Utilizing large-eddy simulation in high-resolution wind resource modelling : a coastal low-level jet case study

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2017

Tay, K. (2017). Utilizing large-eddy simulation in high-resolution wind resource modelling : a coastal low-level jet case study. Doctoral thesis, Nanyang Technological University, Singapore.

<http://hdl.handle.net/10356/72652>

<https://doi.org/10.32657/10356/72652>

NANYANG
TECHNOLOGICAL
UNIVERSITY

**UTILIZING LARGE-EDDY SIMULATION IN
HIGH-RESOLUTION WIND RESOURCE
MODELLING - A COASTAL LOW-LEVEL JET
CASE STUDY**

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A thesis submitted to the Nanyang Technological University in partial
fulfilment of the requirement for the degree of
Doctor of Philosophy

2017

Abstract

The challenges of nesting large-eddy simulation (LES) within the Weather Research Forecast (WRF) model are examined and addressed in this work. Coastal low-level jet (LLJ) events in field campaigns CBLAST-LOW 2001 and NEAQS-ITCT 2004 are used as case studies. Nested LES models offer an advantage in LLJ modeling by encompassing a wide range of length scales. Broad mesoscale conditions that drives LLJ events are taken into account in the coarse outer domains while local turbulence interactions are resolved within the nested high-resolution LES model.

The accuracy of nested LES models are largely dependent on the mesoscale parent domains used. The sensitivity of the mesoscale domains to initial/boundary conditions (IBC), planetary boundary layer (PBL) scheme and sea-surface temperature (SST) forcing are characterized using a suite of diagnostic metrics. In the CBLAST-LOW 2001 case study, IBC was the greatest source of variability due to the differences in the representation of a warm frontal passage. For a hypothetical wind turbine located in the center of the domain, over the duration of the LLJ event, the variability in the simulations results in 5% difference in predicted wind power.

In the NEAQS-ITCT 2004 case study, a continuous 48-hour period that featured two different LLJ events, a WEAK LLJ ($\sim 10 \text{ m s}^{-1}$) and a STRONG LLJ ($\sim 18 \text{ m s}^{-1}$), was selected for analysis. All the simulations were able to successfully model both LLJ events to varying degree. The simulations underestimated the WEAK LLJ peak speed, which also arrived before the observed LLJ; whereas the STRONG LLJ peak speed was overestimated and lagged the observed LLJ. Regardless of IBC or PBL used, the simulations showed an average cold bias ($\sim 2 \text{ K}$) in the surface temperature when compared to ASOS data.

The normalized bias (NBIAS) and normalized root-mean-square error (NRMSE) of wind velocity prediction were used to score model performance. IBC was the greatest source of variability in wind velocity prediction during the WEAK LLJ event. Different SST forcings, which include warming the SST within the Gulf of Maine, had little to no impact on the NRMSE scores during this event. However, during the STRONG LLJ event, regardless of model configurations, the simulations scored similarly in NBIAS within the subject layer ($< 100 \text{ m}$). The NBIAS peaked at approximately the height of the LLJ core ($\sim 100 \text{ m}$). Above this layer, the variability induced by the choice of IBC, PBL and SST were much more pronounced. In both

WEAK and STRONG LLJ events, the maximum NRMSE occurs within the subjet layer, highlighting the need to improve parameterization near the surface layer. Based on the NBIAS and NRMSE scores, GFS_MYNN was the best performing configuration for the mesoscale simulation and hence selected as the parent domain for nested WRF-LES simulation.

Two sets of nested WRF-LES simulations were performed, one each for the two LLJ events identified. A new method to spin-up turbulence between mesoscale and LES domain was also tested. WRF-LES reduced the NBIAS and NRMSE for both WEAK and STRONG LLJ within the subjet and jet core layer as compared to the parent domain. Spectral analysis showed that the Cell-Perturbation Method (CPM) reduced the fetch required for the turbulence spectra to reach equilibrium. CPM was more effective in STRONG than WEAK. CPM was able to inject the missing turbulence scales from the transition between the mesoscale and LES domain when the horizontal resolution was sufficient to resolve most of the LES eddies. The improvement in turbulence representation due to CPM also resulted in further improvements in the NBIAS and NRMSE scores of the modeled LLJ. CPM was shown to be an effective tool for real nested WRF-LES studies.

Acknowledgements

If I learnt only one lesson over the course of my PhD, it is that I certainly did not complete this journey alone. This thesis would never come to fruition without any of the following people.

I will first like to thank my academic supervisors, Asst. Prof. Martin Skote and Assoc. Prof. Koh Tieh-Yong and my industrial supervisor, Dr. Johan Gullman-Strand for their invaluable guidance over the course of this journey. Thank you for being my pillars of support and making sure that I did not stray too far from my path.

I also wish to express my deepest gratitude to Assoc. Prof. Julie Lundquist and her wonderful research group at University of Colorado Boulder for hosting me. Though I was just a visiting researcher at CU for 10 months, she has taken me under her wing and mentored me as one of her own.

I have benefited tremendously from briefly being a part of the Boulder atmospheric community. Of all the researchers I have got to meet and interact with in Boulder, I will like to give special thanks to my collaborator, Dr. Domingo Muñoz-Esparza for generously sharing his work and insights with the LES runs.

Many thanks to my fellow RR2 friends for the countless tea-break sessions when we commiserate about life as a PhD student. To my second family in Boulder: the Heidelberg House, my fellow DUANE folks and my CRC gymbros; a very big thank you for making me feel at home in Boulder, my stay was very much enriched having had the opportunity to cross paths with you all. To my personal friends outside the bounds of my PhD life, your words of encouragement and listening ears have certainly eased this journey of mine.

In addition, none of this would have been possible without the support of Lloyd's Register (Global Technology Center), Energy Research Institute @ NTU (ERI@N) and the Interdisciplinary Graduate School (IGS). The resources and opportunities provided by these organizations are all a young researcher can ask for and more.

Last and most importantly, I will like to thank my mom, Serene, for always supporting this dream of mine.

-Clear eyes. Full hearts. Can't lose.-

List of Conferences and Publications

Thesis Chapter 2

Tay, K., Lundquist, J., Skote, M. and Koh, T-Y., “*Characterization of mesoscale variability in WRF - a coastal low-level jet case study*”, American Geophysical Union Fall Meeting, San Francisco, California, 15 - 19 December 2014.

Tay, K., Lundquist, J., Skote, M. and Koh, T-Y., “*Characterising mesoscale variability in low-level jet simulations for CBLAST-LOW 2001 campaign*”, (Under Preparation)

Thesis Chapter 3

Tay, K., Lundquist, J., Skote, M., Koh, T-Y. and Pichugina, Y.L., “*Mesoscale variability study of coastal low-level jets observed in NEAQS-ITCT 2004 campaign*”. (Under Preparation)

Thesis Chapter 4

Tay, K., Lundquist, J., Muñoz-Esparza, D., Skote, M. and Koh, T-Y., “*Nested WRF-LES of coastal low-level jets in New England*”. (Under Preparation)

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Chapter 1

Introduction

1.1 Background and Motivation

In this current era of climate change, there exists a strong impetus to carry out research into renewable energy resources and sustainable development. Alternative renewable energy is needed to supplement rising global energy need which has so far been reliant on fossil fuels, one of the chief causes of climate change. According to the European Wind Energy Association 69% of all new power capacity in 2012 installed in Europe is provided by renewable energy, wind energy installations alone representing 26% of the total new power capacity. Wind energy has long been used as a source of energy and has been proven to be a viable renewable source. However, despite our familiarity with this resource, the harvesting of wind energy is far from optimal or efficient. One of the various ways to address these shortcomings is to develop better tools to identify and evaluate areas suitable for wind harvesting.

Wind resource mapping requires a combination of satellite imagery, meteorological data from observation towers, weather balloons and numerical weather models to produce an accurate picture. Improvements in any of these areas will help to produce higher quality maps, in terms of accuracy and resolution, that can reflect the true potential of an area.

Studies conducted by DTU Wind Energy has shown how increasing spatial resolution is able to highlight more features of the wind field in numerical simulations of the Horns Rev wind farm. This can be seen in Fig. 1.1.

Fig. 1.2 shows how the effect of grid resolution impacts the amount of wind

Figure 1.1: Mesoscale variability of wind speeds at hub-height over the North Sea Horns Rev farm. Adapted from Vincent et al. (2011).

resource assessed in a target region. With coarse resolution, wind resource is underestimated by almost 50%, compared to finer grid resolution. An inaccurate wind resource map would result in a failure by policy-makers to develop a potentially lucrative region for wind harvesting. This problem is especially relevant in the current global economic climate where investments in the wind energy industry have been not been growing well. Thus, it is important for increased research and development to contribute to make wind energy even more competitive compared to traditional fossil fuel energy, and accurate wind assessment plays a role in achieving that goal.

This research project seeks to look at numerical weather modeling efforts to improve wind resource prediction. Resolution and predictive accuracy tend to go hand-in-hand in numerical simulations; in an ideal situation, where computational resource were unlimited, having an extremely fine grid would tend to provide the most accurate picture. However, this is not currently affordable, as increasing resolution would require increasing computation resource.

The range of length scales in a turbulent flow, (i.e the ratio of the largest to the smallest eddy size) is in the order of $R_t^{3/4}$, the turbulence Reynolds number raised to the power of 3/4 (Wyngaard 2010). $R_t = ul/\nu$. R_t is the turbulence Reynolds number, it is defined as the Reynolds for the larger, energy-containing eddies of size l and velocity u relative to the mean flow. This can be used as an estimate for the number of grid points necessary for an accurate simulation. In a three dimensional simulation, the number of grid points needed is proportional to the

Figure 1.2: Impact of grid resolution on wind resource mapping. Adapted from Vincent et al. (2011).

cube of the length scale, which means that the grid points needed are proportional to $R_t^{9/4}$. For high turbulent flow, the computational requirement is immense since in typical atmospheric flow, $R_t \sim 10^8$. Parameterization of physical processes and usage of numerical averaging techniques are needed to overcome these computational requirements. Hence models are used to achieve a compromise between efficiency and accuracy.

The traditional approach for wind resource modelling is based on Jackson-Hunt type linear flow models such as WAsP, MS3DJH and MS-Micro (Brower 2012). Linear flow models solve steady-state linearized Navier-Stokes equations with simplified assumptions for terrain effects (Jackson and Hunt 1975). They are computationally efficient to use and are able to provide a suitable first approximation for the wind climate of a chosen site. However, the increased availability of computational resources and power has made it more attractive to consider more complex models to be used for wind resource assessment. Beaucage et al. (2014) compared four numerical models used in wind resource estimation in a recent study. They found that coupled models perform better than linear or CFD models mainly due to the

ability of coupled models to simulate the unsteady flow aspect of real meteorological conditions and accounting for phenomena that arise from atmospheric stability and thermal effects. Other studies also showed that higher-order models are able to better resolve nonlinear effects and hence provide better approximation of the available wind resource (Mahoney et al. 2012; Badger et al. 2014; Gopalan et al. 2014).

Numerical weather prediction models (NWP) such as the Weather Research and Forecasting model (WRF) can also be applied to wind resource assessment. NWP models are typically compressible, non-hydrostatic models that are able to simulate a wider range of meteorological phenomena driven by atmospheric stability such as low-level jets (see Section 1.3.4), and barrier jets. NWP models are used as operational weather forecast models and the interest in the use of NWP for wind resource mapping is growing (Storm et al. 2009; Badger et al. 2014; Beaucage et al. 2014; Hahmann et al. 2014; Wilczak et al. 2014).

With increased focus on offshore wind power (Dvorak et al. 2010; Archer et al. 2014; Gryning et al. 2014; Hahmann et al. 2014), NWP offers another advantage for wind resource mapping. NWP provides a larger spatial and temporal coverage for wind climate studies compared to in-situ offshore measurements. Few long-term measurements of offshore wind are available due to the cost and difficulty in setting up mast towers. NWP can bridge this gap and provide better approximations of the wind climate at a potential offshore site before further commitment into costly long-term measurement campaigns.

Despite the greater accuracy of NWP models in wind prediction, the modelling outcome requires familiarity with both the geographic region of interest and the available model parameterization schemes (Storm et al. 2009; Talbot et al. 2012; Nunalee and Basu 2014; Vanderwende et al. 2015; Berg et al. 2015b). No universal model set-up performs optimally in all geographic region and mesoscale conditions. Hence it would be useful to be able to objectively quantify the skill of mesoscale simulations and pinpoint the sources of such variability.

1.2 Aims and Objectives

This thesis aims to use Large-Eddy Simulation (LES) nested within Weather Research Forecast (WRF) model to perform high-resolution modelling of coastal wind

conditions, with an emphasis on coastal low-level jet (LLJ). In order to achieve this aim, the thesis is carried out in the following parts; (1) a review of the basic principles of LES and turbulence modelling, atmospheric dynamics pertaining to LLJ, (2) quantify the sensitivity of LLJ to mesoscale configurations and selection of the best performing model configuration and (3) nested LES study of LLJ modeling, with an emphasis on improving turbulence transition from mesoscale into LES domain. The scope of this thesis is limited to one-way nested LES in a stable atmospheric boundary layer.

