as of the beginning of 2011 [105]. In the U.S., an offshore wind energy infrastructure has struggled to materialize due to political and social barriers. However, these challenges are expected to be overcome in the near future given the fact that abundant wind resources can be found relatively close to high energy consumption areas such as New York, Boston, and Philadelphia [220, 61]. As proof of this imminent development, over 2,000 MW of capacity is currently in the planning and permitting process along the Mid-Atlantic and Northeast coastlines [161]. Similarly, the U.S. West Coast also boasts high wind power potential and suitable bathymetry for the production of up to 75.5 GW of wind power off of the coast of California alone [60]. Despite its advantages, appeal, and recent development, international offshore wind energy

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1The material presented in this chapter appears in the following publication:
development is challenged by unique meteorological factors which are more or less negligible in
the case of onshore development of the same size.

Like their onshore counterparts, offshore wind farms are designed to extract the maximum
amount of wind power possible based on local wind classifications. However, coastal and off-
shore wind regimes present a distinct suite of complex atmospheric processes which heavily
dictate local wind climatologies and, in turn, wind power production [23]. Additionally, the
accuracy of conventional wind resource assessments in these locations may be significantly con-
strained by limited observational data [146]. In coastal areas, the primary factors which induce
these offshore atmospheric processes are attributed to the intrinsic thermal and roughness differ-
ces of water versus land [22]. The differences in heat capacity between water and land
drives differential surface heating on a diurnal time-scale at land/sea interfaces. This differential
heating often times forces circulations which manifest into propagating mesoscale frontal
boundaries (e.g. land-sea breeze front). In some instances these land-sea breeze fronts may
be characterized by wind speed perturbations greater than 5 m/s [36]. Additionally, internal
boundary layers (IBLs) can develop as synoptically driven surface winds adjust to differences
in roughness as they traverse coastal boundaries; the effects of these flow adjustments can be
evident up to 70 km from coastlines [24]. IBLs may also develop near sea surface temperature
(SST) gradients associated with distinct oceanic currents; buoyant heterogeneities near these
SST boundaries can lead to enhanced turbulence [143]. Potential offshore wind farm sites are
also exposed to coastal low-level jets (LLJs) which are relatively narrow elevated channels of
wind maxima. Although the dimensions of offshore LLJs have not been fully documented,
onshore LLJs are typically bounded between ~100 m and ~1000 m above ground level (AGL)
and span a width of ~100 km [19, 176]. Similar to onshore LLJs [82, 83], offshore LLJs are
also expected to transport moisture, heat and other scalars (i.e., passive, active, and reactive)
hundreds of kilometers. In addition to these mesoscale atmospheric phenomena, other types of
phenomena on different scales, such as hurricanes and waterspouts, may also influence offshore
wind energy development.

Given the current lack of offshore wind measurements at hub height, the necessity for
computationally modeled wind climatology data is significant [161, 220]. Despite this necessity,
atmospheric boundary layer (ABL) simulations intended for a variety of applications, including
but not limited to wind energy, have been shown to be heavily sensitive to planetary boundary
layer (PBL) parameterization. This sensitivity is also known to be amplified when complex
mesoscale atmospheric processes are involved [122, 150, 213, 199, 201]. Recent coastal and
offshore mesoscale modeling efforts devoted to wind resource assessment typically employ a
standard PBL scheme for all simulations regardless of the atmospheric features commonly
present in their computational domain. In this appendix, we provide evidence of the diversity
in coastal mesoscale model solutions attributed to PBL parameterization. We also advocate for the use of ensemble physics (i.e. the consensus of multiple physics schemes [104]) when simulating diverse coastal mesoscale phenomena for offshore wind resource estimation.

In summary, mesoscale atmospheric phenomena are particularly relevant to offshore wind energy both from productivity and sustainability perspectives. The motivation for this appendix is rooted in the fact that the limitations of modern numerical mesoscale models are not well understood with respect to these offshore phenomena. In order to partially address this knowledge-gap, we will focus on a single phenomenon, the coastal LLJ\(^2\). The coastal LLJ is chosen above other processes because, like onshore LLJs, it can be especially influential in affecting wind power production and is often overlooked.

The content of this appendix is as follows: in Section A.2 we discuss the state-of-the-science concerning onshore and offshore LLJs. Section A.3 outlines a specific offshore LLJ case study used for model validation and summarizes the details of the mesoscale modeling system. Modeled results are presented and compared to observations in Section A.4 while model solution sensitivities are identified with respect to specific modeling components. Section A.5 summarizes the key results from this study. Finally, in Section A.6 we discuss the implications of our findings and directions for future work.

### A.2 Background

The characteristics which define an LLJ lack scientific precision and many different definitions have been suggested over the past 60 years. While establishing an LLJ climatology for the Eastern U.S. and Great Plains, Bonner et al. [33] defined an LLJ as a wind maximum below 1.5 km AGL and having a peak wind speed of greater than 12 m/s with a 50% decrease in wind speed above the maximum (but below 3 km AGL). Many years later, Whiteman et al. [238] modified the definition of Bonner et al. [33] to allow weaker jets (i.e. with wind maximum less than 12 m/s) to also be classified as LLJs. Additionally, Song et al. [211] further modified the definition of Whiteman et al. [238] and defined the maximum height of LLJs to be below 2 km AGL (as opposed to 3 km AGL). Vastly different from each of these classifications, Andreas et al. [11] and Banta et al. [19] put forth a more relaxed definition of LLJs and characterized them as having wind speed decreases of 2 and 1.5 m/s (respectively) both above and below a wind maximum of any intensity. As exemplified by the diversity of each of these definitions, to classify a particular phenomenon as a LLJ can sometimes be a subjective process. It is for this reason that in this appendix we have selected to study an event which appears to abide by all popular LLJ definitions.

\(^2\)Here we use the terminology coastal LLJs and offshore LLJs interchangeably. In both cases we are referring to LLJs associated with land-sea interfaces which may be positioned completely, or partly, over land or sea.
Before investigating the ability of a conventional mesoscale model to simulate coastal LLJs, it is important to understand the fundamentals of LLJs and the conditions which spawn them. For this reason, it is necessary to review the recent studies and findings related to offshore LLJs. However, considering the lack of literature pertaining explicitly to offshore LLJs, we will also review material concerning onshore LLJs as they are expected to share similar characteristics.

A.2.1 Onshore Low-Level Jets

To date, the majority of scientific studies related to LLJs have focused primarily on onshore nocturnal jets. Through these investigations much has been learned about the inherent features of LLJs and their implications for wind energy [213, 20, 18]. Aside from the obvious advantage of boosting nighttime wind power production, LLJs are capable of producing enhanced low-level (∼100 - 200 m AGL) turbulent kinetic energy (TKE) through shear instabilities [203, 248]. Observational data from field studies such as the CASES99 experiment in the U.S. Great Plains [184] support the hypothesis that strong turbulence could be generated aloft near the jet core and then mixed downward [19]. This enhancement of turbulence beneath LLJs, which is introduced into the nocturnal stable boundary layer, can threaten the sustainability of modern wind turbine blades, which are not designed with bursting turbulence in mind [109]. Also, enhanced ABL turbulence has been documented to degrade wind turbine efficiency [236].

Onshore LLJs are hypothesized to be invoked by a number of different atmospheric mechanisms such as inertial oscillation [32, 174, 226], differential surface heating [115], and diurnal thermal wind balance associated with sloping terrain [174]. Bonner [33] first collected observational data from the U.S. Great Plains LLJs and found them to occur over 60% of the nights. More recently, other LLJ climatology studies have been conducted using records collected by meteorological masts, sodars and wind profilers [156, 238, 211, 16]; some of these studies found warm season LLJ events to occur more frequently than cool season events in the U.S.

A.2.2 Offshore Low-Level Jets

The characteristics of offshore LLJs are much more elusive than onshore LLJs due to a relative lack of observations. The maritime environments which support offshore LLJs have a distinctly weaker diurnal stability cycle compared to onshore environments [137]. They have also been observed to yield vertical wind speed gradients greater than predicted by the standard Monin-Obukhov similarity theory [124]. This is especially important for wind energy applications because the LLJs that develop offshore may be closer to the surface than onshore LLJs. For example, in the Baltic Sea LLJs have been documented as low as 30 m above sea level in stably stratified boundary layers [206]. While the characteristics of offshore and onshore LLJs may be similar in some regards, the formation mechanisms of offshore LLJs are sometimes different
In the early 1990’s Doyle and Warner [56] studied an offshore LLJ event along the Carolina coastline which appeared to have been initiated by geostrophic forcing in a narrow baroclinic zone. This study found the jet height and intensity to be sensitive to both surface flux modifications and latitude. On the other hand, other studies of offshore LLJs [247] in the same region found that the sloping terrain related to the Appalachian mountain chain, which essentially runs parallel to the coastline, may also aid in U.S. east coast LLJ formation. Yet even more recently, the common summertime U.S. east coast LLJ was hypothesized to be most typically initiated by the differential surface heating associated with the highly fluctuating ground temperatures onshore and the adjacent steady ocean surface temperatures [48]. In another part of the world, Jiang et al. [100] studied a Pacific Ocean LLJ off the coast of Chile and concluded that geostrophic forcing, coupled with baroclinic amplification induced by the Andes mountains and differential surface heating, were critical components to the jet’s existence and behavior. Near the northern German coastline four nocturnal LLJs were observed during the PUKK experiment; from these observations Kraus et al. [116] documented one of these jets to have been associated with the familiar onshore formation mechanism known as inertial oscillation [32].

Another, less popular yet equally plausible, coastal LLJ formation mechanism is based on the theory of Hsu [90]. This theory states that coastal LLJs can develop between mesoscale elevated inversions and nocturnal microscale ground-based inversions. This scenario may be established through nighttime radiational cooling and can induce an LLJ as a result of both Venturi and gravity-wind effects [90]. A criteria of this formation mechanism is that the over-sea air temperature should ideally differ from the adjacent over-land temperature by more than 5°C since this allows the land-surface cold pool to form [89, 92]. From a wind energy perspective, this theory is important because it supports the possibility of very low LLJs (<100 m AGL) within ~20 km of global coastlines (this is the region along which the near-surface thermocline is most significantly sloped and gravity-effects are maximized). All in all, literature from past years has indicated that many coastal jets may form due to a combination of dynamics which in some cases may act harmoniously with one another.