The thesis is organized as follows:

Chapter 1 - Literature Review

A review of the basic principles of LES modelling is carried out to provide a background understanding of the advantages and shortcomings of the method. A summary of atmospheric dynamics is presented, focusing on the lower planetary boundary layer and the mechanism of the low-level jet. A study of nesting LES within WRF is done and the current challenges of using such a method are highlighted. Lastly, an overview of the set of metrics used to diagnose model performance in this work is given.

Chapter 2 - CBLAST-LOW 2001 Case Study - Mesoscale Simulations

In Chapter 2, the CBLAST-LOW 2001 campaign is identified as a case study. Mesoscale simulations of a coastal LLJ observed during the campaign are carried out. The sensitivity of the simulated coastal LLJ to model parameters such as initial and boundary conditions (IBC), the planetary boundary layer (PBL) scheme, vertical resolution and sea-surface temperature (SST) is tested. The simulations are first evaluated against available Automated Surface Observing System (ASOS) and rawinsonde data. Then, inter-model comparisons are compared and analyzed. IBC is identified as the largest source of mesoscale variability in simulating the LLJ. There was a significant difference in frontal passage temporal representation between the IBC tested.

Chapter 3 - NEAQS-ITCT 2004 Case Study - Mesoscale Simulations

Based on the results of the Chapter 2 study, the NEAQS-ITCT 2004 campaign is identified as a better alternate case study, it featured a more extensive observational dataset for model evaluation, in particular, a month-long set of high-resolution lidar wind profiles available at 15-min intervals. A continuous 48-hour period that featured two different LLJ events, a weak LLJ ($\sim 10 \text{ m s}^{-1}$) and a strong LLJ ($\sim 18 \text{ m s}^{-1}$), was selected for analysis. A similar sensitivity study is carried out, where IBC, PBL schemes and SST forcings are tested. Based on the diagnostic metrics used, the GFS_MYNN configuration is identified as the best performing model configuration and is used as the parent domain in the subsequent nested WRF-LES simulations.

Chapter 4 - NEAQS-ITCT 2004 Case Study - WRF-LES Simulations

Having selected the best performing mesoscale simulation and characterized its variability, GFS_MYNN is used for the nested LES study in this chapter. Two sets of nested WRF-LES simulations are designed and carried out, one for each of the two LLJ events identified. A method (Cell-Perturbation Method) to generate turbulence between mesoscale and LES domain is also tested. The high-resolution LES simulation improved LLJ predictions within the subject and jet core layer compared to the parent mesoscale domain. Spectral analysis shows that the Cell-Perturbation Method (CPM) is able to reduce the fetch required for the turbulence spectra to reach equilibrium. The improvement in turbulence representation due to CPM also resulted in further improvements in the NBIAS and NRMSE scores of the simulated LLJ. CPM is thus shown to be an effective tool for nested WRF-LES studies.

In the final chapter of the thesis, a summary of the work done and the significance of using WRF-LES and CPM to advance and improve wind modelling is presented. Additionally, extensions of the current work are explored and further development studies involving the use of WRF-LES and CPM are proposed.

1.3 Literature Review

1.3.1 Large-Eddy Simulation

Large Eddy Simulation (LES) is a method formulated in the 1970s to tackle the mathematical intractability of the Navier-Stokes equations. Deardorff (1970) was one of the first to use this method to study turbulent channel flow. The LES method is built upon the theory of an equilibrium high Reynolds number energy cascade (Kolmogorov 1991). The key principle of LES is the separation of the wide range of eddy scales present in turbulent flows into large (resolved) and small (modeled) scales through the use of spatial filtering. This offers an advantage over the ensemble averaged, Reynolds-Averaged Navier-Stokes (RANS) method where all eddy scales are modeled.

Spatial Filtering

A spatial filtering operator is used in LES to separate resolved and sub-filter eddy scales, in contrast with RANS where ensemble averaging is carried out. This averaging operator, also known as a filter function in LES literature is as follows:

$$\bar{f}(\mathbf{x}, t) = \int_{-\infty}^{\infty} \int_{-\infty}^{\infty} \int_{-\infty}^{\infty} f(\mathbf{x}', t) G(\mathbf{x} - \mathbf{x}') dx'_1 dx'_2 dx'_3 \quad (1.1)$$

where G is the filter kernel, f is the original function and $\bar{f}$ is the filtered kernel. Common filter functions used in CFD are:

- Top-hat or box filter:

$$G(\mathbf{x}) = \begin{cases} \frac{1}{\Delta}; |\mathbf{x}| \leq \frac{\Delta}{2} \\ 0; \text{otherwise} \end{cases} \quad (1.2)$$

The top-hat filter averages over the space specified by Δ , the filter width. This filter is most commonly used in finite volume methods in CFD. The filter width is often chosen to be twice the grid spacing. This is convenient because the computational resolution is limited by the grid size and hence, it would not serve any purpose to select a filter width that is smaller (Versteeg and Malalasekera 2007).

- Gaussian Filter:

$$G(x) = \left(\frac{\gamma}{\pi\Delta^2}\right)^{\frac{3}{2}} \exp\left(-\frac{\gamma x^2}{\Delta^2}\right) \quad (1.3)$$

Where the typical value for $\gamma = 6$ (Versteeg and Malalasekera 2007)

The Gaussian filter does not have negative values in physical or wavenumber space. It is also smooth and infinitely differentiable.

- Wave cutoff filter:

The wave cutoff filter is defined in Fourier space, and its function is to cut-off at a specific wavenumber, k_c .

$$G(x) = \frac{\sin(k_c x)}{\pi x} \quad (1.4)$$

Or in Fourier space:

$$\hat{G}(k) = \begin{cases} 1 & ; k \leq k_c \\ 0 & ; k \geq k_c \end{cases} \quad (1.5)$$

The above function, $\hat{G}(k)$, is a low-pass filter. It will allow wavenumbers smaller than cutoff through the filter, while larger wavenumbers are rejected. This can be interpreted as allowing large eddies (small wavenumbers) while smoothing out small eddies (larger wavenumbers). This filter function is used in conjunction with spectral methods.

The different filter functions are shown in Fig. 1.3, in both physical and wave space.

Filtered Navier Stokes Equation

The governing equations need to be treated with the filter operation before these can be used in LES runs. Thus, using the incompressible Navier-Stokes equation:

$$u_{i,i} = 0 \quad (1.6)$$

$$u_{i,t} + (u_i u_j)_{,j} = -\frac{1}{\rho} p_{,i} + \nu u_{i,jj} + S_i \quad (1.7)$$

Where S_i is a source term, (e.g. gravity).

The velocity field, u_i can be expressed as the sum of resolved and sub-filter terms:

$$u_i(\mathbf{x}, t) = u_i^r(\mathbf{x}, t) + u_i^s(\mathbf{x}, t) \quad (1.8)$$

Figure 1.3: Filter functions in physical space (a) — : wave cutoff; ---: Gaussian; -.-: top-hat; wave space (b) — : top-hat cutoff; ---: Gaussian; -.-: wave cutoff. Adapted from Piomelli (1999).

Where the r term represents the part that passes through the filtering operation and can be expressed as the following:

$$u_i^r(\mathbf{x}, t) = \int_{-\infty}^{\infty} \int_{-\infty}^{\infty} \int_{-\infty}^{\infty} G(\mathbf{x} - \mathbf{x}', t) u_i(\mathbf{x}', t) dx'_1 dx'_2 dx'_3 \quad (1.9)$$

The filter function $G(x)$ is required to be:

- Commutative in both temporal and spatial differentiation

That is when (1.9) is partially differentiated in time and space should be:

$$\frac{\partial u_i^r}{\partial t} = \left(\frac{\partial u_i}{\partial t} \right)^r \quad (1.10)$$

$$\frac{\partial u_i^r}{\partial x_i} = \left(\frac{\partial u_i}{\partial x_i} \right)^r \quad (1.11)$$

- Able to remove fluctuations of scale much less than the filter cutoff scale, Δ :

In order to satisfy this requirement, the choice of Δ will have to be much smaller than l , the characteristic length of large eddies such that the filter does not smooth out their features. However, Δ has to be much larger than η , the Kolmogorov eddy length. This will ensure that the filter removes these small eddies, which are computationally intensive to calculate, especially at high turbulent Reynolds number, R_t . Hence the filter scale requirement is as such: $l \gg \Delta \gg \eta$

- Able to leave variations at scales that are greater than Δ unchanged:

This requirement ensures that the filtering operation will not change any constants that undergo filtering. This means that:

$$\begin{aligned} C^r &= C = \int_{-\infty}^{\infty} \int_{-\infty}^{\infty} \int_{-\infty}^{\infty} G(\mathbf{x}', t) C \, dx'_1 \, dx'_2 \, dx'_3 \\ &= C \int_{-\infty}^{\infty} \int_{-\infty}^{\infty} \int_{-\infty}^{\infty} G(\mathbf{x}', t) \, dx'_1 \, dx'_2 \, dx'_3 \end{aligned} \quad (1.12)$$

The above requirements thus imply that the filter function $G(\mathbf{x})$ should be:

$$\int_{-\infty}^{\infty} G(\mathbf{x}', t) \, d^3\mathbf{x}' \quad (1.13)$$

Having specified the above requirements for the filter function, it is then applied to the Navier-Stokes equations and yields the following set of equations:

$$u_{i,i}^r = 0 \quad (1.14)$$

$$u_{i,t}^r + (u_i u_j)^r_{,j} = -\frac{1}{\rho} p^r_{,i} + \nu u_{i,jj}^r + S_i^r \quad (1.15)$$

These equations can now be solved numerically. However, only the following terms can be obtained from the numerical solution: u_i^r and p^r , while the convective terms, $(u_i u_j)^r_{,j}$ are an unknown because

$$(u_i u_j)^r \neq u_i^r u_j^r \quad (1.16)$$

Hence, the following expression is used to modify the filtered equations

$$(u_i u_j)^r = u_i^r u_j^r + ((u_i u_j)^r - u_i^r u_j^r) \quad (1.17)$$

By substituting the expression into equation (1.15), the following is obtained:

$$u_{i,t}^r + (u_i^r u_j^r)_{,j} = -\frac{1}{\rho} p_{,i}^r + \nu u_{i,jj}^r - \tau_{ij,j} + S_i^r \quad (1.18)$$

$$\text{where } \tau_{ij} = (u_i u_j)^r - u_i^r u_j^r$$

Now, because of the additional terms, τ_{ij} , called sub-filter scale stresses, there are more unknowns than there are equations. This is where sub-filter scale terms modeling comes in to address this closure problem.

Turbulence

A formal definition of turbulence remains elusive despite being a phenomena that we experience in our daily life. Examples of turbulence are easily seen such as rising plumes of smoke from a chimney, when cream is mixed with our daily cup of coffee, and even in nature where we see frothy white rapids. Hard as it is to define exactly what turbulence is, turbulent flows do share common characteristics as described (Tennekes and Lumley 1972)

- Irregularity: Turbulent flows are highly irregular and chaotic. Turbulent flows were shown to be highly dependent on initial conditions and slight changes will lead to exponentially diverged solutions in time (Lorenz 1963).
- Dissipative: Turbulent flows are dissipative; the presence of turbulent eddies extract kinetic energy from the mean flow and eventually dissipates the kinetic energy as thermal energy at the smallest scales, through a mechanism known as the energy cascade.
- Three-dimensional: Turbulence is always three-dimensional and unsteady in nature
- Large Reynolds number: Turbulent flows are observed to occur at high Reynolds number. In internal flow, turbulent transition occurs at approximately $Re \geq 2300$, while in external flow, such as the atmospheric boundary layer at $Re \geq 500,000$.

Turbulence Scales

Turbulent flows contain eddies at a wide range of scales, from the large eddies that contain the majority of the kinetic energy to the smallest eddies that dissipate the kinetic energy.

Figure 1.4: Turbulent eddy scales from large to dissipative scales. Adapted from Davidson (2011).

Fig. 1.4 depicts the flow of kinetic energy from large scale eddies to dissipative scale eddies through the energy cascade process. l_0 represent energy-containing eddies, while l_1 and l_2 represent eddies in the inertial subrange and l_η represent the Kolmogorov scale eddies.

For a characteristic length of large eddies, l_0 are scaled to be the order of the characteristic flow geometry. For example, in an atmospheric boundary layer, l_0 would be scaled to the depth of the boundary layer while in internal pipe flow, the diameter would be the characteristic dimensional length. Similarly, there exists a characteristic velocity scale to describe such eddies, u_0 , which is scaled to the square root of TKE. Thus, the turbulent time-scale can be described as

$$\tau_0 = O\left(\frac{l_0}{u_0}\right) \quad (1.19)$$

This turbulent time scale, also known as eddy turnover time, provides an order of magnitude estimate for how long the eddies remain in the flow before decaying.