A.2.3 Offshore Low-Level Jet Climatology

Many studies have been devoted to studying the climatology of inland LLJs [19, 33, 238, 211, 16]. Findings from these studies have exposed the fact that LLJ activity is not a rarity and can be quite common in certain favored locations (i.e. the U.S. Great Plains are known to experience LLJs up to 60% of the nights). On the other hand, studies documenting the frequency of coastal LLJs are much fewer. Nonetheless, in their global nocturnal low level jet index Rife et al. [190] highlighted the presence of increased LLJ frequency in certain coastal regions around the world;
the Eastern U.S. being one of them. Additionally, Zhang et al. [247] studied specifically the
U.S. Mid-Atlantic LLJ jet and found it to have a magnitude of 8 - 23 m/s with a mean height
of 670 m AGL and a frequency of 15 - 25 days per month during the warm season. Similarly,
Ryan [195] found jets over Maryland, of durations equal to or greater than 2 hours, to occur
on 43% of days with a mean intensity of 14 m/s. Also, Colle and Novak [48] found LLJs near
the New York Bight to have typical magnitudes between 11 - 12 m/s at a height of only 29 m
above sea level as measured by buoy wind speed sensors.

The information provided by the aforementioned studies, coupled with the known LLJ
formation mechanisms previously discussed, lead the authors to believe that coastal LLJs along
the U.S. East Coast are common enough, and strong enough, to be influential to offshore wind
ergy development and performance. In order to support this assumption, a preliminary
climatology study was conducted mainly to assess the annual number of occurrences of coastal
LLJs in the Northeastern U.S. coastal region. This climatology used hourly windspeed data
from the RUTNJ vertical wind profiler (see Section A.3) to document the frequency, magnitude
and height of LLJs that occurred in that region throughout 2010. It was found that coastal
LLJs occur over the RUTNJ profiler at least 118 days out of the year (i.e. 32%) with a mean
magnitude of \(~15\) m/s and mean height of \(~525\) m AGL. A vertical wind profile was considered
to have an LLJ signature when a clearly discernible wind maxima was present for at least 2
samplings (i.e. 2 hours) with at least a 4 m/s decrease in wind speed both above and below the
jet core. The magnitude of an LLJ was taken to be its highest wind speed maxima recorded
by the profiler and its height was the height at which its peak wind speed occurred.

We feel that the statistical estimates yielded from this simple study are extremely conserva-
tive and open to a number of sources for error: (1) our LLJ criteria was fairly strict (compared
to [11, 19]), (2) the dataset was not free of data-gaps (i.e. many LLJ events may not have been
captured by the profiler), (3) the profiler was unable to capture very low LLJs as profilers do not
record measurements within about the lowest 100 m AGL and (4) the coarse dataset sampling
interval (1 hour) makes classifying average LLJ magnitude and height potentially biased con-
sidering their temporal morphology. Additionally, it should be noted that the year selected for
analysis may or may not have been meteorologically representative of a typical year. Nonethe-
less, this brief coastal LLJ climatology investigation indicates that the LLJ studied in this work
is characteristically not an outlier for the region and that offshore wind energy developers can
expect similar LLJ activity to occur about 1 out of every 3 days or more.

A.3 Data and Methodology

In this study, a prominent coastal LLJ event (July 15-17, 2011) was identified based on observa-
tional data provided by two 915 MHz MADIS vertical wind profilers along the US Northeastern
Seaboard (RUTNJ in New Brunswick, NJ and BLTMD in Beltsville, MD) and a radiosonde from Upton, NY (Figure A.1 left panel). Profilers, such as RUTNJ and BLTMD, have been previously used to identify and analyze similar Mid-Atlantic coastal LLJs with reasonable accuracy [195, 178]. The MADIS profilers used in this study provide wind speed and direction data for heights between \( \sim 100 \) m and \( \sim 2000 \) m AGL and have a temporal resolution of 10 minutes. The radiosonde provided wind and temperature profiles at 00Z and 12Z each day at about seven vertical heights below 1 km AGL. The quality control procedures for these datasets are provided in the National Weather Service Technique Specification Package [166]. Observational data from the profilers and radiosonde was used to verify version 3.3 of the Weather Research and Forecasting model (WRF) while simulating the identified LLJ event.

The WRF model is a non-hydrostatic mesoscale numerical weather prediction (NWP) model used extensively within academia, government, and industry. The WRF model is capable of simulating a plethora of synoptic scale, mesoscale and microscale atmospheric phenomena and is equipped with multiple parameterization schemes (e.g. micro-physics, land-surface interaction, radiation, planetary boundary layer, etc.) [205]. In order to evaluate the performance of the WRF model's PBL parameterization options in simulating this offshore LLJ event, WRF was run multiple times over the U.S. East Coast domain for the dates of the LLJ event occurrence. For each model run a different PBL parameterization was invoked while all other model
configurations were held constant. The various PBL schemes evaluated here were: Yonsei University [85], Mellor-Yamada-Janjic [99], Quasi-Normal Scale Elimination [217], Mellor-Yamada Nakanishi and Niino Level 2.5 [162], Asymmetric Convective Model [180, 181], and BouLac [34]. Other parameterization schemes used in these simulations were: WRF Single-Moment 5-class Microphysics [86], RRTM Longwave Radiation [157], Dudhia Shortwave Radiation [59], Kain-Fritsch Cumulus [106], and the Noah Land Surface Model [43].

A.3.1 Modeling Domain Configuration

For all LLJ simulations, the WRF model’s preprocessing system (WPS) was used to construct a nested numerical modeling domain centered on the location of the RUTNJ vertical wind profiler. This domain configuration consisted of a large parent domain with 2 nested domains. The outer domain, d01, employed a 18 x 18 km horizontal resolution; the intermediate domain, d02, used a horizontal resolution of 6 x 6 km and the target domain, d03, encompassed the observation sites and operated using a horizontal grid resolution of 2 x 2 km (Table A.1). Throughout all the domains, the vertical resolution remained constant with 50 grid points between the surface and 16,000 m AGL, the lowest of which was ~8 m AGL (Figure A.1 right panel). About 18 out of the 50 total vertical grid points were below 1 km. Land use and topographical properties for these simulations were taken from the U.S. Geological Survey at resolutions comparable to the grid resolution. Model runs were executed for a total of 96 hours with the first 24 hours being used as spin-up time for the simulations. The simulations operated using a 90 second integration time step on d01, a 30 second time step on d02, and a 10 second time step on d03. Model output was extracted for analysis every 10 minutes for d03. Standard wind resource models use slightly lower vertical resolutions and about the same horizontal resolutions [183]. Operational NWP models, on the other hand, have a much coarser horizontal resolution but about the same vertical resolution [1]. Appendix A.7 contains a detailed study of the effect of vertical resolution on accurate offshore LLJ simulation.

<table>
<thead>
<tr>
<th>Domain</th>
<th>∆x</th>
<th>∆t</th>
<th>Nx</th>
<th>Ny</th>
<th>Domain Size</th>
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<td>90 sec.</td>
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<td>100</td>
<td>1782 km × 1782 km</td>
</tr>
<tr>
<td>d02</td>
<td>6 km</td>
<td>30 sec.</td>
<td>181</td>
<td>181</td>
<td>1080 km × 1080 km</td>
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<tr>
<td>d03</td>
<td>2 km</td>
<td>10 sec.</td>
<td>181</td>
<td>181</td>
<td>360 km × 360 km</td>
</tr>
</tbody>
</table>
A.3.2 Model Initialization

Like all computational fluid dynamics models, the WRF model requires initial and boundary condition data to initiate and time integrate atmospheric simulations. For the case study presented here, the WRF model was prescribed meteorological initialization conditions from the ERA-Interim and NARR datasets. ERA-Interim is a data assimilation system produced by the European Centre for Medium Range Weather Forecasts and contains environmental data records from 1979 through the present at 6-hour intervals. The ERA-Interim dataset includes a horizontal resolution of approximately 79 km and 60 vertical coordinate levels [29]. NARR is produced by the National Centers for Environmental Prediction and is similar to ERA-Interim; however, its geographic coverage only encompasses North America. The NARR dataset has a horizontal resolution of approximately 32 km with 42 vertical levels [2]. As will be discussed, model initialization data appears to have a substantial influence on the WRF model’s wind output (see Section A.4.2). Ultimately, ERA-Interim data were chosen for most of the analysis in this appendix as it allowed the WRF model to capture the LLJ event much more accurately.

A.4 Results

During July 14 - 17, 2011 the U.S. Northeastern Seaboard experienced a relatively uneventful synoptic-scale atmospheric pattern. Surface high pressure, centered off of the coast of Virginia, persisted throughout the majority of the period, preventing the encroachment of any large-scale weather system (Figure A.2 upper left panel). Based on modeled data, the individual nocturnal LLJs which developed during this time frame were oriented parallel to the U.S. northeastern coastline with a crosswise width of about 100 km and a streamwise length of about 1000 km (Figure A.2 upper right panel).