The smallest eddies have scales known as the Kolmogorov length, l_η and velocity, u_η . The scaling of these parameters is not the same as those of the large eddies. Kolmogorov provided a scaling of these dissipative eddies in his seminal work (Kolmogorov 1991) using dimensional analysis. He surmised that the dissipative eddies are isotropic in nature, having no spatial directional preference while large eddies are anisotropic, with geometrical features governed by the flow dimensions. He then postulated that the smallest eddies are governed by two parameters: the kinematic viscosity of the fluid, ν and the rate of kinetic energy transfer from the large eddies that is dissipated by ε . Hence, using simple dimensional analysis, it can be shown that:

$$l_\eta = \left(\frac{\nu^3}{\varepsilon}\right)^{\frac{1}{4}}, \quad v_\eta = (\nu\varepsilon)^{\frac{1}{4}}, \quad \tau_\eta = \left(\frac{\nu}{\varepsilon}\right)^{\frac{1}{2}} \quad (1.20)$$

And since the rate of energy dissipation is determined by the rate at which energy is extracted by the large eddies:

$$\varepsilon = \varepsilon_0(u_0, l_0) \sim \frac{u_0^3}{l_0} \quad (1.21)$$

And comparing l_0 with l_η , it can be shown that:

$$\frac{l_0}{l_\eta} = \frac{l_0 \varepsilon^{\frac{1}{4}}}{\nu^{\frac{3}{4}}} \sim \frac{l_0 u_0^{\frac{3}{4}}}{l_0^{\frac{1}{4}} \nu^{\frac{3}{4}}} = \left(\frac{u_0 l_0}{\nu}\right)^{\frac{3}{4}} = R_t^{\frac{3}{4}} \quad (1.22)$$

This relation shows the wide range of scales associated in a turbulent flow such as in atmospheric boundary layer where $R_t \sim 10^8$.

It is useful to study and represent these eddies in Fourier space. Let $f(x)$ be a real function defined over length L and it can be represented in Fourier space as such:

$$f(x) = \frac{a_0}{2} + \sum_{n=1}^{\infty} a_n \cos(\kappa_n x) + \sum_{n=1}^{\infty} b_n \sin(\kappa_n x) \quad (1.23)$$

Where x is a spatial coordinate and the wavenumber:

$$\kappa_n = \frac{2n\pi}{L} \quad (1.24)$$

The function $f(x)$ can be used to represent any of the relevant variables in the Navier-Stokes equations. For example, let $f(x)$ be the fluctuating velocity of the fluid flow, u'_1 and its corresponding Fourier transform be $\hat{f}(\kappa_n)$, where the wavenumber vector, $\kappa_n = (\kappa_1, \kappa_2, \kappa_3)$ is defined to be inversely proportional to eddy length scale, L . The kinetic energy of the eddy can be then expressed in terms of wavenumber vector in Fourier space:

$$\overline{u'_i u'_i} = \sum \hat{f}^2(\kappa_n) (\Delta\kappa)^3 = \iiint_{-\infty}^{\infty} \hat{f}^2(\kappa_n) d^3\kappa \quad (1.25)$$

And energy spectrum $E(\kappa)$ is defined as

$$E(\kappa) = \iint \frac{1}{2} \hat{f}^2(\kappa_n) d^2\kappa \quad (1.26)$$

Hence

$$kinetic\ energy = \frac{1}{2} \overline{u'_i u'_i} = \int_0^{\infty} E(\kappa) d\kappa \quad (1.27)$$

Hence, it is easy to see that each eddy size is prescribed a unique wave number and that the kinetic energy spectrum is the sum of all contributions from each wave number, just as the physical interpretation that the kinetic energy of the flow is the sum of all kinetic energy contained in each eddy.

Fig. 1.5 depicts the energy cascade mechanism. It should also be noted that the energy cascade spectrum is plotted in a log-log scale to highlight the -5/3 law.

Region I: This region is where the large energy containing eddies are. They extract kinetic energy from the mean flow. The interaction between vorticity and velocity gradients is the key mechanism that creates and maintain turbulence. These large eddies creates smaller eddies through vortex stretching and tilting. A mathematically rigorous treatment of the vorticity interaction can be found in Wyngaard (2010); Davidson (2011). In the filtered kinetic energy equation, the energy extraction mechanism would be represented by a negative Source term. Eddies in this region have velocity and length scales defined as v_0 and l_0 .

Region II: This region is the inertial subrange. This linear region exists in fully turbulent flow. This is also known as the transfer region because in the energy cas-

Figure 1.5: Energy spectrum of turbulent kinetic energy in wave space. P^k is the kinetic energy extracted from mean flow and ϵ is the kinetic energy dissipated at Kolmogorov scale. Adapted from Davidson (2011).

cade process, eddies in this region perform the intermediary function of transferring the kinetic energy extracted from the mean flow by the large eddies to the dissipative eddies. Using the assumption that energy transfer rate is constant between eddy sizes (wavenumbers) and whatever energy transferred is dissipated at a rate ϵ , the kinetic energy of the eddies in this range can be written as:

$$E = E(\kappa, \epsilon) \quad (1.28)$$

Using dimensional analysis, the following relationship is obtained:

$$E(\kappa) = C_K \epsilon^{\frac{2}{3}} \kappa^{-\frac{5}{3}} \quad (1.29)$$

Where C_K is the Kolmogorov constant

The relation derived above describes the Kolmogorov spectrum law or -5/3 law. Energy spectral analysis of turbulent flow simulation should exhibit this decay relation in this region.

Region III: This region is the dissipation region. Eddies in this region are small and isotropic. These eddies dissipate the kinetic energy extracted from the mean flow and transferred by the inertial subrange as thermal energy. In the TKE energy equation, this viscous dissipation is represented by the ε term. Eddies in this region are in the Kolmogorov scales, v_η and l_η .

Conservation equations for SFS quantities

Now, using the filtered Navier-Stokes equations, we can derive the conservation equations for the SFS fluxes. This will give us a mathematical model to describe the changes of the SFS quantities in time and space. Using the formulation derived by Lilly (1967) and starting with equation (1.18), while assuming that Reynolds number of the filtered scale motion is high enough to neglect molecular diffusion leads to

$$u^r_{i,t} + (u^r_i u^r_j)_{,j} = - \left(\frac{1}{\rho} p^r + \frac{2}{3} e \right)_{,i} - \tau^d_{ij,j} \quad (1.30)$$

$$\text{Where } \tau^d_{ij} = (u_i u_j)^r - u^r_i u^r_j - \frac{2}{3} e \delta_{ij} ; \quad e = \frac{1}{2} \tau_{ii} = \frac{1}{2} ((u_i u_i)^r - u^r_i u^r_i)$$

The term τ^d_{ij} is known as the deviatoric stress tensor, which is the difference between the SFS stress tensor τ_{ij} and one-third of the trace of the tensor. The term e , is the SFS turbulent kinetic energy. The deviatoric tensor is most often used in the modeling of SFS stress.

Using the following relation:

$$\frac{\partial}{\partial t} [(u_i u_j)^r - u^r_i u^r_j] = \left(u_i \frac{\partial u_j}{\partial t} + u_j \frac{\partial u_i}{\partial t} \right)^r - u^r_i \frac{\partial u^r_j}{\partial t} - u^r_j \frac{\partial u^r_i}{\partial t} \quad (1.31)$$

and using the conservation equation for u_i , the conservation equation for τ^d_{ij} is defined as

$$\begin{aligned}
\frac{\partial \tau_{ij}^d}{\partial t} + u_k^r \frac{\partial \tau_{ij}^d}{\partial x_k} = & \frac{\partial}{\partial x_k} [(u_i u_j u_k)^r - u_i^r (u_j u_k)^r - u_j^r (u_i u_k)^r - u_k^r (u_i u_j)^r + 2u_i^r u_j^r u_k^r \\
& - \frac{\delta_{ij}}{3} ((u_l^2 u_k)^r - 2u_l^r (u_l u_k)^r - u_k^r (u_l^2)^r + 2(u_l^r)^2 u_k^r)] \\
& + \frac{2}{3} e S_{ij}^r - [\tau_{ik}^d \frac{\partial u_j^r}{\partial x_k} + \tau_{jk}^d \frac{\partial u_i^r}{\partial x_k} - \frac{1}{3} \delta_{ij} \tau_{kl}^d S_{kl}^r] \\
& - [\frac{p}{\rho} S_{ij}]^r + \frac{p^r}{\rho} S_{ij}^r \\
& + \frac{1}{\rho} \frac{\partial}{\partial x_k} (\delta_{ik} [(u_j p)^r - u_j^r p^r] + \delta_{jk} [(u_i p)^r - u_i^r p^r] - \frac{2}{3} \delta_{ij} [(u_k p)^r - u_k^r p^r]) \quad (1.32)
\end{aligned}$$

The two terms on the left-hand side of (1.32) are the change in time and the advection by the local resolved velocity field respectively. The first two lines on the right hand side represent a set of divergence transport terms. There are two terms on the third line which describe the interaction between τ_{ij} and the resolved strain rate tensor, S_{ij}^r . The first of this set refers to the gradient production of τ_{ij}^d that are aligned with S_{ij}^r while the second of this set refers to the tilting production of τ_{ij}^d that is reoriented by the strain rate interaction. The two terms on the fourth line are destruction by pressure, while the last set of terms is the pressure transport.

The conservation equation for e , the SFS turbulent kinetic energy is derived as:

$$\begin{aligned}
\frac{\partial e}{\partial t} + u_k^r \frac{\partial e}{\partial x_k} = & \frac{\tau_{ij}^d}{2} S_{ij}^r \\
& - \frac{\partial}{\partial x_k} \left(\frac{(u_k u_i u_i)^r}{2} - \frac{u_k^r (u_i u_i)^r}{2} - u_i^r (u_k u_i)^r + u_i^r u_i^r u_k^r \right. \\
& \left. + \frac{(u_k p)^r}{\rho} - \frac{u_k^r p^r}{\rho} \right) - \varepsilon \quad (1.33)
\end{aligned}$$

The terms on the left hand side are the change in time and advection of e . On the right hand side, are the shear production, divergence of fluxes and rate of viscous dissipation terms.

Due to the mathematical intractability of evaluating the turbulence moments that involve even higher order of unknowns, researchers have proposed several methods to approximate and model these SFS fluxes and the following models represent the commonly used in both engineering and research applications of LES.

SFS Models

To analyze the SFS stress terms, the decomposition introduced in the previous section is used, which yields:

$$\begin{aligned}\tau_{ij} &= (u_i u_j)^r - u_i^r u_j^r \\ &= \underbrace{\left((u_i^r u_j^r)^r - u_i^r u_j^r\right)}_{(I)} + \underbrace{\left((u_i^r u_j^s)^r + (u_i^s u_j^r)^r\right)}_{(II)} + \underbrace{(u_i^s u_j^s)^r}_{(III)}\end{aligned}\tag{1.34}$$

The three sets of terms are commonly referred to as:

- Term (I), Leonard stresses: These terms can be resolved directly from the resolved velocity field. They arise from the eddy interactions in the resolved scale.
- Term (II), Cross stresses: These terms represent the eddy interaction between the resolved and sub-filter scales. It is also related to the interscale energy transfer between the large and small eddies.
- Term (III), SFS Reynolds stresses: These represent the sub-filter interactions between the small eddies.

Although these terms possess different characteristics and are not necessary similar in nature, they are commonly lumped together and modeled as a whole rather than as separate entities, especially in finite volume methods (Peyret and Krause 2000)

Smagorinsky-Lilly Model

The Smagorinsky-Lilly model was one of the first models developed to model the SFS stresses. Lilly (1967) used the simplest first order closure assumption of (1.32) for the proposed SFS model. It assumes a quasi-steady locally homogenous state where rate of production is equal to dissipation of τ_{ij}^d (Wyngaard 2010). This can be expressed as:

$$\tau_{ij}^d = \nu_{SFS} S_{ij}^r\tag{1.35}$$

where $S_{ij}^r = \frac{1}{2} \left(\frac{\partial u_i^r}{\partial x_j} + \frac{\partial u_j^r}{\partial x_i} \right)$ and ν_{SFS} is the eddy viscosity.

He then proposed to use Smagorinsky model (Smagorinsky 1963) for the eddy viscosity:

$$\nu_{SFS} = (C_s \Delta)^2 |S^r| \quad (1.36)$$

where Δ is often taken to be filter cutoff scale and C_s is a constant.