The initiation of the three consecutive LLJs could easily have been due to a combination of multiple atmospheric processes. Of note, the geostrophic flow was from land to sea and fairly weak (Figure A.2 bottom row) throughout the event. This was ideal forcing for the jet formation mechanism outlined by Hsu [90] as the offshore pressure gradient could have potentially amplified PBL winds in a baroclinic zone near the coast as they were decoupled from the surface at night. Furthermore, the event took place in July when the sea surface temperatures (SST) were generally warmer than the adjacent nighttime over-land temperatures. This temperature difference likely played a significant role in jet formation due to the across-shore changes it induced to the near-surface microscale cold-pool. In addition, other mechanisms common for onshore LLJs, such as sloping terrain, may have contributed to the formation of these coastal LLJs. On the other hand, inertial oscillation which is a very common onshore LLJ formation mechanism was not observed in the measured directional wind speed hodographs.
Figure A.2: The Eastern U.S. seaboard was dominated by a synoptic-scale surface high pressure system on 7/16/2011 as shown by the Hydrometeorological Prediction Center surface analysis (upper left). A plan view of 300m AGL wind speed displays the coastal LLJ (outlined in black) along the northeastern U.S. seaboard as simulated by the WRF model valid 0Z (upper right). Wind data is masked over areas where the terrain height is greater than 300 m AGL and verification data source locations are labeled the same as in Figure A.1. 1000 mb geostrophic wind vectors also show the weak offshore geostrophic forcing along the U.S. Northeastern coastline at 00Z 7/16/2011 (bottom left) and 12Z 7/16/2011 (bottom right). Geostrophic wind figures were provided by Plymouth State Weather Center; value of maximum vector size (m/s) is indicated beneath each figure.
Wind speed observations from the RUTNJ profiler (Figure A.3 left panel) display the structure and temporal evolution of three consecutive nocturnal LLJs. The three jets were observed on the nights of July 15th, 16th, and 17th between roughly 00UTC and 12UTC. The timing of each jet’s development and disappearance, along with their vertical limits (\( \sim 100 - 800 \) m AGL), is fairly consistent throughout all nights. Nevertheless, the intensity and core structure of each jet varies throughout the event. The LLJ which occurred on July 17th was the most intense with peak wind speeds greater than 15 m/s over RUTNJ. Although the BLTMD profiler site was not directly beneath the jet cores, it did record weak LLJ signatures on each of the three nights (Figure A.3 right panel).

![Figure A.3: Time vs. height plots of wind speed observations from the RUTNJ (left) and BLTMD (right) MADIS vertical wind profilers during July 14th - 17th, 2011. Note that the color scale of the BLTMD plot has been adjusted to reflect weaker wind maxima.](image)

A.4.1 Sensitivity to Planetary Boundary Layer Parameterization

Individual PBL schemes within the WRF model simulated the series of Northeastern U.S. coastal LLJs with noticeable diversity. Figure A.4 illustrates the fact that each PBL scheme struggled to capture the intensity and structure of each of the LLJs; however, the timings and vertical bounds of each jet feature were modeled reasonably well. The MYJ, ACM2 and QNSE PBL schemes simulated the most intense, and most accurate, coastal LLJ event with maximum winds approaching 15 m/s for the final jet\(^3\). The initial nocturnal jet was the weakest and the most difficult for each scheme to represent.

\(^3\)During the course of the current study, a bug in the YSU scheme (WRF version 3.3) was discovered (pers. comm. Heather Richardson, 2012). This bug has been corrected in WRF version 3.4.1 and should result in improved nighttime low-level winds.
Figure A.4: WRF model derived time vs. height plots of wind speed during July 14th - 17th, 2011 at RUTNJ using ACM2 (top left), BLc (top right), MYJ (middle left), MYNN2 (middle right), QNSE (bottom left) and YSU (bottom right).
Wind Speed and Potential Temperature Profiles

Validation using radiosonde profiles also show interesting qualities of the WRF-PBL schemes. When compared to observations at 00Z on July 16th; it is evident that all PBL schemes over-predicted the LLJ’s intensity and under-predicted its height with the MYJ and QNSE schemes performing the poorest (Figure A.5 left column). Also, at this time the near surface modeled atmosphere has a cold potential temperature bias. Nevertheless, looking at the jet which occurred on July 17th at a later time, the MYJ and QNSE schemes most closely match the shape of the wind speed profile (Figure A.5 right column). At this same instance, other PBL schemes overestimated the height of the jet core (∼300 m AGL) and underestimated its maximum intensity (∼13 m/s). The BouLac scheme was the most extreme in its intensity underestimation with an error of about 4 m/s. On the other hand, wind speeds above the LLJ were unanimously overestimated by all schemes. With regards to potential temperature correlation, simulation results were consistent throughout the boundary layer and free atmosphere (Figure A.5); however, there was a warm thermal bias near the surface below ∼300 m AGL.

Wind Speed Distribution

The diversity among PBL-schemes translates to non-negligible differences in the wind resource characterization at particular locations. To quantify these differences, the modeled 3-day wind speed distributions for 179 m AGL at the RUTNJ site are compared to the observed wind speed distribution in Figure A.6. It is evident that all schemes underestimated the Weibull scale parameter ($C$) and also demonstrated significant diversity in terms of the shape parameter ($K$) [148]. A quantitative assessment of the Weibull parameters, mean wind speed, power density and annual energy production (AEP) error is shown in Table A.2. AEP estimates are based on the power curve of the GE 1.5 S model wind turbine with a rotor plane diameter of 70.5 m. In order to avoid extrapolating wind down to the recommended hub height, a hub height of 179 m AGL was assumed as that was the lowest height of consistently good quality profiler data. Power density errors ranged from -8.82% to -75.96% while estimated AEP errors ranged from 0.28% to -71.86%. The multi-physics ensemble AEP estimate had an error of about -30% which ranks 5th among all other PBL schemes. In addition to potentially offering increased wind resource estimates, the ensemble-physics assessments also provide a measure of uncertainty for the expected wind regime. This is provided through the spread in solutions from the ensemble members and can be beneficial for financial risk assessments and planning.
Figure A.5: Wind speed (top row) and potential temperature (bottom row) as a function of height for the OKX radiosonde launch site. Modeled profiles are colored and sounding observations are denoted by a line with stars. Data valid July 16th, 2011 00Z (left column) and July 17th, 2011 12Z (right column). Dashed black lines represent typical wind turbine rotor plane top and bottom heights (40 m and 140 m AGL).

Vertical Wind Shear

The power law exponent ($\alpha$) is a common metric used to quantify magnitudinal wind shear between two vertical heights above ground level. The variable, $\alpha$, is defined in the power law equation (Eq. A.1) where $Z_1$ is the height of the lower level, $Z_2$ is the height of the upper level, and $U$ represents wind speed at each height.

$$\left( \frac{U_{Z_2}}{U_{Z_1}} \right) \sim \left( \frac{Z_2}{Z_1} \right)^\alpha \quad (A.1)$$
Figure A.6: Wind speed distribution at 179 m AGL at the RUTNJ site along with modeled wind speed distribution. Distributions are representative of the three days with LLJs present.

Figure A.7 (upper left panel) displays the temporal evolution of $\alpha$ values computed using $Z_1 = 40$ m and $Z_2 = 140$ m AGL for the LLJ event. Increases in these values up to 1 indicate a highly sheared boundary layer flow. The heights of $Z_1$ and $Z_2$ roughly correspond to typical rotor plane bottom and tip heights of modern offshore wind turbines. As expected, all WRF-PBL schemes modeled an increase in vertical wind shear in association with the formation of each LLJ with $\alpha$ upwards of 0.8 in some cases. Despite these general increases, the degree of $\alpha$ increase varied with each scheme, sometimes by a factor of over two. These differences can likely be attributed to the specific treatment of diffusion which is unique to each scheme. To put these values in perspective, wind energy developers typically assume values of $\alpha$ between 0.11 and 0.20 which are typically standard for onshore locations [63] and the International Electrotechnical Commission has set wind turbine standards to withstand $\alpha = 0.2$ [97]. Obviously, high shear environments such as those induced by the coastal LLJs discussed here will generate unanticipated, enhanced stress on tall turbines [72]. The presence of the three consecutive coastal LLJs also enhanced the simulated absolute directional wind shear ($\beta$) in the WRF model. The variable, $\beta$, is defined as the absolute value of the difference in wind direction between two heights above ground level (Eq. A.2),

$$\beta = |\phi_{Z_2} - \phi_{Z_1}|$$  \hspace{1cm} (A.2)

where $\phi$ represents wind direction at a specific height ($Z$). The inconsistency between each PBL scheme’s physics is highlighted by this parameter as some schemes exhibit minimal directional wind shear whereas others produce directional wind shear over 45 degrees (Figure A.7). Nevertheless, a PBL-scheme ensemble analysis of this coastal LLJ event clearly illustrates an increase in absolute directional wind shear of about 20 degrees with each jet. Over land, van Ulden and Holtslag [228] estimated the typical value of $\beta$ taken over the entire nighttime stable
Table A.2: Wind Resource Statistics at 179 m AGL. AEP estimates are based on the power curve of the GE 1.5 S model wind turbine with a rotor plane diameter of 70.5 m. A hub height of 179 m was assumed as that was the lowest height of consistently good quality profiler data.

<table>
<thead>
<tr>
<th></th>
<th>Mean Wind Speed</th>
<th>Power Density</th>
<th>AEP</th>
</tr>
</thead>
<tbody>
<tr>
<td>ACM2</td>
<td>5.6427</td>
<td>107.81</td>
<td>5.46</td>
</tr>
<tr>
<td>Blc</td>
<td>4.1630</td>
<td>43.30</td>
<td>2.08</td>
</tr>
<tr>
<td>MYJ</td>
<td>6.3152</td>
<td>151.12</td>
<td>6.88</td>
</tr>
<tr>
<td>MYNN2</td>
<td>5.9222</td>
<td>124.63</td>
<td>5.78</td>
</tr>
<tr>
<td>QNSE</td>
<td>6.4926</td>
<td>164.21</td>
<td>7.41</td>
</tr>
<tr>
<td>YSU</td>
<td>4.8900</td>
<td>70.17</td>
<td>2.91</td>
</tr>
<tr>
<td>Ensemble</td>
<td>5.5709</td>
<td>110.20</td>
<td>5.09</td>
</tr>
<tr>
<td>RUTNJ-MADIS</td>
<td>6.6956</td>
<td>180.0989</td>
<td>7.39</td>
</tr>
</tbody>
</table>

Stability

Another interesting response of the WRF model to the coastal LLJs was evident in $R_{i_{rotor}}$, a derivative of the popular Bulk Richardson number [216]. The Bulk Richardson number accounts for atmospheric stability and also vertical wind shear in both horizontal axes throughout the entire boundary layer depth. Negative Bulk Richardson Number values indicate a positively buoyant/convective atmosphere whereas values greater than zero correspond to a stably stratified condition. The variable $R_{i_{rotor}}$, which is defined by Eq. A.3, behaves similarly to the Bulk Richardson number; however, it only considers shear and static stability parameters across the depth of typical wind turbine rotor planes (40-140 m AGL).

$$R_{i_{rotor}} = \frac{2(Z_2 - Z_1)g}{\theta_{Z_2} + \theta_{Z_1}} \left[ \frac{\theta_{Z_2} - \theta_{Z_1}}{(U_{Z_2} - U_{Z_1})^2 + (V_{Z_2} - V_{Z_1})^2} \right]$$  (A.3)

In the lower left panel of Figure A.7 the diurnal stability cycle of the RUTNJ location is evident by the very negative $R_{i_{rotor}}$ values during the afternoon hours and large positive values during the nighttime hours. Additionally, a prolonged period of slightly positive $R_{i_{rotor}}$ values corresponds to the presence of each nocturnal LLJ occurrence. This observation supports the
hypothesis that coastal LLJs promote boundary layer mixing and reduce nocturnal stability. These results were drawn from the general consensus of all PBL schemes, which were fairly consistent for the majority of each of the LLJ’s prevalence. However, variability among schemes was observed both during the afternoon hours and the times when each coastal LLJ was forming.

Turbulent Kinetic Energy

The WRF model-derived 90 m AGL turbulent kinetic energy (TKE) time series displays small increases in magnitude from 00Z-12Z each night; the times when the LLJ was present (Figure A.7 lower right panel). Although dwarfed by the level of turbulence present during daytime hours, the nocturnal turbulence associated with the presence of the LLJ approaches 0.5 m²/s² on the final night of simulation. The MYJ PBL scheme demonstrates artificial TKE clipping at 0.1 m²/s² while the BLc scheme maintains a near-zero TKE production environment during nocturnal hours. Also during the nighttime (i.e., the hours when the LLJs are present) simulated TKE is fairly intermittent and characterized by a turbulent bursting-type nature. Observations of TKE bursting similar to this has been observed over the U.S. Great Plains in association with onshore LLJs [19, 18, 109].
A.4.2 Sensitivity to Initial and Boundary Conditions

Although the primary purpose of this appendix is to assess the sensitivity of PBL parameterization on the ability to accurately model coastal LLJs, we similarly investigated the sensitivity of initial and boundary conditions on this accuracy. Surprisingly, we found the ability of the WRF model to capture this LLJ event to be very strongly related to initialization data. To illustrate this connection, the WRF model was run once again using identical configurations only this time using NARR data for initialization and boundary conditions. In Figure A.8 wind speed values from the RUTNJ profiler are subtracted from both the NARR-based simulation and the ERA-Interim based simulation previously mentioned (both simulations used the QNSE PBL scheme). This time-height comparison clearly show that the performance of the NARR-based simulation was much weaker compared to that of the ERA-Interim with regards to the first two LLJ intensities.

![Figure A.8](image)

Figure A.8: Time vs. height wind speed difference plot of WRF-model derived output minus observations from the RUTNJ MADIS vertical wind profiler using ERA-Interim (left) and NARR (right) initialization conditions. Data valid for July 14th - 17th, 2011.

The differences between the data in these two reanalyses are many and associated with multiple model variables. With that being said, it is beyond the scope of this appendix to identify the specific parameters responsible for the differences highlighted in Figure A.8. Nevertheless, scientifically speaking these parameters must be related to one or more of the formation and/or amplification mechanisms at play here. Assuming that the primary LLJ development mechanism at work here was that described in Section 2 (i.e. Hsu [90]), two key environmental conditions would be SSTs warmer than the adjacent land surface and weak offshore geostrophic wind.
From satellite observations, it can be seen that SSTs within the innermost modeling domain (d03) were much more heterogeneous than other regions in the vicinity (Figure A.9 bottom). On July 17th, multiple warm SST pockets were identifiable offshore of Northern New Jersey and Long Island. The ERA-Interim data better represented this area of warmer water compared to the NARR dataset, which was about 1-2 degrees cooler (Figure A.9 bottom). Warm SST anomalies, such as those visible in the satellite imagery, may have increased the nocturnal temperature gradient at the coast and thus contributed to LLJ formation. From a modeling perspective, this may have been one of many sources of error that potentially contributed to the poor performance of the NARR-based simulation. In order to test this speculation we attempted to improve the weaker NARR-based simulation by incorporating real-time, high-resolution satellite-based SST observations [3] as boundary conditions to the WRF run. Interestingly, this yielded little change in terms of the wind speed vertical profiles throughout
the event especially for the first two jets as they went largely unsimulated (Figure A.10). This indicates that efforts to improve modeled representation of coastal LLJs by simply addressing individual variables (i.e. SST) may offer little success. An alternative, possibly more helpful tactic from an offshore wind energy standpoint, would be to use a suite of various initialization and boundary conditions to produce modeled wind resource assessments.

Figure A.10: Time-height wind speed values for the RUTNJ site using QNSE PBL parameterization and NARR initialization data supplemented by high-resolution satellite-derived SSTs (left). Wind speed difference between original NARR and the NARR supplemented by satellite SSTs (right).

A.5 Discussion

In this appendix, the implications of coastal LLJs on mesoscale modeling for offshore wind energy applications have been outlined. We have emphasized the importance of accurate initial and boundary conditions for coastal LLJ simulation. In addition, multiple PBL parameterization schemes were evaluated based on their ability to simulate the atmospheric boundary layer with respect to coastal LLJs. For the particular Northeastern U.S. coastal LLJ event studied here, it was acknowledged that multiple process may have led to its development; however, weak offshore geostrophic forcing combined with significant land-sea temperature differences appeared to be important ingredients.

From a modeling perspective, our results showed that each PBL scheme tended to underestimate the intensity of each jet core while also struggling to capture the presence of the weakest LLJ altogether. On the other hand, the timing of each jet’s development and termination was accurately simulated by all PBL schemes. Furthermore, an extraordinary amount of magnitudinal and directional vertical wind shear ($\alpha \sim 1$ and $\beta \sim 20$) was present throughout depths common for wind turbine rotor planes ($\sim 40 - 140$ m AGL). As would be expected, the en-
hanced vertical wind shear promoted PBL mixing and also led to intermittent bursting of TKE at hub height. Essentially this degree of inflow heterogeneity has the potential to degrade the structural health of current wind turbines over time.

A.6 Future Directions

In a 2011 report [53], the U.S. Department of Energy (DOE) outlined a national offshore wind strategy for the development of an offshore wind energy infrastructure in the U.S. In this report, the DOE identified a number of major market barriers. Of these barriers, two important meteorological issues were highlighted: (1) “The offshore wind resource is not well characterized. This significantly increases uncertainty related to potential project power production and turbine and array design considerations, which in turn increases financing costs.” and (2) “Most meteorological, wave, and seabed data used in assessing potential offshore wind sites are based on extrapolations of data from on-shore sites, buoys, or limited surveys. The accuracy of these projections has not been properly validated.” Moving forward, the limitation of insufficient offshore meteorological datasets may be addressed in two ways. The first and the most obvious way to address this barrier is to install tall instrumented towers on the U.S. offshore continental shelf, in the Gulf of Mexico, and in the Great Lakes. The DOE has anticipated that these types of facilities will be “critical in assessing the costs, energy production, design requirements and overall economic viability of (offshore wind) projects.” Rationally speaking this is a long-term solution which has already been adopted in Europe (e.g. FINO towers in the North Sea) and Asia; however, it is a financially expensive proposition. An alternative and much more cost-effective way to address the issue of limited offshore data is to use mesoscale models to simulate the expected wind behavior where observations are sparse. This is an attractive solution given the constantly advancing state of high-performance computing and the increasing usability of models such as WRF. Despite the simplicity and efficiency of mesoscale models, the limitations imposed by every physical parameterization along with uncertainty in initial and lateral boundary conditions means that no model will be perfectly accurate as reinforced by this appendix. One might speculate that the long-term averaging (say averaging of 365 random day simulations taken from the past decade or so) might reduce or possibly even eliminate the dependence of simulated wind resource climatologies on the model physics parameterizations. Our present study (admittedly, limited in scope) does not support such speculation.⁴ In Figure A.11, we compare the 3-day averaged 90 m AGL wind speed from two different model physics options. At some offshore locations, the differences could be as high as 2 m/s. Based on this, we advocate

⁴ A recent mesoscale model intercomparison study [224], which compared six mesoscale models spanning over a period of one year, corroborates our finding (qualitatively). Over the Arctic Ocean, they found that the annual bias of 10 m wind speed among various models ranged from -0.98 m/s to 1.47 m/s.
the use of multi-physics ensembles which should provide more accurate wind resource averages and more importantly wind resource uncertainty outlooks.

Even with increased reliance and confidence in mesoscale models, the necessity for tall offshore meteorological towers cannot be ignored as they are desperately needed for model validation. Recently, NREL used ocean buoys, marine automated stations, Coast Guard stations/lighthouses and satellite derived 10 m wind speeds to validate offshore wind maps produced using a physics based mesoscale model [220]. “The wind measurements from these (sources) were extrapolated to 50 m above the surface and compared to the model estimates at the same height using shear exponents from the Power Law Equation (Elliott et. al. 1987).” As emphasized in this appendix, this validation technique is compromised by the fact that offshore wind regimes may often be characterized by shear exponents much larger than those derived through the Power Law Equation (Eq. A.1).

Regardless of the data constraints mentioned above, offshore wind energy remains an attractive avenue of development. However, if mesoscale modeling prevails as a offshore wind resource assessment tool, then it should incorporate new modeling techniques such as: four-dimensional data assimilation, coupling to microscale models (e.g. large-eddy models) and coupling to ocean-wave models. Each of these techniques offers unique benefits that may help to relax erroneous mesoscale model sensitivities, particularly those presented in this appendix.
Employing these types of new techniques may also prove valuable when attempting to simulate unique coastal atmospheric phenomena with high implications to wind energy. Also, further work should be done to climatologically characterize the prevalence and diversity of these phenomena. For example, occurrences of very low LLJs have been observed offshore (Figure A.12) yet their frequency and intensity remains largely unknown. To the best of our knowledge simulations of such phenomena have not been reported in the literature. We expect the ability of conventional mesoscale models to capture events such as this to be extremely weak considering the small-scale nature of the phenomena.