The dissipation of turbulent kinetic energy is given simply as

$$\varepsilon = \frac{1}{2} \tau_{ij}^d S_{ij}^r \quad (1.37)$$

Using the assumption that the spectrum at the cutoff length scale lie in the inertial sub-range, Lilly, using theoretical analysis of the decay of isotropic eddies, showed that C_s is related to the Kolmogorov constant, $\alpha \approx 1.5$

$$C_s \approx 0.23 \alpha^{-\frac{3}{4}} \approx 0.17 \quad (1.38)$$

However, the constant C_s has been shown to vary across a range of values from 0.17 to 0.24 (Versteeg and Malalasekera 2007) depending on various factors such as grid resolution, geometric complexity, filter functions etc. In one of the first applications of this model (Deardorff 1970) in LES simulation of turbulent channel flow, $C_s = 0.1$ was found to be the suitable value for such cases. This indicates that in regions with large shear strains, the eddy viscosity model is too dissipative and hence the constant C_s needs to be adjusted downwards to account for this. Herein lies the weakness with this constant coefficient model, it has to be adjusted for different case simulations, often from empirical evidence. Despite this, this model has been implemented for common engineering applications due to its computation simplicity and relative success in turbulence predictions.

One-Equation Eddy-viscosity Model

For the Smagorinsky-Lilly model, the underlying assumption of equilibrium for turbulence production and dissipation that happens instantaneously is not strictly valid especially for cases that involve wall-dominated internal flows, flow separation etc. To address this, we can solve additional transport conservation equations for each transported turbulence characteristics such as $(u_i u_j)^r$. However, with the additional

of multiple transport equations would be computational expensive. A simplified approach would to use a single transport for one representative turbulence characteristic.

Most commonly used one-equation model is based on the eddy viscosity concept, where the eddy viscosity is now dependent on this transported characteristic instead of strain rate. One such characteristic used in modeling is the SFS turbulent kinetic energy, e defined as

$$e = \frac{1}{2} ((u_i u_i)^r - u_i^r u_i^r) \quad (1.39)$$

Using a simplified form of the transport equation for e

$$\frac{\partial e}{\partial t} + u_i^r \frac{\partial e}{\partial x_i} = -\tau_{ij}^d S_{ij}^r + \frac{\partial}{\partial x_i} \left(\nu_{SFS} \frac{\partial e}{\partial x_i} \right) - \varepsilon \quad (1.40)$$

Where:

$$\begin{aligned} \nu_{SFS} &= C_k \Delta \sqrt{e} \\ \varepsilon &= C_\varepsilon \frac{e^{\frac{3}{2}}}{\Delta} \end{aligned}$$

The constants C_k and C_ε are non-dimensional coefficients similar to C_k that are obtained empirically. It can be noted that this closure method reduces to the Smagorinsky-Lilly model when $\tau_{ij}^d S_{ij} + \varepsilon = 0$, which describes the equilibrium assumption used. The strain rate tensor, τ_{ij}^d is modeled the same way as the Smagorinsky-Lilly model, ie $\tau_{ij}^d = \nu_{SFS} S_{ij}^r$. This closure method is also referred to as the 1.5-order closure or TKE closure model because it is between a first-order and second-order closure method as the turbulent kinetic energy (TKE) is the only second order moment that is explicitly evaluated through a transport equation.

The advantage of this one-equation eddy viscosity model over the standard Smagorinsky-Lilly is that it takes into account the time evolution of the TKE and responds better to changing local flow conditions, thereby providing a more accurate description of the flow field, where turbulence is generated and destroyed at different places. This model has been shown to perform well (Fureby et al. 1997) compared to the standard Smagorinsky-Lilly model.

Dynamic SFS model

Germano et al. (1991) first proposed this procedure to improve some of the deficiencies with the above models where the constant C_s or C_k and C_ε is predetermined and remains constant throughout the whole flow domain. With this procedure, the constants will be computed locally instead. This is accomplished by carrying out an additional filtering operation at a filter width larger than the grid cut-off and this coarser filter is known as the test filter.

Recall that the SFS stress that has to be modeled is

$$\tau_{ij} = (u_i u_j)^r - u_i^r u_j^r \quad (1.41)$$

Let us define the test filter with cut-off width $\tilde{\Delta}$ (typically defined as $\tilde{\Delta} = 2\Delta$, where Δ is the grid cut-off width). Applying the second filtering operation to the filtered Navier-Stokes Equation will yield the following SFS stress term that has to be modeled

$$T_{ij} = (\widetilde{u_i u_j})^r - \widetilde{u_i^r u_j^r} \quad (1.42)$$

Recall the definition of Leonard Stresses in (1.34) and it can be expressed as:

$$L_{ij} = \widetilde{u_i^r u_j^r} - \widetilde{u_i}^r \widetilde{u_j}^r \quad (1.43)$$

The Leonard stress terms can also be computed from this relation:

$$L_{ij} = T_{ij} - \widetilde{\tau_{ij}} \quad (1.44)$$

We then use the Smagorinsky closure and assuming that both T_{ij} and τ_{ij} can be related by a scaling law, we can derive the following relation:

$$\tau_{ij} - \frac{1}{3}\delta_{ij}\tau_{kk} = -2C'_s\Delta^2 |S^r| S_{ij}^r \quad (1.45)$$

$$T_{ij} - \frac{1}{3}\delta_{ij}T_{kk} = -2C'_s\widetilde{\Delta^2 |S^r|} \widetilde{S_{ij}^r} \quad (1.46)$$

$$L_{ij} - \frac{1}{3}\delta_{ij}L_{kk} = -2C'_s \left(\widetilde{\Delta^2 |S^r|} \widetilde{S_{ij}^r} - \Delta^2 |S^r| S_{ij}^r \right) \quad (1.47)$$

Since (1.47) is a symmetric tensor equation with no trace, we have five equations to evaluate C'_s which make this an over-determined system. Lilly suggested utilizing

the least-squares approach (Lilly D. , 1992) to compute the constant. The error is defined as the square of the difference between the left-hand side and the right-hand side of equation (1.47)

$$Q = \left(L_{ij} - \frac{1}{3} \delta_{ij} L_{kk} + 2C'_s M_{ij} \right)^2 \quad (1.48)$$

Where $M_{ij} = \left(\widetilde{\Delta^2 |S^r|} \widetilde{S_{ij}^r} - \Delta^2 |S^r| S_{ij}^r \right)$

The error, Q, will then be a minimum when $\frac{\partial Q}{\partial C'_s} = 0$. The derivative of Q is then given as

$$\frac{\partial Q}{\partial C'_s} = 4M_{ij} \left(L_{ij} - \frac{1}{3} \delta_{ij} L_{kk} + 2C'_s M_{ij} \right) = 0 \quad (1.49)$$

Simplifying the above expression will yield

$$C'_s = - \frac{\langle L_{ij} M_{ij} \rangle}{2 \langle M_{ij} M_{ij} \rangle} \quad (1.50)$$

The $\langle \rangle$ brackets indicate an averaging.

Germano et al. (1991) noted that the dynamic SFS model resulted in C'_s values that varied sharply throughout the domain, including negative values of eddy viscosity, which led to numerical instabilities. Hence, ensemble averaging in homogeneous direction(s) was employed by Germano to overcome this issue. Also, C'_s is constrained to be positive such that eddy-viscosity values remain positive ($\nu_{SFS} \geq 0$).

It should also be noted that the averaging over homogeneous directions is restrictive because not all real 3D flows have such homogeneity, resulting in mathematical inconsistencies. Ghosal et al. (1995) used an integral formulation over the entire domain to evaluate C'_s instead and was thus able to apply this dynamic procedure to general inhomogeneous flows.

This dynamic procedure can be also generalized for use with higher order closure methods such as the one-equation model. This can be achieved by deriving similar forms of equation (1.50) for the coefficients such as C_k and C_ε .

Should there be no constraints imposed on the solution for C'_s , there will result in locally negative eddy-viscosity as noted before. This negative dissipation has been interpreted as a form of backscatter, where energy from SFS fluxes transfer back towards the resolved scale, a phenomenal that has been observed in experimental

studies, which could be advantageous. However, this transfer of energy can lead to unrealistic solutions of the flow field and violation of energy conservation.

Scale-similarity and Linear Combination Model

The models discussed before, based on eddy-viscosity, model the SFS stress as a whole. However, it is shown in equation (1.34) that SFS stress can be further split into different stress terms arising from different interactions. The central concept in scale-similarity model is that the primary interaction between resolved scale and SFS is between the smallest resolved eddies ($l = \Delta^+$) and the biggest SFS eddies ($l = \Delta^-$) and they are similar in their interaction.

Recalling that the SFS stress can be split as

$$\tau_{ij} = L_{ij} + C_{ij} + R_{ij} \quad (1.51)$$

Where L_{ij} is the Leonard stresses, C_{ij} is the Cross stresses and R_{ij} is the SFS Reynolds stresses.

The term of interest is C_{ij} as this term represents the interaction between resolved eddies and SFS eddies. Bardina et al. (1980) proposed that while L_{ij} can be computed explicitly, the Cross stresses and Reynolds stresses be modeled together as such

$$C_{ij} = C_c \left(u_i^r u_j^r + \tilde{u}_i^r \tilde{u}_j^r \right) \quad (1.52)$$

$$R_{ij} = 0 \quad (1.53)$$

It was found that this model was not dissipative, which required a combination approach, where R_{ij} will be modeled using the Smagorinsky approach and C_{ij} will still be modeled as equation (1.52).

$$C_{ij} = C_c \left(u_i^r u_j^r + \tilde{u}_i^r \tilde{u}_j^r \right) \quad (1.54)$$

$$R_{ij} = (C_s \Delta)^2 |S^r| S_{ij}^r \quad (1.55)$$

An advantage of the scale-similarity model is that the energy transfer from SFS to resolved grid is modeled while avoiding the numerical instabilities in Dynamic

models where eddy-viscosity is allowed to become negative. This has also shown to yield more accurate simulation results but at the expense of greater computational cost (Fureby et al. 1997). In this study, the One-Equation Eddy-viscosity model is chosen as the turbulence closure model.

1.3.2 Planetary Boundary Layer

The concept of boundary layer stems from early aerodynamic studies of Ludwig Prandtl. This revolutionary idea that at the interface between a solid boundary and moving fluid exists a buffer layer where fluid velocity slows down, led to great advancement in the understanding of fluid dynamics. Likewise, in our atmosphere (a moving fluid) there is a boundary layer at the ground surface of Earth.

Figure 1.6: Acoustic sounder record from a convective PBL. PBL is at 200m at 1000 till 1100, which then grows to 400m at 1230 before falling back to 200m at 1330. It grows back to 400m at 1400. Adapted from Wyngaard (2010).

The existence of the planetary boundary layer (PBL) was first made visible in the early 1970s with use of acoustic sounding technique, pioneered by McAllister et al. (1969) and Little (1969). In Fig. 1.6, the acoustic sounding shows the inversion layer, the distinct the dark line that separates the lower and upper atmosphere.

Understanding the mechanics that govern the behavior of the PBL is of interest to researchers because of several reasons such as

- Anthropological interactions with the atmosphere are mainly in this region

- Local weather forecasts
- Aerosol dispersion

Characteristics of PBL

Figure 1.7: (Left) Typical potential temperature, horizontal wind velocity and humidity profiles over land in mid-latitudes during cloudless conditions. (Right) Typical schematic diagram of eddy structure of PBL during day and night. Adapted from Kaimal and Finnigan (1994).

The PBL in the day is made of three separate regions:

- **Surface Layer:** this layer is typically about 10% of the height of the PBL. The temperature, velocity and humidity vertical gradient is the largest in this region. The turbulence moments are strongly influenced by the surface
- **Well-mixed Layer:** in this layer, the flow quantities tend to be constant throughout the layer. This is because of the large eddy structures or convective plumes that span this layer which promotes mixing
- **Inversion:** Inversion is defined as a layer of temperature increase with height.

The presence of this inversion inhibits mixing and thus preventing any interaction between the PBL and the free troposphere. This layer also marks the upper limit of the PBL. The top of PBL can also be determined by change in wind speed/direction and air mass variables such as humidity.

In the night, the PBL structure is different from that in the day, due to the changes in surface heat flux. The nocturnal PBL is made up of these regions:

- **Stable Nocturnal Boundary Layer:** Due to cooling in the night, there is a marked decrease in convective effects, resulting in a more stable boundary layer. This layer is much more shallow compared to the daytime PBL. There is also a nocturnal low level jet that forms, characterized by the a maximum horizontal wind speed higher than geostrophic wind.
- **Residual Layer:** This layer is what remains of the daytime mixed layer, where temperature and humidity quantities remain approximately constant.
- **Weak Inversion:** The inversion layer in the nocturnal PBL is not as sharply defined, compared to the daytime inversion layer. And thus, it is sometimes difficult to prescribe an exact height of the nocturnal PBL. This surface inversion is due to a stable surface layer caused by radiative cooling at the surface at night, without radiative forcing. The net downward sensible heat-flux causes the stratification.

1.3.3 Surface Layer

Surface Energy Balance

The diurnal variation in the PBL structure is due to stability effects related to the surface energy balance (Wynngaard 2010). Using an assumption of an infinitely thin slab for the earth's surface, the energy balance relation can be written as

$$\frac{\partial I}{\partial t} = C_t + R_n + C_b \quad (1.56)$$

where

I = internal energy of the slab

C_t = net rate of energy input by conduction through air

R_n = net rate of energy input by radiation

C_b = net rate of energy input by conduction through soil

Internal energy, I is a measure of the surface temperature. Coupled with wind speed and near-surface air temperature, this would determine the rate of surface heat transfer, C_t . If the surface is warmer than the air, C_t would have an upward direction resulting in instability of the near-surface air. In contrast, if the surface is cooler than the air, C_t would have a downward direction resulting in stable stratification of the near-surface air. However, in general, there is a need to consider the effect of moisture, thus a better quantity to describe the stability of the air would be the virtual temperature. The strong influence of the surface energy balance on the atmosphere cannot be overstated however detailed discussion of this topic is beyond the scope of this report and readers are referred to Wyngaard (2010) and Barry and Chorley (2009) for further exploration on this topic.

Our understanding of the surface layer is by the far the most developed when compared to the mechanics that drive the other layers of the PBL, despite the complex interaction between the surface and the turbulent flow. This is due to its accessibility (lowest 10% of PBL) that allows large number of observational studies to be carried out. Much of our theoretical understanding is founded upon the Monin-Obukhov (M-O) similarity theory.

Monin-Obukhov Similarity Theory

The M-O similarity theory comes from the seminal work of Obukhov (1971) and Monin and Obukhov (1954). Its derivation is based on the Buckingham- π principle. It hypothesize that the turbulence structure near the surface is governed by five parameters that are:

1. Length scale, l , taken as distance z from the surface; this assumes that the outer-part of the boundary layer does not influence the surface layer thus $l \sim z$.

2. Velocity scale, u , taken as friction velocity u_* , the square root of the magnitude of the mean kinematic surface stress, $\sqrt{\tau_s/\rho}$. This is based on the assumption that just above the surface the mean shear stress is transported by turbulence only.
3. Mean surface flux of a conserved scalar constituent, C_0 . Mean surface flux C_0 is also similarly scaled by u_* .
4. Mean temperature flux, Q_0 . Q_0 is included because it reflects the rate of buoyancy production of turbulent kinetic energy.
5. Buoyancy parameter, g/θ . This parameter is a component of the equations of motion for the PBL; it is a ratio of inertia and buoyancy forces which provides a sense of the stability of the atmospheric condition.

Now, there are six parameters comprising of the five governing M-O parameter and an unknown dependent variable, ϕ that is related to stability of the surface layer ($m=6$). There four dimensions – length, time, temperature, mass ($n=4$). Thus $m - n = 2$ independent dimensionless quantities that are functionally related. Following from Buckingham- π analysis, the first dimensionless quantity is

$$\xi = \frac{z}{L} \quad (1.57)$$

Where $L = -\frac{u_*^3 \theta_0}{\kappa g Q_0}$, Obukhov length.

And then the other dependent variable, ϕ , nondimensionalized with z , u_* and Q_0 , is function of this dimensionless parameter, ξ ,

$$\phi_m = f(\xi) \quad (1.58)$$

L is negative in an unstable PBL when ($Q_0 > 0$), positive in stable conditions when ($Q_0 < 0$) and infinite when ($Q_0 = 0$). Near-neutral conditions are when $|\frac{z}{L}| \approx 0$ which occurs near surface as $z \ll |L|$.

The similarity theory implies that the surface layer wind and temperature profiles are related as such

$$\frac{\kappa z}{u_*} \frac{\partial U}{\partial z} = \phi_m(\xi) \quad (1.59)$$

$$-\frac{\kappa z u_*}{Q_0} \frac{\partial \Theta}{\partial z} = \phi_h(\xi) \quad (1.60)$$

where κ is the von Karman constant.

The functions ϕ_m and ϕ_h need to be determined experimentally but should be universal in nature.

Physical interpretation of Obukhov length, L

From the turbulent kinetic energy equation,

$$\text{Shear production rate} = -\overline{uw} \frac{\partial U}{\partial z} \quad (1.61)$$

$$\text{Buoyant production rate} = \frac{g}{\theta_0} \overline{\theta w} \approx \frac{g}{\theta_0} Q_0 \quad (1.62)$$

Buoyant production rate of turbulent kinetic energy is independent of height, z and it is the shear production that varies with the boundary layer height. So, if we define a height, z_e at which shear production rate is equal to the buoyant production rate, in neutral conditions, we obtain

$$u_*^2 \left(\frac{u_*}{\kappa z_e} \right) = \frac{g}{\theta_0} Q_0 \quad (1.63)$$

$$z_e = \frac{u_*^3 \theta_0}{\kappa g Q_0} = |L| \quad (1.64)$$

The Obukhov length is then recovered. Thus, the Obukhov length can be described as the height at which shear and buoyant production are in equilibrium. At heights above it, buoyancy becomes dynamically important while below it, shear effects are more significant.

Experimental Validation of M-O Similarity

The Kansas experiment carried out by Businger et al. (1971) was one of the first major experiments that tested the validity of the M-O Similarity theory. The mean wind and temperature profiles as well as the turbulence fluxes for the surface layer were measured over a range of stability conditions. This can be seen in Fig. 1.8.

Figure 1.8: The solid line represents the similarity theory prediction. Mean wind shear (left) and mean potential temperature gradient (right) Adapted from Businger et al. (1971).

There have been numerous forms proposed for the stability functions, ϕ_m and ϕ_h over the years and are based on fitting the functions to experimental data set. One of the forms proposed by Högström (1988) are as follows

$$\begin{aligned} \text{Stable conditions : } \phi_m &= 1 + 4.8\xi, & \phi_h &= 1 + 7.8\xi \\ \text{Unstable conditions : } \phi_m &= (1 - 19.3\xi)^{-\frac{1}{4}}, & \phi_h &= (1 - 12\xi)^{-\frac{1}{2}} \end{aligned}$$

Using these functions, the wind and temperature profiles can be then be computed through numerical integration methods.

1.3.4 Low-Level Jet

Low-level jets, as shown in the Night profile of Fig. 1.7, are of interest to both the atmospheric community as well as the wind industry. LLJs have been frequently documented in many regions of the world (Bonner 1968; Doyle and Warner 1993;

Whiteman et al. 1997; Banta et al. 2002; Song et al. 2005; Lundquist and Mirocha 2008; Colle and Novak 2010; Rife et al. 2010; Hu et al. 2013b; Berg et al. 2015b; Du et al. 2015; Vanderwende et al. 2015). LLJs have an impact on the transportation of momentum, moisture and pollutants in regions where they occur (Angevine et al. 2006). LLJs can be a source of strong winds for wind power production but the associated strong wind shear and veer have significant influences on the stresses experienced by a wind turbine (Kelley et al. 2004). A better understanding of LLJ interaction with wind turbines can advance optimal wind turbine design strategies.

The LLJ which occurs over the Great Plains of the United States is one the most well-studied LLJ phenomena. Two main theories have been developed over the past 60 years to explain the formation of this LLJ. These mechanisms are not however just limited to LLJs in this geographical region.

The first mechanism of LLJ formation (see Blackadar (1957)) was based on inertial oscillation of the ageostrophic wind triggered by diurnal variation of eddy viscosity in the boundary layer. At sunset, radiative cooling of the surface rapidly stabilizes the boundary layer. The decay in surface-based turbulence causes the ageostrophic wind to accelerate at the top of this stable boundary layer. Van de Wiel et al. (2010) extended this mechanism to also include frictional effects within the boundary layer and the possibility that the ageostrophic winds oscillate about an equilibrium nocturnal wind rather than the geostrophic wind.

Holton (1967) proposed an alternate mechanism that relates the effect of diurnal thermal forcing has on the oscillation of boundary layer winds above sloping terrain. Baroclinicity arise from the sloping terrain and the horizontal temperature contrast between high and low terrain. Thermal variation due to the diurnal cycle yields thermal wind reversals that enhance or degrade the geostrophic winds at the top of the boundary layer.

However, both theories are not able or sufficient on their own to fully explain the peak jet speed and the timing for the formation, evolution and decay of the LLJs observed. Bonner and Paegle (1970) and more recently, Du and Rotunno (2014) combined both forcing mechanism into a model and showed that both mechanisms are important for LLJ formation.

1.3.5 Nesting LES within WRF

There is a keen interest in nesting Large-Eddy Simulations (LES) within mesoscale models to improve weather predictions (Talbot et al. 2012; Zhou and Chow 2014) as well as for wind energy applications (Liu et al. 2011; Mirocha et al. 2014a; Aitken et al. 2014; Mirocha et al. 2015). The performance of nested LES largely depends on the ability of the mesoscale domain simulation to capture real atmospheric conditions.

The Weather Research and Forecasting (WRF) model is a non-hydrostatic mesoscale numerical weather prediction model used extensively for both operational and research purposes (Skamarock et al. 2008). WRF allows for either idealized or real-data simulation studies. WRF also allows for nested domain runs that can be one-way or two-way nested.

There are two forms of nesting procedure that can be used: 1-way or 2-way nesting. In 1-way nesting, the nested inner domain receives initial and boundary conditions from the parent outer domain and does not relay any information outside, hence changes within the inner domain does not affect the outer but the reverse is true. In 2-way nesting, both nested inner and parent outer domain interact with each other, hence changes in either domain will affect the other.

WRF has been shown to simulate LLJs reasonably well for the US Great Plains (Storm et al. 2009; Vanderwende et al. 2015) and for coastal cases (Nunalee and Basu 2014). These studies have shown that the timing of LLJ formation, height and strength of the jet core are dependent on the WRF model configurations. In this study, we seek to characterize the model variability related to the model configurations and to determine sensitivity of the modelled LLJ to each.

In a study of nesting of LES in WRF, there are several issues to consider. Wyngaard (2004) discussed problems associated with modelling turbulent scales of motions that are comparable to the grid resolution and referred to this region as “terra incognita”. This is relevant when choosing the appropriate turbulence closure for the appropriate resolution. Moeng et al. (2007) also highlighted another difficulty in nesting LES within mesoscale models, which is the prediction of turbulence transition between the domains. Mesoscale models use PBL parameterization schemes that give ensemble mean turbulence which is far less accurate than 3D LES simulation. Hence there is a need to design a method that allows the transition from

parameterization in the mesoscale mother domain to the LES daughter domain. While the flow in the mesoscale domain is a turbulent flow, there is no exchange of turbulence information between mesoscale and LES domains in the current formulation of WRF model, thus requiring a spin-up process to introduce turbulence within the LES domain.

1.3.6 Diagnostic Suite

To quantify the skill of model configurations in simulating the LLJ, it is imperative to select a suite of diagnostic metrics that is robust and has the ability to elucidate the underlying biases and errors. We selected the diagnostic suite of Koh et al. (2012) which was designed to evaluate NWP performance. A brief description of the metrics used in our study will be given in this section.

In this diagnostic framework, root-mean-squared-error (RMSE) is normalized by the sum of variances for both observation, σ_O^2 and forecast, σ_F^2 . The normalized RMSE (NRMSE) can be expressed as

$$\delta^2 = \sigma^2(1 + |\mu|^2), \quad (1.65)$$

where

$$\delta \stackrel{def}{=} \frac{\sqrt{\langle \mathbf{D}^2 \rangle}}{\sqrt{\sigma_O^2 + \sigma_F^2}} \quad (1.66)$$

$$\sigma \stackrel{def}{=} \frac{\sigma_D}{\sqrt{\sigma_O^2 + \sigma_F^2}} = \sqrt{\alpha}, \quad \sigma \leq \sqrt{2} \quad (1.67)$$

$$\mu \stackrel{def}{=} \frac{\langle \mathbf{D} \rangle}{\sigma_D}. \quad (1.68)$$

$\langle \mathbf{D} \rangle$ is defined as the bias between forecast and observation and is valid for both vector and scalar variables, while σ_D^2 is the variance of the bias. The above relation for NRMSE decomposes total error into two main components - bias and variance. The variance error is also referred to as pattern error. Normalized bias (NBIAS), μ , is the bias, D , normalized with the standard deviation of the bias. It measures the statistical significance of the bias in the dataset. A larger NBIAS implies a larger statistical bias. Also, arising from the relationship between NBIAS and NRMSE, it can be shown that a NBIAS of less than 0.3 would contribute to less than 5% of the NRMSE, suggesting that the total error is dominated by pattern

error rather than bias error. Normalized Pattern Error (NPE), σ , scores the variance error between forecast and observation. It can be expressed as

$$\sigma = \sqrt{\alpha}, \quad (1.69)$$

where

$$\alpha \equiv 1 - \rho\eta; \quad (1.70)$$

$$\alpha \stackrel{\text{def}}{=} \frac{\sigma_D^2}{\sigma_O^2 + \sigma_F^2}, \quad 0 \leq \alpha \leq 2; \quad (1.71)$$

$$\rho \stackrel{\text{def}}{=} \frac{\text{tr}[\text{cov}(\mathbf{F}, \mathbf{O})]}{\sigma_O \sigma_F}, \quad -1 \leq \rho \leq 1; \quad (1.72)$$

$$\eta \stackrel{\text{def}}{=} \frac{\sigma_O \sigma_F}{\frac{1}{2}(\sigma_O^2 + \sigma_F^2)}, \quad 0 \leq \eta \leq 1. \quad (1.73)$$

NPE can be decomposed further into two other metrics, ρ and η , which assess the contribution of phase and amplitude errors, respectively. Normalized error variance, α , does not depend on variability in the observation dataset, which is useful for evaluating forecast errors and is bound between 0 and 2.