Figure A.12: Very low LLJ offshore of Martha’s Vineyard, MA on August 8, 2001 17UTC. Data measured during the Coupled Boundary Layers/Air-Sea Transfer experiment [62] and provided courtesy of Larry Mahrt.

A.7 Appendix: Vertical Resolution Sensitivity

While many components of mesoscale models can contribute the errors identified in this appendix, vertical resolution is particularly important for simulations focused on boundary layer flows. For this reason we present a concise sensitivity study dedicated to understanding the effects of vertical resolution on model accuracy while verifying against the coastal LLJ observational data. To test the effects of vertical resolution, the number of gridpoints in the ABL (i.e. about the lowest 3000 m AGL) were doubled and the coastal LLJ was re-simulated (all other modeling configurations identical to those described in Section A.3). In Figure A.13 the vertical wind profiles produced by simulations with increased vertical resolution are compared to the baseline simulations presented in Section A.4. Two different PBL schemes were chosen
for this resolution sensitivity study, one that performed fairly well (QNSE; Figure A.13 left) and another that performed poorly (MYNN2; Figure A.13 right). The quintessential difference between the baseline simulations and the higher resolution simulations is a slightly weaker and slightly lower jet core produced by the higher resolution runs. The increased resolution did not improve the comparably poorer MYNN2 simulation; in fact it deteriorated the PBL scheme's representation of the LLJ. The same can be said for the QNSE scheme, the result of increasing the vertical resolution only slightly weakened its performance which was nearly perfect in the baseline simulation.

Nonetheless, these observations are only valid for one time and at one location; nothing can be said regarding the capacity to which these observations can be generalized to describe effects at other locations or times. However, these results do agree with those recently found by Bernier and Bélair [28] who also found increased vertical resolution to not significantly improve or degrade the quality of wind forecasts for wind energy applications. On the other hand, Bernier and Bélair [28] did find that increased vertical resolution may in some cases greatly improve other, more sensitive, model prognostic variables (i.e. moisture fields). Theoretically, an improvement in model representation of these fields should indirectly improve wind representation; however, this improvement is likely insignificant considering the multiple sources of error inherent to mesoscale models. In light of the information presented in this sensitivity study, a justification cannot be made concerning the advantages of increased vertical resolution compared to lower vertical resolution when simulating coastal LLJs. This is especially true given the fact that PBL parameterization has been found to be responsible for much more extreme wind speed errors. On a final note, other model components beyond those assessed in this appendix also have the potential to affect the performance of individual PBL schemes such as the lowest gridpoint level AGL [202].
Figure A.13: Wind speed as a function of height for the OKX radiosonde launch site. QNSE (left) and MYNN2 (right) modeled profiles are colored and sounding observations are denoted by a line with stars. Data valid July 17th, 2011 12Z.
Appendix B

Eulerian Dispersion Modeling with WRF-LES of Plume Impingement in Neutrally and Stably Stratified Turbulent Boundary Layers

B.1 Introduction

The predictive modeling of scalar transport and dispersion is significantly more difficult in regions of sloping terrain compared to regions of flat homogeneous terrain. This increased difficulty is primarily attributed to two additional elements of complexity introduced by the presence of variable topography. One source of uncertainty is associated with the unique turbulent diffusion features commonly induced by sloped terrain, for example, in hill recirculation zones [96]. In some cases, these features can enhance localized scalar deposition through trapping, or alternatively promote mixing, thereby modifying the spatial dimensions and core concentration of a scalar plume with respect to transport distance [79, 135]. Another added element of uncertainty is potential flow redirection by the terrain [209]. For a point source scalar, and associated downstream plume, this redirection can vastly alter the trajectory of the plume under variable winds (e.g., plume impingement, pooling in basins, channeling through valleys, etc.) [21, Ch. 6]. The level of uncertainty contributed by these two elements is dependent upon a number of critical factors such as terrain slope, source location with respect to obstacle locations, and...
Despite the above challenges, regulatory dispersion modeling systems (e.g., COMPLEX1, CTDM, and AERMOD) have experienced great improvements in accuracy over the past few decades with regards to dispersion in complex terrain. Furthermore, the value of such analytical models has also been compounded by their relatively low computational cost and execution time. However, the empirical relationships used in the analytical models have been shown to demonstrate certain weaknesses when dealing with heterogeneous terrain such as 1-hr to 3-hr concentration errors of a factor of 2 or 3 [139]. Moreover, concentration biases have been found to have several sensitivities. For example, tendencies to over (under) predict mean areal concentrations under neutral (stable) stratification were attributes of the early Complex Terrain Dispersion Model (CTDM) while the position of scalar sources with respect to the dividing streamline was also associated with biases in CTDM [215]. These issues motivate the evaluation of a new and improved generation of physically-based models which are also capable of capitalizing on the recent growth in computational power. One such class of models is large-eddy simulation (LES) models, which have the ability to realistically simulate turbulent atmospheric boundary layer (ABL) flow down to spatial resolutions on the order of 10 m or less [159, 12].

Currently, the extension of LES to scalar dispersion and transport modeling has shown promise in both homogeneous [235, 168] and heterogeneous terrain situations [151, 113]. Over sloping surfaces especially, where spatio-temporally varying turbulence regimes are expected, the use of LES versus simpler dispersion models offers the advantage of potentially capturing a vast spectrum of non-linear microscale flow features [241]. These include lee recirculation and terrain induced turbulence enhancement, which are known to impact time averaged scalar distributions [50]. Given these capabilities, many researchers are exploring use of LES in addressing some historically difficult, yet high priority, problems such as scalar transportation in urban settings [163, 140].

While the allure of LES-based approaches to dispersion modeling is clear, they are nonetheless handicapped by several outstanding problems, some of which have long plagued other dispersion modeling approaches. Most notably, the accuracy of LES models, like all computational fluid dynamics models, are absolutely reliant upon accurate input forcing conditions e.g., wind speed, wind direction, thermal stability [221]. For dispersion timescales, these input conditions may vary highly with time making it difficult to prescribe realistic forcing conditions using coarse temporal resolution input data e.g., hourly averaged wind and temperature profiles [78]. This issue may be even more problematic when simulating flow near hills and other bluff-bodies [173, 17]. Therefore, errors introduced by a lack of knowledge of boundary conditions can degrade the accuracy of LES-based dispersion simulations. This concept has been supported by
high fidelity laboratory tank experiments and numerical simulations which have demonstrated that the inability to capture meandering wind direction can easily lead to predicted concentration overestimates of factors of four to ten along the plume centerline [210, 117, 68]. At the same time, concentration underestimates can also be produced at locations off of the plume centerline.

In this appendix, we implement passive tracers into the Weather Research and Forecasting - Large Eddy Simulation (WRF-LES) model in order to simulate two realistic field experiments from the Cinder Cone Butte (CCB) dispersion field study [126, 214]. The WRF-LES model was selected for this study for multiple reasons but primarily because it is freely accessible and therefore used by more than 20,000 users worldwide (Weather Research and Forecasting Model Homepage, Accessed: 2014). This popularity stems also from the fact that the mesoscale WRF model has demonstrated skill in short to medium range weather forecasting and because of its extensive data assimilation capabilities [120]. In addition, the robustness of the WRF model has been tested for conditions ranging from turbulence-scale [158] to global climate-scale [138]. Despite challenges linked to multiscale coupling, initial attempts to generate coupled mesoscale and microscale atmospheric simulations are encouraging [245, 155]. For dispersion modeling in the planetary boundary layer, this presents great opportunity since variability in large scale flows influences microscale flow features. In this study, we evaluate the ability of the non-coupled (i.e., only microscale) WRF-LES model to simulate accurate scalar transport and dispersion over complex terrain. This work represents an exploratory effort to assess the potential strengths, weaknesses, and overall value of WRF-LES for modeling plume impingement onto an isolated hill in the presence of neutral and stable stratification. This work is valuable as it presents the first documented effort, to the knowledge of the authors, to apply WRF-LES to stable boundary layer flows and to the problem of turbulent transport and dispersion in an area influenced by complex terrain features.

B.2 Field Experiments and Simulation Details

In this section, we detail the CCB field study and discuss the specific hours which our results focus on. Additionally, we introduce the WRF-LES modeling system and describe the model configuration and simulation specifics.

The Cinder Cone Butte field campaign took place in 1980 and was designed to study the complex problem of plume impingement onto a bluff body (e.g., a hill) under various stability regimes. The CCB field study was part of a much larger initiative carried out by the Environmental Protection Agency (EPA) designed to develop and improve numerical models (e.g., Complex Terrain Dispersion Model; CTDM) suitable for estimating concentration distributions in regions of complex terrain. This effort included field studies (e.g., Cinder Cone Butte,
Hogback Ridge), laboratory experiments, and theoretical studies.

All CCB experiments were undertaken during various nighttime hours in order to represent multiple stability classes. Each experiment involved the continuous release of passive tracer gases from a mobile crane upstream of the butte. In parallel surface concentration measurements were recorded at various locations on and near the butte. CCB itself is a cone-shaped, nearly symmetrical hill near Mountain Home, ID bounded by relatively flat, homogeneous terrain in all directions (i.e., ~10 km radius). It is roughly 1 km in diameter and rises approximately 100 m above its surroundings with a surface roughness length of ~1 m (Figure B.1).

In this appendix, we numerically simulate two specific CCB experiment hours, or cases, using a modified version of the WRF-LES model complimented with Eulerian dispersion (discussed in Section B.2.3). The first case was characterized by a nearly neutral stability profile and high geostrophic wind speed while the second case was moderate to strongly stable with weaker wind speeds (Figure B.2). The differences between these two cases enabled us to study the versatility of the LES dispersion modeling system in conditions with different buoyancy forces.

Figure B.1: Overhead view (left) and view from the South (right) of the topography of Cinder Cone Butte, Idaho.