The correlation coefficient, ρ , is the familiar Pearson correlation coefficient. For ρ equal to 1, the variances are perfectly correlated while ρ equal -1 means that the variances are perfectly anti-correlated and ρ equal 0 means that there is no correlation.

Variance similarity, η , scores how similar the standard deviation of forecast and observations are. The invariant property of the metric penalizes equally the under- or over-prediction of variability. For good variance prediction, η equal to 1 means that both σ_O and σ_F match exactly. However, for poor prediction, η equal to 0 means that σ_O and σ_F match poorly.

For an ideal forecast, α equal to 0, the amplitude and phase of forecast variability are predicted perfectly hence there is no pattern error. For α equal to 1, this implies that the model performs as well as a random forecast. Lastly, for α equal to 2, this denotes maximum pattern error between forecast and observation, when η equal to 1 and ρ equal to -1. Although the amplitude of variability is well predicted in such a situation, observations and forecasts fluctuate completely out-of-phase, leading to a negative correlation. The goal of a forecast would thus be to minimize

Table 1.1: Summary of acronyms, symbols and their meanings for diagnostic suite variables

Acronym/Symbol	Meaning
RMSE	Root-mean-square-error
NRMSE	Normalized root-mean-square error
NBIAS	Normalized bias
NPE	Normalized pattern error
O	Observed variable
F	Forecasted variable
D	Difference between forecast from observed variable
σ_O	Standard deviation of observed variable
σ_F	Standard deviation of forecasted variable
σ_D	Standard deviation of difference
δ	Normalized root-mean-square error, NRMSE
μ	Normalized bias, NBIAS
σ	Normalized pattern error, NPE
α	Normalized error variance, ALPHA
ρ	Correlation
η	Variance similarity

the pattern error, α . This goal can be achieved in various ways. For any non-zero score of η , improving the phase error between forecast and observation, hence increasing correlation, ρ , will minimize α . For a fixed positive ρ score, maximizing η by matching the amplitude of forecast variability to observations will lead to minimizing α . Conversely, for a negative ρ score, η will need to be minimized to minimize α . In a scenario when either forecast or observation is nearly constant and the forecast is varying in opposite direction with observation, minimizing η (maximizing the amplitude error) would result in least overall disagreement between forecast and observation. Also, α being decomposed into amplitude and phase errors allows model evaluators to distinguish between models, with the same α scores. Model evaluators will then be able to decide if small phase or amplitude errors are better predictors of their models' performance.

Chapter 2

Variability in Mesoscale Simulation of Coastal Low-Level Jet - A CBLAST-LOW 2001 Case Study

2.1 Introduction

The WRF model is utilized to simulate a coastal LLJ observed in August 2001 during the Coupled Boundary Layers and Air-Sea Transfer Experiment in Low Winds (CBLAST-LOW) field campaign. CBLAST-LOW was conducted in the New England coastal region, which is also an area identified for offshore wind development such as the Cape Wind project (Cape Wind cited 2015). I address the following questions in this chapter:

1. How sensitive are the timing of LLJ formation, height of the LLJ core and the strength of the LLJ core to the model configuration options such as choice of initial/boundary conditions, planetary boundary layer schemes and vertical resolution?
2. How can I quantify the influence that the model parameters have on the simulated jet?
3. How sensitive are estimates of annual energy production to changes in simu-

lated LLJs due to model parameters?

In Section 2, I describe the simulation domain set-up and the different model parameters tested in this case study. In Section 3, I present the results of the simulations and evaluate the performance of the simulations. Lastly, I conclude with a discussion on the implication of the choice of model parameters and its impact on wind power production and wind resource assessment.

2.2 Domain Description and Simulation Set-up

The CBLAST-LOW campaign was conducted over the summer months of 2001 through 2003 in the North Atlantic, south of Martha’s Vineyard Island, MA, USA (Edson et al. 2007). The objectives of the campaign were to understand air-sea interaction and the dynamics of the coupled atmospheric-oceanic boundary layer. To achieve these goals, direct measurements of vertical fluxes of momentum, heat and mass were made across these coupled boundary layers. The spatial and temporal characteristics of coupled boundary layers were captured in order to identify the dynamic processes occurring within the boundary layers. A combination of radiosondes and remote sensing of the vertical structure of the coastal boundary layer yielded further insights into the dynamics of the coastal LLJ and the influence of synoptic events on its formation and structure (Mahrt et al. 2014; Helmis et al. 2015).

I selected a prominent LLJ case observed in the CBLAST-LOW 2001 study. A strong LLJ was captured by a Woods Hole Oceanographic Institution (WHOI) sounding at approximately 07 August 2001 0030 UTC (1930 LST) (Fig. 2.1). The radiosonde data provided temperature, wind speed and wind direction profiles which showed the characteristics associated with a LLJ: a temperature inversion up to 350 m above sea-level, supergeostrophic wind just above the temperature inversion, as well as wind veer of $\sim 20^\circ$. A cold frontal passage occurred during this observational period and is believed to have contributed to the formation of this coastal LLJ (Helmis et al. 2015).

Figure 2.1: Sounding profiles for wind speed (red), wind direction (green) and temperature (blue) measured with a rawinsonde launched on 08 August 2001 0030 UTC (1930 LST) at 41.52°N, 70.67°W from WHOI Ship. No wind data was available between 3 m and 226 m from the available sounding archive. (retrieved from <ftp://ftp.whoi.edu/pub/users/cblast/2001/rawinsonde/01080720.CAP>)

2.2.1 WRF Model Domain Configuration

The domain consists of three nests, centered upon Martha’s Vineyard Coastal Observatory (Fig. 2.2). The horizontal resolutions from the outermost domain, d01 to the innermost domain, d03 are 30 km, 6 km and 1.2 km respectively. The horizontal domain sizes ($N_x \times N_y$) are 100x100, 300x300 and 300x300 respectively. The time-steps used are 100 s, 20 s and 10 s respectively. All simulations used the same Noah Land surface model, Dudhia (Dudhia 1989) and RRTM (Mlawer et al. 1997) radiation scheme; the cumulus scheme is switched off for d03; the microphysics scheme used is the WSM 3-class scheme Hong et al. (2004). All simulations ran from 07 August 2001 0000 UTC (1900 LST) until 09 August 2001 0000 UTC (1900 LST) with the first 12 hours discarded as spin-up.

For the sensitivity studies, I have chosen to vary the initial and boundary conditions used, planetary boundary layer (PBL) schemes, vertical resolution and sea-surface temperature (SST) fields.

2.2.2 Model Configuration Options

Boundary Conditions

Previous studies have shown that the choice of initial and boundary condition datasets heavily influenced the characteristics of modelled LLJ (Nunalee and Basu 2014; Vanderwende et al. 2015). LLJ formation is dependent on multiple factors that include baroclinicity, topography and synoptic events such as fronts (Ostdiek and Blumen 1997; Banta et al. 2002; Van de Wiel et al. 2010; Vanderwende et al. 2015). There exist differences between various reanalysis products in their representation of synoptic events, dependent on the methods used to produce these reanalysis models. Hence it would be prudent to test the sensitivity of the modelled LLJ due to such possible differences.

I have explored two options for boundary conditions. First, ERA-Interim is a third generation global atmospheric reanalysis product from 1979 to present (Dee et al. 2011). It contains global atmospheric and surface parameters at T255 spectral resolution (~ 80 km) on 60 vertical levels, with 12 levels below 850 hPa. It uses 4D-VAR for data assimilation to produce atmospheric fields on model pressure levels at 6-hour intervals and surface fields available at 3-hour intervals. The North-

Figure 2.2: WRF domain set-up showing 3 nested domains centred on the location of the Martha’s Vineyard Coastal Observatory. D01 is the outermost domain, d02 is the nested intermediate domain and d03 is the innermost domain nested within d02.

American Regional Reanalysis (NARR) is a second-generation regional reanalysis product from 1979 to present, a high-resolution extension of the NCEP Global Reanalysis, focused over the North American Region (Mesinger et al. 2006). It contains regional atmospheric and surface parameters at 32 km horizontal resolution on 29 vertical levels, with 7 levels below 850 hPa. NARR uses 3D-VAR to assimilate meteorological variables into the NCEP Eta model to produce atmospheric fields that are available at 3-hour intervals. These datasets are summarized in Table 2.1.

Table 2.1: Comparison of boundary condition datasets

	ERA	NARR
Model Spatial Coverage	Global Reanalysis	2nd generation reanalysis Downscaled from NCEP Global Reanalysis focusing on North Amer- ica Region
Spatial Resolution	Spectral T255 ($\sim 80\text{km}$)	32km
Data-assimilation Method	4D-VAR	3D-VAR
Total Vertical Levels	60	29
Vertical Levels below 850 hpa	12	7
Temporal Resolution	6-hour intervals for verti- cal levels 3-hour intervals for sur- face	3-hour intervals

PBL Scheme

In mesoscale WRF simulations, the vertical structure of the atmosphere is parameterized by PBL schemes, which are used to represent sub-grid scale turbulent vertical fluxes of momentum, heat and moisture. Because a stable boundary layer suppresses buoyancy-driven mixing while having increased turbulence shear production aloft due to strong wind shear, in contrast to a convectively-driven boundary layer, the choice of closure schemes suitable for convective boundary layer might not be optimal for stable boundary layer (Nielsen-Gammon et al. 2010; Shin and Hong 2011).

In this study, I focus on the different results obtained by local and non-local closure schemes. These closure schemes differ in how the sub-grid fluxes are modelled as a function of mean flow quantities. Local closure schemes model these fluxes at each grid point based on the mean values of their respective gradients at that point. Non-local closure schemes would include an additional flux term to account for large-

scale vertical transport that is non-local in nature (Troen and Mahrt 1986). Local schemes such as the Mellor-Yamada-Nakanishi-Niino (MYNN) scheme are expected to perform better for stable boundary layer (Hu et al. 2010; Shin and Hong 2011; Hu et al. 2013a) because local mixing is the primary mode of turbulence in a stable boundary layer. Non-local closure schemes such as the Yonsei-University (YSU) scheme have been shown to be effective for simulating convective boundary layers (Hong 2010; Hu et al. 2010; Shin and Hong 2011). The MYNN PBL scheme is a turbulent kinetic energy (TKE)-based local mixing scheme implemented in the WRF-ARW model. It has the option to run in either level 2 or 3 closure. The closure constants are tuned to a set of LES simulations (Nakanishi and Niino 2004).

A prognostic TKE equation is used in the MYNN scheme

$$\frac{d(q^2/2)}{dt} - \frac{\partial}{\partial z} \left[l_m q S_q \frac{\partial(q^2/2)}{\partial z} \right] = P_s + P_b + \varepsilon \quad (2.1)$$

where $q^2 = 2 \times TKE$. The first left-hand side term represents the total derivative of q and the second term refers to the vertical distribution of q . On the right-hand side, the first and second term represents shear, P_s and buoyancy production, P_b respectively and the last term on the right-hand side is the dissipation term, $\varepsilon = q^3/B_1 l_m$, where B_1 is a constant.

The mixing length, l_m determines the vertical distribution term of q and its dissipation. The MYNN scheme formulates the mixing length as:

$$\frac{1}{l_m} = \frac{1}{l_s} + \frac{1}{l_t} + \frac{1}{l_b} \quad (2.2)$$

The mixing length term is a function of the surface layer length, l_s , turbulent length, l_t , and buoyancy length, l_b . Detailed formulation for these length scales can be found in Nakanish (2001). It has been formulated as such in order to allow the smallest length scale to dominate and determine the characteristic mixing length. This feature allows the MYNN scheme to transit across different stability regime smoothly.

In contrast, the YSU PBL scheme is a first-order non-local mixing scheme with a non-local counter-gradient term (Hong et al. 2006; Hong 2010). The YSU diagnoses the height of the PBL, h based on a critical Richardson number. Using the height of the PBL, an eddy diffusivity profile is formulated as:

$$K_m = \kappa w_s z \left(1 - \frac{z}{h}\right)^2 \quad (2.3)$$

where ($w_s = \frac{u_*}{\phi_m}$) is the velocity scale, u_* is the friction velocity, κ is the von Karman constant, z is the height above ground and h is the boundary layer height. The atmospheric stability function, ϕ_m in WRF is formulated as:

$$\phi_m = 1 + 5 \frac{z}{L} \quad (2.4)$$

where z is, again, the height above ground and L is the Monin-Obukhov length scale.