B.2.1 Neutral Experiment

The first case of our study, henceforth referred to as the neutral case, was recorded as CCB experiment 218 and took place on November 12th, 1980 and lasted from about 2AM - 10AM (local time). During this time period, the experiment hour lasting from 4AM - 5AM was designated in this study to be the neutral case. This hour was selected for a number of reasons;
first, the wind direction in the lowest 175 m AGL was measured by an upstream tethersonde to be consistently from the Northwest (∼315°) during this time period with a geostrophic wind greater than 12 m s\(^{-1}\). Second, the stability maintained relative neutrality in the lowest levels of the atmosphere with a dry bulb temperature of about 274 K. The atmospheric state was relatively consistent during this hour making it attractive for our modeling study because it allowed us to neglect large scale changes and focus on the microscale turbulent effects on the dispersion of the tracer. Lastly, concentration measurements from this case were of high quality with few measurement system malfunctions.

The CCB experiments used a cylindrical coordinate system focused on the center of the butte. Source release positions were defined according to radial distance (r) and azimuth (ϕ; 0° was due North) in relation to the center of CCB. In addition, the source location was elevated at a specific height above ground level (z) at r and θ. For the experiment of interest here, the release location \((S(R, \theta, z)) = S(1119 \text{ m}, 327.3^\circ, 30 \text{ m})\) and the release rate of \(SF_6\) was 0.156 g/s ± 1.2%. Additionally, during this experiment hour \(CF_3Br\) (i.e. Freon) was released from the same location except at 45 m above ground level (AGL) and at a flow rate of 0.831 g/s ± 2.7%. However, for this work we chose to focus our analysis on \(SF_6\) since its measured distribution was more complete and encompassed a much larger range of concentrations.

For the neutral case, scalar concentration measurements were recorded by approximately 47 bag samplers dispersed across CCB. The average concentration measurements collected from the neutral case hour are shown in Figure B.3 (left panel). In this case, the surface layer wind was approaching the butte from the northwest and high tracer concentrations were
Figure B.3: (Left panel) Measured concentrations of $SF_6$ at CCB during experiment 218, hour 5 (i.e. the neutral experiment). (Right panel) Measured concentrations of $SF_6$ at CCB during experiment 215, hour 4 (i.e. the stable experiment). Contours and labels represent terrain height while colored circles represent measured concentrations. The black crosses symbolize the release locations of the $SF_6$ tracer.

recorded in a narrow diagonal zone from northwest to southeast. The highest concentrations were recorded along the northern tier of CCB. Additionally, there was a small area of localized high concentration on the downwind face of CCB which may have been due to boundary layer separation there.

B.2.2 Stable Experiment

The second case of our study, henceforth referred to as the stable case, was recorded as CCB experiment 215 and took place on November 6th, 1980 and lasted from about 12AM - 7AM (local time). For the stable case, we focus on experiment hour 4 lasting from 3AM - 4AM. The stable case had similar appealing characteristics as with the neutral case (e.g., comprehensive concentration measurements); however, strong overnight radiational cooling had formed a classic stable surface layer (i.e., there was not a complex multi-layered thermal profile). This stratification also produced a vertical wind profile that was challenging to model. As can be seen in Figure B.2, the wind direction in the lowest 50 meters veered by nearly 90 degrees. Above this thin veering layer, the wind direction was from Southeast with little change in direction with height. In addition, near the surface the wind speed was very weak with possible intermittent calm conditions. The source location for the tracer release was $S = S(1033 \text{ m}, 107.2^\circ,30 \text{ m})$ and the release rate was $0.190 \text{ g/s} \pm 3.2\%$.

Measurements reveal that the region of highest tracer concentration around CCB in this stable case was much more localized than with the neutral case (Figure B.3, right panel). The area of highest concentration is on the upwind face of the butte with some spillage around the
flanks of CCB. Also, there is no maximum of concentration at the peak of the butte nor is there significantly enhanced concentrations on the downwind face. This distribution appears to be an artifact of the thermal stability and adheres to what would be expected in the case of a scalar release below the dividing streamline [95]. In other words, the stratification prevents the scalars from predominately ascending over the crest of the butte and instead they are diverted laterally around the obstacle. Further analysis of the flows associated with both the neutral and stable cases will be discussed in Section B.4.

B.2.3 Numerical Dispersion Modeling System

In this study, we present simulation results from the WRF-LES model with a modification to include passive tracer transport and dispersion. The WRF model is a finite difference computational fluid dynamics system which is widely-used in the numerical weather prediction community. While conventional WRF model simulations (i.e. mesoscale simulations) use coarse resolution and consequently turbulence, or planetary boundary layer, parameterizations, WRF-LES explicitly resolves the most energetically significant turbulent eddies. This is a beneficial property for simulations of dispersion and scalar transport as it provides transient characterizations of microscale turbulent fluctuations and structures (i.e. updrafts, rolls, velocity streaks, meandering, etc.) [110, 58]. Individually and collectively, these features can heavily influence the behavior of passive scalars. While the WRF-LES model can resolve many of the turbulent structures within the boundary layer explicitly, the representation of the sub-grid scale fluxes and transfer of turbulent kinetic energy must be parameterized by a SGS closure model. Here we comment on results obtained by invoking two different versions of SGS closure: [1] a linear eddy viscosity closure scheme based on a prognostic equation for turbulent kinetic energy known as the Deardorff scheme (abbreviated DDF; [51]) and [2] a non-linear eddy viscosity scheme which incorporates energy backscatter and anisotropy (abbreviated NBA; [114]). These two subgrid models are included in publicly released version of WRF and detailed comparison of their performance in WRF-LES is described in [153]. Essentially, while the DDF model is absolutely dissipative (i.e., acts as a sink of turbulent kinetic energy everywhere), the NBA model allows for local backscatter of energy and in general produces more realistic resolved turbulent fields. Also, it should be noted that this work includes results from a stable boundary layer (SBL) simulation which is, as far as the authors can tell, the first of its kind in WRF-LES. Stably stratified turbulence is historically much more difficult to model using LES approaches. This is due to the fact that the relative turbulent length scales are much smaller in SBL flows than in neutrally stratified and convective flows; at the same time, the shape of turbulent eddies also becomes compressed vertically in SBL turbulence.

Simulations were run for a total of 7 hours; the first six hours were used to ”spin-up” realistic
turbulence and develop appropriate mean profiles of temperature and momentum. Results were taken from the seventh hour and were compared to the hour averages for both the neutral and stable cases. The source of the passive scalar \( (SF_6) \) was defined to be a plume source; in other words, the scalar was introduced into the domain at one specific grid cell and was then continuously emitted for the entire 1 hour of interest in a single source grid cell in units of \( \text{kg m}^{-2} \text{s}^{-1} \) within the source grid cell. This treatment of the source and transport and dispersion is essentially identical to that in [140]. Inside the source grid cell, we assume a spatially uniform emission rate with a total grid cell emission rate equal to the emission rate associated with the experiments in units of mass per unit time. Since the tracer was passive, the spatio-temporal concentrations and emission rates were scaled against the actual source emission rates through post processing.

The modeling dimensions used in this study are described in Table B.1. The inner most domain had a uniform horizontal terrain following mesh with a horizontal resolution of 10 m and a vertical resolution of 10 m in the lowest 2/3 of the domain. Additionally, measures were taken to maintain the numerical stability of the simulations which included prescribing a Rayleigh damping layer near the top of the domain along with relaxation of vertically propagating acoustic waves. In order to simulate turbulent inflow into the domain, unaffected by the complex terrain surrounding CCB, two domains were used with one nested inside the other. Both domains resolved turbulence; however, the outer domain (d01) used a “flat-plat” lower surface boundary with periodic boundary conditions along the longitudinal and transverse boundaries. Domain d01 provided turbulent inflow boundary conditions for the smaller inner domain (d02) which used real (i.e., Cinder Cone Butte) terrain as the lower surface boundary. Most importantly, d02 did not feed information back to d01 as that would have caused the wake effects of CCB to be recycled into the inflow turbulence. Using this technique, we effectively provided the target domain (i.e., d02) with realistic turbulent inflow free from anomalous terrain-induced flow features. Further details on this technique can be found in [154].

<table>
<thead>
<tr>
<th>Domain</th>
<th>( \Delta x, y )</th>
<th>( \Delta t )</th>
<th>( N_z )</th>
<th>Dimensions ([x, y, z (\text{neutral,stable})])</th>
</tr>
</thead>
<tbody>
<tr>
<td>d01</td>
<td>30 m</td>
<td>( \frac{1}{4} ) s</td>
<td>98</td>
<td>6 km ( \times ) 6 km ( \times ) (2, 1.2) km</td>
</tr>
<tr>
<td>d02</td>
<td>10 m</td>
<td>( \frac{1}{24} ) s</td>
<td>98</td>
<td>4 km ( \times ) 4 km ( \times ) (2, 1.2) km</td>
</tr>
</tbody>
</table>

For initialization of the WRF-LES simulations, vertical profiles of constant wind velocity and potential temperature were prescribed for the neutral case. Before forward time integration
began, the flow was essentially laminar and as time evolved, surface stress developed a realistic boundary layer complete with turbulence. For the stable case simulation, uniform wind velocity was also prescribed while the potential temperature profile was set to the observed potential temperature profile above 100 m agl. Below 100 m, there was initially no vertical potential temperature gradient but a negative surface skin temperature and sensible heat flux were prescribed so that the thermal profile matched the observed profile after the turbulence spin-up time period\(^2\). Obviously, realistic wind and stability forcing data is difficult to measure and also difficult to prescribe in the simulations. The method used here assumed stationary flow for the neutral case and quasi-stationary flow for the stable case with the surface cooling being the only dynamic forcing in the latter. However, in reality, ABLs are rarely perfectly stationary which presents one of the major shortcomings of the modeling approach used here. Coupling WRF-LES with larger domain mesoscale simulations would potentially provide more realistic forcing. Such an approach has recently been explored by [155] but is beyond the scope of this current study.

### B.3 Neutral Case Results

In this section, we evaluate the performance of the WRF-LES modeling system for the neutrally stratified case described in Section B.2.1. First we compare the mean wind and thermal profiles generated by the model to those measured by the tethersonde. In addition, we discuss some of the complex physics captured by the model (i.e., longitudinal rolls). Then, we compare the modeled concentrations to the observed concentrations using contour plots overlaid with measurements, Quantile-Quantile (QQ) plots, and a suite of standard statistical accuracy metrics. For consistency, the statistical analyses performed in this appendix generally follow those used by [13] in a previous dispersion modeling study of CCB and those presented in the EPA CTDM final report. Based on the comparison, we comment on the strengths and weaknesses of the WRF-LES dispersion modeling system and its sensitivity to sub-grid scale parameterization.

Since the neutral case was characterized by a nearly uniform potential temperature profile, a surface heat flux was not prescribed. This allowed for easy replication of the temperature profile. Fortunately in this case, the lack of surface heat flux meant that directional wind shear was minimal in the surface layer. Nevertheless, the simulated plume impingement on CCB proved to be extremely sensitive to wind direction and therefore much care was taken to closely align the steady state wind profile to the observed hourly mean wind profile. In Figure B.4, the WRF-LES steady state vertical profiles are shown with respect to the tethersonde measurements. The modeled wind speed was similar to the observations in the lower half of the

\(^2\)Surface skin temperature and heat flux was ingested into the model using WRF’s auxiliary data input stream functionality.
surface layer and then a slight overestimate was recorded above that. Since the measurements represent a profile from one particular point near CCB and the simulated profile is an average over the flat outer domain, it is expected that some slight differences will be present near the top of the surface layer as larger turbulent effects there can be heavily dependent upon surface heterogeneity.

As discussed earlier, one of the most attractive benefits of LES-based dispersion modeling, compared to conventional techniques such as Gaussian-plume models, is the ability of LES to explicitly resolve large turbulent motions which can transport scalars in ways not describable by less sophisticated techniques, an example being longitudinal rolls [131]. Longitudinal rolls are secondary circulations embedded within the turbulent ABL flow and are manifestations of thermal and inflectional instabilities in the flow. In the neutral case simulation, evidence of the presence of such rolls was found due to the lateral meandering of the simulated plume. In Figure B.5 the plume can be seen traversing CCB on opposite flanks at two different times.
Similar plume flip-flopping behavior has been documented in the past, notably in laboratory experiments carried out for CCB cases [210]; however, their physical dynamics are not entirely understood in the literature. That being said, the fact that lateral plume meandering was captured by the WRF-LES model demonstrates one of its potential strengths in reproducing physically realistic plume behavior.

Figure B.5: Instantaneous tracer visualization displaying plume meandering from one side of the butte to the other at two different times with steady-state boundary conditions.

As stated in Section B.2, measured $SF_6$ concentrations were averaged over each experiment hour. In Figure B.6, concentrations from the WRF-LES simulation were averaged over the hour and compared to the $SF_6$ point measurements. The mean trajectory of the plume was well captured as it crossed CCB in a northwest-to-southeast orientation and aligned with the swath of highest observed concentrations. The lateral diffusion of the plume with respect to distance from the source was also similar to what was measured by the bag-samplers (i.e. about half of the diameter of CCB), although it appears that the simulated plume was slightly wider than what was indicated by measurements. This could perhaps be a result of numerical diffusion associated with the Eulerian-based modeling framework.

As with most dispersion modeling endeavors, the specific magnitude of the simulated $SF_6$ concentrations across CCB were highly variable. That being said, the WRF-LES model underestimated the highest scalar concentrations in the core of the plume by about a factor of two and overestimated the concentrations near the peak of CCB, to the south of the plume centerline, by about 100 PPT. While the averaged concentrations depict a nearly uniform plume of decreasing intensity with distance from the source (Figure B.6), instantaneous simulated tracer surface concentrations (not shown here) show very localized high concentration regions which propagate over CCB in unique patterns. This transient behavior attests to the complications associated with modeling plume impingement against a terrain obstacle. Despite this challenge,
Figure B.6: Plan view of simulated tracer concentrations (contours) and measurements (colored circles) averaged over 1 hour. The left panel corresponds to the DDF-SGS simulation and the right panel corresponds to the NBA-SGS simulation.

The large-eddy simulation still demonstrated skill in capturing the 1-hr average point-by-point concentrations.

Figure B.7: Quantile-Quantile plot of observed $SF_6$ concentrations ($C_o$) versus predicted concentrations ($C_p$) as simulated by the DDF-SGS (left) and NBA-SGS (right) for the neutral case. Black dashed line represents 1-1 correlation while green dashed lines represent the bounds of data within a factor of two of the observations.

In Figure B.7, Quantile-Quantile (QQ) plots demonstrate the correlation of the simulated concentrations with the measured concentrations. Note that the observed concentrations below
5 PPT were set to 5 PPT since that was the lower concentration measurement threshold of the bag samplers. Figure B.7 reflects the fact that the WRF-LES model over estimated the lower magnitude concentrations while at the same time under estimated the higher magnitude concentrations. This behavior was observed in other models (e.g., CTDM) by the EPA in their model evaluation of CCB. In the current simulated data, the low magnitude over estimation bias appears to be primarily due to the fact that the width of the plume was over predicted while the high magnitude under estimation bias could be an artifact of enhanced diffusion associated with Eulerian dispersion in the plume core. In addition, out of the two SGS schemes tested here, the NBA scheme generally produced a more horizontally compact plume which seemed to agree better with the measurements. This property slightly reduced the underestimation bias associated with the high concentration points in the plume core. The fact that the NBA scheme produced a more compact and focused plume whereas the DDF scheme produced a slightly more diffused plume, may perhaps be related to the less pronounced longitudinal roll signatures observed in the NBA simulations (not shown here). This result has previously been observed and mentioned by others [114, 153].

Table B.2: Statistical accuracy metrics for the neutral case simulation. $C_o$ and $C_p$ denotes the observed and predicted concentrations, respectively. Overbars represent averaging.

<table>
<thead>
<tr>
<th>Metric</th>
<th>Formulation</th>
<th>DDF</th>
<th>NBA</th>
</tr>
</thead>
<tbody>
<tr>
<td>Bias (PPT)</td>
<td>$\bar{C_p} - \bar{C_o}$</td>
<td>-39.286</td>
<td>-24.260</td>
</tr>
<tr>
<td>Root Mean Squared Error</td>
<td>$\sqrt{\frac{(C_p - C_o)^2}{C_p}}$</td>
<td>2.028</td>
<td>1.457</td>
</tr>
<tr>
<td>Linear Pearson Correlation Coefficient</td>
<td>$\frac{(C_o - \bar{C_o})(C_p - \bar{C_p})}{\sigma_{C_p}\sigma_{C_o}}$</td>
<td>0.796</td>
<td>0.767</td>
</tr>
<tr>
<td>Fraction of data within a factor of 2</td>
<td>$\frac{1}{2} \leq \frac{C_p}{C_o} \leq 2$</td>
<td>0.558</td>
<td>0.519</td>
</tr>
<tr>
<td>Fraction of data within a factor of 5</td>
<td>$\frac{1}{5} \leq \frac{C_p}{C_o} \leq 5$</td>
<td>0.865</td>
<td>0.827</td>
</tr>
<tr>
<td>Fraction of data within a factor of 10</td>
<td>$\frac{1}{10} \leq \frac{C_p}{C_o} \leq 10$</td>
<td>0.981</td>
<td>0.961</td>
</tr>
</tbody>
</table>

In this study, we used a suite of statistical comparison metrics which are standard in trans-
port and dispersion modeling (e.g. [80, 233]); these metrics are presented in Table B.2. The simplest and perhaps the most descriptive of the metrics presented in Table B.2 are the fractions of data within a factor of 2, 5, and 10 of the measurements. For the most part, the WRF model simulated about 52% to 56% of points within a factor of two of the measurements. While this range of error leaves significant room for improvement, the WRF-LES model’s performance for this case is roughly in line with the current level of accuracy of achieved by alternative dispersion modeling platforms. Moreover, an EPA evaluation of the CTDM model (i.e., the predecessor of AERMOD) for this particular CCB case computed $\frac{\overline{H_{5p}}}{\overline{H_{5o}}} = 0.505$ where $H_{5p}$ and $H_{5o}$ represent the the highest 5 predicted and observed concentrations respectively. The DDF-SGS and NBA-SGS WRF-LES model simulations presented here were competitive with $\frac{\overline{H_{5p}}}{\overline{H_{5o}}} = 0.421$ and 0.510, respectively.

In addition, the linear Pearson correlation coefficient ranged between 0.76 and 0.80 which was slightly better than 0.73, a value achieved by a previous attempt to model a different CCB experiment by [13] using a Reynolds-Averaged Navier Stokes model. In their study, [13] also reported strong sensitivity to background wind direction, as was the case in the WRF-LES model runs presented here.

### B.4 Stable Case Results

In addition to the neutral case, we also evaluated the performance of the WRF-LES dispersion modeling system for a significantly more complex, stably stratified case. The contents of this section are similar to those in Section B.3; however, more attention is paid to the physical phenomena related to the stratification (e.g., flow splitting, etc.) and the challenges it presents to the model accuracy.

The simulated 1-hr averaged profiles of wind speed, wind direction, and temperature for the stable case are shown in Figure B.8. In order to reproduce the unique mean lapse rate recorded by the tethersonde during CCB experiment 215, a negative (i.e., downward) sensible heat flux boundary condition was applied at the surface which decreased with time (approx. -1 W/m²hr) starting from -0.01 W/m² at 30 minute intervals during the spin-up and simulation time. At the same intervals, the surface skin temperature cooled at a rate of approx. 1.14 K/hr starting from 292.67 K. This surface cooling boundary condition was implemented in order to mimic the natural radiational cooling which occurred during the real experiment. The one hour mean potential temperature profile showed general agreement with that which was observed; however, the simulated wind conditions within the lowest 30 m did not have as strong of directional and magnitudinal wind shear as measured by the tethersonde. Low wind speed, combined with strong directional shear, in this layer near the surface seemed to indicate the presence of a non-classical stable boundary layer. That is, the very stable surface layer was
likely not idealized, and instead varied throughout the experiment hour, thereby producing the complex mean wind velocity profile. Such behaviors are not unexpected in stable boundary layers [94] and have been thought to be driven by several multiscale processes such as mesoscale meandering, gravity waves, and turbulence decay [68, 144]. While these processes are very important for accurately predicting plume impingement in the stable boundary layer, many of them unfortunately originate outside of our 4 km by 4 km LES domain. This highlights a weakness of using quasi-idealized boundary conditions as forcing for WRF-LES-based dispersion and also emphasizes the potential value of a successfully coupled mesoscale-LES dispersion model.