SST Field

The contrast in surface temperature between land and sea can contribute to LLJ formation in coastal regions (Mahrt et al. 2014). The land-sea temperature contrast can result in baroclinicity that is conducive to the formation of coastal LLJ, similar to the role of baroclinicity over the sloping Great Plains (Holton 1967). When warm air from the land surface advects offshore over cooler water, a strong stable marine boundary inversion can form (Smedman et al. 1997). The stratification of this marine boundary layer depends on the temperature difference between the warm land and the cooler sea. The stability profile will then influence the characteristics of the resultant LLJ above this inversion layer.

The standard WRF-ARW model configuration lacks a coupled ocean model, and so the SST field is provided as an input boundary condition and updated in time according to data availability. ERA-Interim provides a 6-hourly SST field that can be used to update the SST and thus the skin temperature field (Skamarock et al. 2008). For the NARR dataset, there is no explicit SST field available at 3-hour intervals. Instead, the SST field is derived from the initial air temperature at the surface over water bodies from NARR fields and the SST field remains constant over the duration of the simulation. To assess the importance of SST on LLJ characteristics, the simulations include a case where the SST field was not updated in an ERA-based run during the simulation period. This simulation also serves as a control against which NARR-based runs can be evaluated.

Vertical Resolution

The stable atmospheric boundary layer is characterized by smaller turbulent eddies compared to those of the convective boundary layer. Increasing the vertical reso-

Table 2.2: Names and summary of all the model configurations in the study

	Boundary Condition	PBL Scheme	SST Update	Vertical Grid Lev- els (eta)
ERA_MYNN	ERA-Interim	MYNN	YES	60
ERA_YSU	ERA-Interim	YSU	YES	60
NO_SST	ERA-Interim	MYNN	NO	60
ERA_MYNN_75	ERA-Interim	MYNN	YES	75
NARR_MYNN	NARR	MYNN	NO	60
NARR_YSU	NARR	YSU	NO	60

lution within the stable boundary layer would in theory increase the accuracy of the PBL parameterizations of the eddy fluxes. The improved accuracy near the surface and the LLJ core should then lead to better estimates of winds at a typical wind turbine hub-height of $\sim 80\text{m}$. In a study on high resolution modelling for wind energy purposes, Bernier and Blair (2012) found that increased vertical resolution only has modest improvements on wind profile predictions. A study on coastal LLJ (Nunalee and Basu 2014) also corroborated this finding. However, Bernier and Blair (2012) also noted that prediction of wind shear across the wind turbine blade layer was improved with higher resolution. To test the influence of vertical resolution, I executed a simulation with finer vertical resolution, increasing the total number of vertical levels to 75 from 60. As a result, the effective vertical resolution in the lowest 1000m increased from 35m resolution to 25m resolution. Table 4.4 summarizes the model configurations used in this study.

2.2.3 Observation Data

To validate the model results with observations and to apply the diagnostic metrics, I use surface observations from Automated Surface Observing Systems (ASOS) stations in Massachusetts and Rhode Island for temperature and wind speed (see map in Fig. 2.3).

and reduces the effect of surface roughness. The effect of reduced turbulence leads to an acceleration of the wind field aloft, thus forming a LLJ. The LLJ observed was partially a result of this frontal passage. Wind direction shifts associated with a frontal passage were recorded by some ASOS stations (not shown). Observed surface winds generally shifted from southerly to northeasterly direction.

Although this LLJ was captured in a rawinsonde sounding launched from the WHOI ship (Fig. 2.1), the single sampling of the atmosphere was not sufficient to evaluate the skill of the WRF model in modelling LLJ characteristics that I am interested in: timing, location, height and speed of LLJ. Thus, I employ the use of ASOS data to help fill in the gaps in observations. Even though only surface measurements were available from ASOS stations, the data will be able to capture the frontal passage during the simulation period. The frontal passage is likely a key modulator of the LLJ characteristics (Helmis et al. 2013), hence I can use the surface measurements to evaluate the ability of the WRF model to simulate the front and the condition of LLJ formation by proxy.

2.3 Results

The frontal passage is captured in all simulations although the timing and the spatial extent of the front differ between simulations. Horizontal cross sections at 200 m above sea-level of the temperature contours and wind vectors of ERA_MYNN and NARR_MYNN at 6-hour intervals depict the status of the evolving front (Fig. 2.4). The simulated frontal passage in ERA_MYNN (top panels) appears to extend further offshore compared to NARR_MYNN (bottom panels). The acceleration of the south-westerly winds at 2100 UTC (1600 LST) is represented in both simulations.

At 1500 UTC (1000 LST), no front is observed in d03: westerly winds are fairly steady at 10 m s^{-1} onshore and close to coast with stronger south-westerly winds further offshore at approximately 15 m s^{-1} . At 2100 UTC (1600 LST), the south-westerly winds intensified. Westerly winds onshore can be seen in the north-west of the domain. A distinct wind direction shift between the warmer continental air mass and the cooler maritime air mass can be seen in this north-west region. At 0300 UTC (200 LST), two regions of distinctly different wind directions are seen and this is indicative of a frontal passage passing through the domain from the north-west

Figure 2.4: Temperature contours at 200 m above mean sea-level during frontal passage domain 3. Wind speed and direction indicated by wind barbs in m s^{-1} . Grey area in plots indicate presence of terrain at 200 m above mean sea-level. The black pointer marks the center of the simulation domain. Top panels: ERA_MYNN; Bottom panels: NARR_MYNN

to the south-west, advected offshore.

Although each of the simulations described in Table 2.1 above produced a slightly different LLJ, they all do produce jets, as seen in Fig. 2.5. The resulting LLJs differ by their time of onset, duration, maximum wind speed, height of the wind speed maximum, and depth of the acceleration layer. Many of these differences emerge from different representations of the synoptic situation, specifically the timing and shape of the frontal passage discussed above.

I first explore the impact of boundary conditions on the simulations. Comparing ERA_MYNN and NARR_MYNN (top-left and bottom-left panels of Fig. 2.4), the cold frontal passage as evidenced by the band of wind direction shifts from the northern region of the domain appears to occur earlier in ERA_MYNN and

Figure 2.5: Time-height cross sections of horizontal wind speed at center of d03 for all 6 model configurations. Location is marked with a black pointer in Fig. 2.4. Period is from 07 August 2001 1500 UTC (1000 LST) to 08 August 2001 0300UTC (2200 LST)

extends much further offshore compared to NARR_MYNN. This difference in frontal passage timing is corroborated by the difference in LLJ initiation timing (top-left and bottom-left panels of Fig. 2.5). The difference in spatial extent of the warm air mass offshore is also shown in Fig. 2.6 where ERA_MYNN and ERA_YSU show warmer air mass aloft the marine boundary layer further offshore compared to NARR_MYNN and NARR_YSU. The intensity of the LLJ is also stronger in ERA-based simulations compared to NARR-based simulations, with a longer period of sustained wind speeds equal to or greater than 20 m s^{-1} at the jet-core (Fig. 2.5). Jet-core wind speeds exceeded 20 m s^{-1} for approximately 6 hours in ERA-based simulations and for only 2 hours in NARR-based simulations.

The differences in boundary-layer parameterizations (especially MYNN vs YSU) are constrained to near-surface regions, below the nose of the jet. The wind shear

Figure 2.6: NW-SE transect of domain 3 for all 6 model configurations on 08 August 2001 0030 UTC (1930 LST). Colored contours indicate temperature. Line contours indicate wind speed.

is stronger below 100m in both MYNN-based simulations compared to both YSU-based simulations (Fig. 2.5). Wind contours offshore reflect this behaviour as well, with stronger wind speeds close to the surface within the cooler marine boundary layer for the MYNN simulations (Fig. 2.6). This observation holds true regardless of the choice of boundary condition. The role of the PBL scheme appears to influence the depth of the marine inversion layer and hence the height of the LLJ core, and this finding is similar to previous works such as Olson and Brown (2009); Shin and Hong (2011); Talbot et al. (2012); Carvalho et al. (2012); Steele et al. (2013); Huang and Peng (2016).

Although increased vertical resolution was expected to improve representation of the LLJ, the increase in vertical resolution did not have a significant impact on the LLJ characteristics near the surface where more grid points were added. Wind shear below 100m appears to be slightly weaker over the LLJ duration in ERA_MYNN_75 compared to ERA_MYNN (Fig. 2.5); however the difference between the two MYNN simulations is not as large as the difference between MYNN and YSU. The depth of the LLJ when jet-core speeds are greater than 20 m s^{-1} is shallower in ERA_MYNN_75 (from 200 m to 350 m above mean sea-level) compared

to ERA_MYNN (from 150 m to 400 m). On the other hand, the heights of the simulated jet-core are at approximately 250m for both.

Lastly, in the NO_SST simulation where the SST was not updated during the period of simulation, the differences in simulated LLJ characteristics are less pronounced. The wind shear observed in Fig. 2.5 (top right) within the lowest 100 m was more similar to ERA_MYNN as compared to ERA_YSU and ERA_MYNN_75. The jet is slightly elevated compared to ERA_MYNN but this behaviour is also observed in ERA_YSU and ERA_MYNN_75. In Fig. 2.6, the NO_SST wind contours within the marine boundary layer are also similar to ERA_MYNN and the height of the marine boundary is also at approximately 200 m, in concert with other ERA-based simulations.

These qualitative observations offer some insight into the variability in the LLJ due to configuration options: the choice of the boundary condition exerted the greatest influence on resulting LLJ variability. However, it is still unclear as to which configuration simulated the actual conditions the best.

2.3.1 Comparison to CBLAST-LOW sounding profile

In order to decide which configuration options modelled the actual conditions best, I look to in-situ measurements of the atmospheric conditions during the LLJ event. The profiles for wind speed and temperature from the rawinsonde sounding shown in Fig. 2.1 are compared to all 6 model simulations. The profiles are extracted from the closest grid point to the location of the rawinsonde launch (41.52°N, 70.67°W).

Wind speeds are overestimated by all simulations within the first 500 m (left panel of Fig. 2.7). Wind speeds in ERA-based simulations are closer in magnitude to the sounding compared to NARR-based simulations, particularly around the height of the jet wind speed maximum. The heights of the wind maxima are approximately similar for all 6 simulations with the NO_SST simulation at the lowest height of 220 m and the ERA_MYNN_75 at the highest height of 300 m. However, all 6 simulations overestimate the height of the jet core compared to the sounding.

The NARR-based simulations underestimate the temperature below the inversion layer while also underestimating the strength of the inversion (right panel of Fig. 2.7). Conversely, the ERA-based simulations match the sounding to a good degree, with the exception of the ERA_YSU simulation. I conclude that the choice

Figure 2.7: Comparisons between sounding and simulation profiles on 08 August 2001 0030 UTC (1930 LST) at 41.52°N, 70.67°W. Left panel: horizontal wind speed; Right panel: temperature.

of PBL scheme seems to have an influence on the strength and depth of the marine inversion layer.

However, the LLJ is a 4D phenomenon that varies spatially and temporally (Banta et al. 2002; Rife et al. 2010), and so the single sounding profile comparison does not provide a robust basis to assess which simulation performs optimally.

2.3.2 Comparison to ASOS observations

Due to the scarcity of observational soundings during the simulation period, I utilize available ASOS surface observations to validate the simulations results and to objectively select the optimal combination of configurations.

I first bin the ASOS surface observations for surface temperature and wind speed into the closest 00/30 min for the LLJ duration of 1500 UTC (1000 LST) to 0300 UTC (2200 LST), then I extract the time-series for temperature at 2 m, T2 and wind speed at 10 m of the closest grid-point to each ASOS station in the simulation domain, d03. I perform the diagnostics suite on these multi-station pairs.

Changing from ERA to NARR as the boundary condition generally increased

NRMSE and NPE for most stations, regardless of PBL schemes chosen (Fig. 2.8). Stations that exhibit larger NRMSE and NPE are common within ERA- and NARR-based simulations. ERA-based simulations also show lower overall NBIAS compared to NARR-based simulations; NARR-based simulations show a strong NBIAS of approximately -2 for multiple stations, regardless of PBL scheme.

Comparing the effect of changing MYNN to YSU scheme to the effect of changing boundary conditions, the differences are less pronounced. Overall, MYNN-based simulations have smaller NRMSE due to decreased NBIAS and NPE as compared to YSU-based simulations. This holds true regardless of boundary condition used.

For NO_SST configuration, the effect of not updating SST in ERA_MYNN lead to slight increase in NBIAS for KBID, KOQU and KPVC stations but decreased NPE for the same stations. This shift is expected as these stations are situated in grid-points designed as water-bodies in WRF. When vertical resolution is increased in ERA_MYNN_75 there are marginal changes in error metrics when compared to ERA_MYNN .