Figure B.8: WRF-simulated (blue) and observed (red) hourly mean vertical profiles of wind-speed, potential temperature and wind direction for the stable case. The WRF profiles represent averages over the outer domain.

Figure B.9 depicts the simulated mean $SF_6$ concentrations overlaid with the observations. It is clear that the stable stratification resulted in significant horizontal distortion to the plume trajectory compared to the neutral case. With a source location of 30 m AGL, the plume lacked sufficient momentum to significantly overcome the stable buoyancy and completely cross the butte vertically as it did in the neutral case simulation. Instead, much of the near surface plume
diverted around the flanks of the butte. The suppression of vertical motion due to the thermal stability resulted in reduced scalar transport up to the crest of CCB, which was approximately 75 m above the release height. This lack of vertical transport was also documented by several smoke visualizations in the CCB field experiments and also in towing tank laboratory experiments [126].

![Plan view of simulated tracer concentrations (contours) and measurements (colored circles) averaged over 1 hour for the stable case as simulated by the DDF-SGS simulation (left panel) and the NBA-SGS simulation (right panel). Note that the color scheme is logarithmic.](image)

Figure B.9: Plan view of simulated tracer concentrations (contours) and measurements (colored circles) averaged over 1 hour for the stable case as simulated by the DDF-SGS simulation (left panel) and the NBA-SGS simulation (right panel). Note that the color scheme is logarithmic.

Another feature of the plume which was affected by the stability was its lateral dispersion. Compared to the neutral case simulation, the stable case simulation had a plume that was significantly wider as was documented in the physical CCB experiment [126]. This was due to the flanking of the plume around the butte and also due to the predominance of horizontal dispersion over vertical dispersion. In addition, the horizontal size of the plume simulated by the NBA model was noticeably larger than the one simulated by the DDF model; this could be evidence of increased turbulence-scale meandering in the NBA scheme compared to the DDF scheme. This is because while DDF model is absolutely dissipative (dissipative everywhere), NBA model includes potential for local backscatter of energy and therefore potential for more energetic resolved eddies that cause plume meandering.

In the stable case, the scalar source position was to the east-southeast of the butte. The east-southeast face of CCB contains a subtle concave draw in the relief (Figure B.3 right panel) which appears to be collocated with the peak concentrations measurements (i.e., the impingement zone). This draw may have influenced the scalar pooling location and magnitude by inducing a weak wind funneling effect. Comparing the simulated $SF_6$ concentrations with the observations, both schemes generated large prediction overestimates near the impingement zone (note the
In the impingement zone, the SF$_6$ samplers recorded hourly mean concentrations of 700 - 800 PPT while the WRF-LES model simulated average concentrations over 2000 PPT in these locations. The DDF scheme generally had the larger overestimation bias in the impingement zone while the NBA scheme had an overall larger overestimation bias integrated across the entire butte. This bias may have been associated with the light and potentially variable winds which occurred during this night; following the mesoscale meandering concept, there may have been times when the actual plume did not impact CCB, while in the WRF-LES model the plume always impacted CCB. This could have led to concentration over estimates.

The tendency of the WRF-LES model to produce a significant over estimation concentration bias for this stable case, and not in the neutral case, is reminiscent of what was generally observed for several analytical models considered in the final report of the complex terrain dispersion model development [215]. Moreover, in the simulations presented here, this bias is possibly due to unaccounted for mesoscale lateral plume meandering, which was potentially very active during the real experiment. [78] highlighted the prominence of the meandering phenomenon for many of the Cinder Cone Butte experiments and others have shown that its neglect in numerical models can support over prediction errors on the magnitude of what is reported here [210, 68]. Fruition of this hypothesis also helps to explain the discrepancies between the measured and simulated hourly mean wind speed and direction profiles previously mentioned.

![Quantile-Quantile plot](image)

Figure B.10: Quantile-Quantile plot of measured SF$_6$ concentrations versus DDF simulated (left) and NBA simulated (right) concentrations for the stable case. Black dashed line represents 1-1 correlation while green dashed lines represent the bounds of data within a factor of two of the observations.
Despite the positive 1-hr averaged concentration bias, the qualitative distribution of the simulated concentrations generally agreed with the observed concentration distribution Figure B.9. That being said, a slight inverse of the distribution bias observed with the neutral case was found in the stable case simulations. That is, the over estimation bias was greatest in the high magnitude regime and smallest in the low concentration regime, with another increase in over estimation with the lowest 4-5 concentration samples (Figure B.10). Table B.3 shows the statistical accuracy of the two simulations with respect to the observations. Compared to the neutral case, this case witnessed a substantial decrease in the percentage of samples within a factor of two of the observations with 23% and 26% in the DDF-SGS and NBA-SGS runs, respectively. Also for this case, RMSE’s were around 1.2 and 1.3, the former of which was recorded by [13] for a different stably stratified CCB case. Furthermore, [13] also documented localized high concentration magnitude over estimates near the plume impingement location on CCB, similar to what was found here with WRF-LES. Also, for the stable case simulated here \( \frac{H_5}{H_5} = 4.439 \) and 4.890 for the DDF-SGS and NBA-SGS simulations, respectively. For the CTDM, \( \frac{H_5}{H_5} = 1.185 \) for this particular case; however, for other stably stratified CCB cases \( \frac{H_5}{H_5} \) up to 35.321 were found. Furthermore, it should be noted that the analytical expressions which went into the development of CTDM (and later AERMOD) were based on observations from the CCB experiment. Therefore, the performance of these regulatory models for CCB cases is somewhat biased.

B.5 Conclusion and Future Directions

In this work, the WRF-LES model was evaluated based on its ability to simulate scalar transport and dispersion in ABLs influenced by complex sloping terrain. The WRF-LES model was verified against two experiments of plume impingement conducted under the Cinder Cone Butte field campaign which were representative of neutrally and stably stratified atmospheric conditions. This represents the first documented attempt to simulate dispersion in a stably stratified turbulent ABL using WRF-LES. While the simulations were challenged by uncertainty of large-scale forcing, they demonstrated skill similar to that of competing modeling techniques when validated against hourly mean concentration measurements. In addition, the simulations demonstrated several intricate scalar phenomena such as microscale plume meandering and realistic plume responses to stability regime.

In both the neutral and stable cases, the simulated plume trajectories, for the most part, agreed well with concentration measurements. However, a positive magnitude bias was observed in the stably stratified case which seemed to be due to weak and variable near surface winds possibly associated with the neglect of mesoscale lateral meandering [78, 68, 142]. Nevertheless, the statistical accuracy achieved in this work was similar to previous physical [13] and analytical

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Table B.3: Statistical accuracy metrics for the stable case. \( C_o \) and \( C_p \) denotes the observed and predicted concentrations, respectively. Overbars represent averaging.

<table>
<thead>
<tr>
<th>Metric</th>
<th>Formulation</th>
<th>DDF</th>
<th>NBA</th>
</tr>
</thead>
<tbody>
<tr>
<td>Bias (PPT)</td>
<td>( C_p - C_o )</td>
<td>570.537</td>
<td>619.862</td>
</tr>
<tr>
<td>Root Mean Squared Error</td>
<td>( \sqrt{\frac{(C_p - C_o)^2}{C_p}} )</td>
<td>1.232</td>
<td>1.266</td>
</tr>
<tr>
<td>Linear Pearson Correlation Coefficient</td>
<td>( \frac{(C_o - \overline{C}<em>o)(C_p - \overline{C}<em>p)}{\sigma</em>{C_p}\sigma</em>{C_o}} )</td>
<td>0.353</td>
<td>0.590</td>
</tr>
<tr>
<td>Fraction of data within a factor of 2</td>
<td>( \frac{1}{2} \leq \frac{C_p}{C_o} \leq 2 )</td>
<td>0.226</td>
<td>0.264</td>
</tr>
<tr>
<td>Fraction of data within a factor of 5</td>
<td>( \frac{1}{5} \leq \frac{C_p}{C_o} \leq 5 )</td>
<td>0.509</td>
<td>0.585</td>
</tr>
<tr>
<td>Fraction of data within a factor of 10</td>
<td>( \frac{1}{10} \leq \frac{C_p}{C_o} \leq 10 )</td>
<td>0.736</td>
<td>0.811</td>
</tr>
</tbody>
</table>

[215] dispersion modeling attempts in the presence of a terrain obstacle. Correlation coefficients up to 0.80 were achieved in the neutral case, which were near or slightly better than the accuracy achieved by [13] of 0.73 for a specific stable case from the CCB experiment. For the same case, [13] reported a RMSE of 1.20 which was similar to the values of RMSE recorded for this particular stable case, but at the same time they highlighted the fact that their simulation accuracy was highly sensitive to wind direction as it evidently was here also. Furthermore, values of \( \frac{H_{5p}}{H_{5o}} \) ranged from 0.421 and 0.510 for the neutral case which was similar to the accuracy achieved by the CTDM for the same CCB case. However, for the stable case \( \frac{H_{5p}}{H_{5o}} \) ranged from 4.439 to 4.890 which was considerably higher than that observed by CTDM of 1.185 for the same stably stratified CCB case. On the other hand, it should be noted that the analytical expressions that went into the development of several regulatory dispersion models (i.e., CTDM and its decedent, AERMOD) were based on observations from CCB cases which makes their performance for those cases biased.

Like many other models, the accuracy WRF-LES dispersion simulations is perhaps the most constrained by the notorious plume sensitivity to ambient wind speed and direction fluctuations. Future studies should continue to explore the concept of coupled mesoscale-LES modeling, with both one-way and two-way feedback, and its capability to reduce the uncertainty of dynamic boundary conditions. In addition, investigations should be carried out to assess whether this
WRF-LES Eulerian modeling approach results in significant unphysical numerical diffusion and consequently inaccuracies in dispersion solutions; a problem which could potentially be remedied with the use of an offline Lagrangian particle model. Moving forward, these issues should be better understood and mitigated in order to improve quantitative model accuracy. Nevertheless, this evaluation indicates that a model like WRF-LES could potentially be a good platform for microscale transport and dispersion simulations over complex terrain.