Overall, all simulations NPE scores are similar in magnitude for surface temperatures, indicating comparable skill in pattern prediction. All simulations also show a cold bias for the surface temperature prediction while NARR-based simulations exhibit much stronger significant biases compared to ERA-based simulations.

When the simulations are evaluated based on their predictions of surface wind speeds, the contrast between simulation configurations is much more striking. ERA-based simulations have in general smaller NRMSE and NPE compared to NARR-based simulations, regardless of PBL scheme used (Fig. 2.9). This observation is similar to that of surface temperature comparisons. When the PBL scheme was changed from MYNN to YSU, the average NRMSE and NPE are increased for each configuration. However, it is also interesting to note that although average NPE is increased, the spread of NPE is reduced for both YSU-based simulations. Comparing the performance of NO_SST and ERA_MYNN_75, the average NRMSE and NPE scores are increased slightly, although it is difficult to determine a systematic error associated with each stations performance.

It is clear for this offshore jet case that boundary conditions introduce a stronger bias compared to those introduced by different PBL schemes, SST field, and vertical resolution. This finding is similar to that of (Vanderwende and Lundquist 2012)

modeling of LLJ in the Great Plains.

Based on the above metrics, I choose ERA_MYNN as the optimal set of simulation configurations and to be used as the base simulation for inter-model comparison diagnostics. ERA_MYNN offers the best mix of NRMSE/NPE/NBIAS scores for both T2 and wind speed amongst all the simulations.

2.3.3 Inter-model comparisons

ERA_MYNN is chosen as a reference model based on comparisons to ASOS surface observations. The purpose of the analysis in this section is to use the diagnostic tools to objectively quantify the variability in prediction introduced by different model parameters.

Much stronger temperature biases are introduced by boundary conditions compared to other model parameters such as PBL scheme, SST condition and vertical resolution. NARR-based simulations exhibit a strong cold bias mainly over sea surfaces and strong warm bias over land surfaces over the jet duration (Fig. 2.10). Biases are also seen in other model configurations but the spatial extent and amplitude of NBIAS are less than those in NARR-based simulations.

NARR-based simulations exhibit strong error variance in the south-eastern portion of the domain, over the sea surface (Fig. 2.11). For other model configurations, large error variance is confined mainly to regions just off the south-eastern coast. The prominent area of large α scores in NARR-based simulations over the sea indicates an anti-correlation of temperature between ERA-based and NARR-based simulations in this region which implies a phase error. This pattern suggests that the time of frontal passage that contributed to the LLJ formation was lagging in NARR-based simulations compared to ERA-based simulations

The jet itself and the frontal boundary can be seen in the northwest-southeast vertical transect of the domain (Fig. 2.12). The biases are similar to that in the horizontal domain: NARR-based simulations exhibit overall stronger biases compared to ERA-based simulations with cold bias over the sea and warm bias over land surfaces. ERA_YSU shows stronger warm biases over land compared to other ERA-based simulations.

The α score in the vertical slice of NARR-based simulations shows a spatial pattern that resembles a frontal boundary (Fig. 2.13). The strong α score arises from

Figure 2.8: Error decomposition diagram of surface temperature, T2, for all model configurations. NRMSE, δ , is the radial distance along the bottom of the plot. NPE, $\sigma = \sqrt{\alpha}$, is the upward vertical distance from the bottom axis and bound by $\sqrt{2}$. NBIAS, μ , is labeled for each radial line. The bold radial lines mark $\mu = \pm 0.3$ denoting bias values between that contributes negligibly to total error. Optimal model with $\delta = 0$ is marked with a blue circle.

an anti-correlation of temperatures between the NARR and ERA-based simulations and manifests over the sea surface, especially near the marine surface layer. This behaviour corroborates the α pattern seen in the horizontal (Fig. 2.11) where an associated phase error contributed to this variability in the simulations. In contrast, for the ERA-based simulations, the error variance is mainly confined to the region of sea surface near the coast, with little variation in the vertical profile.

Figure 2.10: Contour plot of NBIAS for temperature during LLJ event between 07 August 2001 1500 UTC (1000 LST) to 08 August 2001 0300 UTC (2200 LST) at 200m cross-section. ERA_MYNN is chosen as the reference dataset for comparison. NARR_MYNN refers to comparing NARR_MYNN to ERA_MYNN. Similarly for ERA_YSU , NARR_YSU , NO_SST and ERA_MYNN.

Figure 2.11: Contour plot of α for temperature during LLJ event between 07 August 2001 1500 UTC (1000 LST) to 08 August 2001 0300 UTC (2200 LST) at 200 m cross-section. ERA_MYNN is chosen as the reference dataset for comparison. NARR_MYNN refers to comparing NARR_MYNN to ERA_MYNN. Similarly for ERA_YSU , NARR_YSU , NO_SST and ERA_MYNN.

Figure 2.12: Contour plot of NBIAS for temperature during LLJ event between 07 August 2001 1500 UTC (1000 LST) to 08 August 2001 0300 UTC (220 LST) along the NW-SE vertical transect across domain 3 shown in Fig. 2.3. ERA_MYNN is chosen as the reference dataset for comparison. NARR_MYNN refers to comparing NARR_MYNN to ERA_MYNN. Similarly for ERA_YSU , NARR_YSU , NO_SST and ERA_MYNN.

Figure 2.13: Contour plot of α for temperature during LLJ event between 07 August 2001 1500 UTC (1000 LST) to 08 August 2001 0300 UTC (2200 LST) along the NW-SE vertical transect across domain 3 shown in Fig. 2.3. ERA_MYNN is chosen as the reference dataset for comparison. NARR_MYNN refers to comparing NARR_MYNN to ERA_MYNN. Similarly for ERA_YSU , NARR_YSU , NO_SST and ERA_MYNN.

2.4 Discussion

The strong α score (> 1) offshore as seen in both Fig. 2.11 and Fig. 2.13 arises from an anti-correlation of the temperature signal in that region. This phase error can be explained by the dynamics of the frontal passage. NARR-based simulations modelled a frontal passage that passes through the domain later in time and does not have the same spatial extent as ERA-based simulations.

Figure 2.14: Pressure field at mean sea-level for domain 3 from ERA and NARR reanalysis dataset used as initial and boundary conditions for simulation runs.

Examination of the boundary forcing in ERA and NARR dataset shows slight variation in pressure fields (Fig. 2.14). A sharper pressure gradient extends across the domain in ERA, as compared to NARR; the low pressure system exists further south in NARR. The difference in the resulting geostrophic wind could result in a different advection rate of the front as well as the formation of the LLJ at the frontal boundary (Ostdiek and Blumen 1997).

In contrast to the role of ERA/NARR boundary conditions, the choice of PBL scheme, vertical resolution and role of SST updates were less significant in model variability. The variability in the ERA-based simulations is more apparent in the vertical profile and mainly manifested as NBIAS.

The choice of PBL scheme, SST and vertical resolution mainly influenced the height of the marine boundary layer and the hence the height of the LLJ core. The choice of YSU scheme over MYNN resulted in an overall warmer profile over land due to enhanced mixing from the land surface that is then advected offshore, which is seen in YSU's warm bias above the marine boundary layer. The role of SST is less significant than expected. Of course, SST fields are updated every 6 hours in ERA and in $0.75^\circ \times 0.75^\circ$ resolution, which may be too coarse in space and time to have a significant impact on the land-sea temperature contrast.

2.4.1 Implication for wind power

With increased interest in using NWP for offshore wind resource assessment (Archer et al. 2014; Badger et al. 2014; Hahmann et al. 2014; Nunalee and Basu 2014), it is important to characterise the variability between model performance that would have an impact on wind power prediction.

To quantify the potential effect of the LLJ has on wind power prediction, it would be useful to calculate the wind power production during the LLJ occurrence. Using the power curve for the VESTAS V100 2.6 MW turbine (Vestas cited 2015) , I calculate the wind power based on rotor equivalent wind speed (REWS) (Wagner et al. 2011) for a turbine situated in the centre of the domain. The REWS is a method that characterizes the kinetic energy flux, hence the wind profile, across the entire turbine blade and takes into account the vertical wind shear experienced. The utilization of REWS is expected to improve annual energy production estimates and reduce uncertainty in power curve measurements.

The variability in the REWS (top panel of Fig. 2.15) results in differences in total power output (bottom panel of Fig. 2.15) between each model configuration (summarized in Table 2.3) . Once again, the choice of boundary conditions resulted in the highest variability between total power output when compared to PBL schemes, SST option or finer vertical resolution. Changing from ERA to NARR resulted in a decrease in predicted power output of approximately 1.6 MWh, regardless of PBL schemes whereas switching MYNN to YSU only resulted in a decrease of approximately 1.1 MWh and other configurations led to even smaller differences. This 1.6 MWh (5% of ERA_MYNN total power output) difference in total power output due to boundary conditions in the above case-study has potential implications for long

Table 2.3: Summary of average rotor equivalent wind speeds during LLJ duration from 1500 UTC (1000 LST) to 0300 UTC (2200 LST) and total power output calculation is sum of power output at each 30 minute time interval over the entire duration of the LLJ. Normalized power output is the total power output of each model configuration normalized by the total power output by ERA_MYNN

	Average tor Wind Speed (m s^{-1})	Ro- Equivalent (m Output (MWh)	Power (MWh)	Normalized Power Output (%)
ERA_MYNN	15.04		30.19	100.0
NARR_MYNN	13.16		28.58	94.7
ERA_YSU	13.13		29.05	96.2
NARR_YSU	12.41		27.44	90.9
NO_SST	13.79		29.13	96.5
ERA_MYNN_75	13.48		29.53	97.8

term wind resource assessment using NWP. However, a longer systematic climatological study will have to be carried out to assess the variability in mesoscale modeling by choice of boundary conditions. Wind veer, where there is a wind directional dependence on height, is another possible source of power variability (Rareshide et al. 2009; Vanderwende and Lundquist 2012) but it is not explored in this study.

Figure 2.15: Top panel: Rotor equivalent wind speed time-series for a single V100 2.6 MW wind turbine situated in the centre of domain d03 over the duration of the LLJ. The swept rotor area extended from 30 m to 130 m above the surface. Bottom panel: Power output time-series for a single V100 2.6 MW based on equivalent wind speed and power curve

Conclusions

In this study, I simulate a LLJ event observed during CBLAST-LOW 2001 campaign off the New England coast and characterize the variability in the LLJ timing, height and speed of the LLJ core due to different model configurations. I varied the choice of initial and boundary condition dataset, PBL scheme, SST update frequency and vertical resolution.

The timing of the LLJ initiation is most dependent on the boundary condition dataset (ERA or NARR) (Fig. 2.5), whereas the height of the LLJ core responds to the choice of PBL scheme. I evaluate simulations against observations to establish an optimal configuration suitable for this case-study. I then compare wind speed and temperature profiles from a rawinsonde sounding and find that wind speeds are overestimated by all models, although ERA-based simulations show the smallest bias. The height of the wind speed maximum is also overestimated by all models. When I examine the temperature profiles, ERA-based simulations are able to model the strength and depth of the temperature inversion layer much better than NARR-based simulations. NARR-based simulations show a strong cool bias of $-2\text{ }^{\circ}\text{C}$ below the inversion layer. The influence of the PBL schemes on wind speed and temperature profile manifests differently in ERA compared to NARR. However, a single sounding is not suitable to provide a good determination of models' performance in modelling a 4D phenomenon. Thus, I utilize surface observations from ASOS stations for further validation.

In order to objectively determine model performance, I employ the diagnostic suite described in Koh et al. (2012). This set of diagnostic tools allows us to decompose errors into bias and pattern errors, which provide further insights into the possible sources of variability between model configurations.

From the diagnostics of surface temperature, I find that all simulations tend to exhibit a cold bias error, although the majority of stations within each simulation have small pattern error. This lead us to conclude that models have comparable skill in variance prediction but the biases are likely driven by differences in boundary conditions between ERA and NARR. For the wind speed diagnostics, the NRMSE are overall higher compared to temperature and the NBIAS varies across stations and simulations. While the magnitude of NBIAS changed between model configurations, stations that with significant NBIAS (> 0.3 or < -0.3) did not change signs. Based

on the diagnostic metrics derived from ASOS stations, ERA_MYNN offered the best mix of NRMSE/NBIAS/NPE performance, and hence I chose ERA_MYNN as the reference model for the inter-model comparison study.

In the inter-model study, I compare the performance of all other model configurations to ERA_MYNN. Between different ERA/NARR datasets, the error is due to a phase difference in the frontal passage through the domain dictated by the boundary conditions. The variability between different model configurations arises mainly from bias errors which can be explained by the parameterization of the PBL scheme and the changes in vertical resolution.

The diagnostics suite also provides metrics to diagnose wind as a vector quantity however it was not performed in this study. The ASOS surface observations for wind speed and direction are not robust enough to decompose the wind into vectors for analysis. Also, the frontal passage was the major synoptic event that heavily influenced the LLJ formation hence it is prudent to diagnose the temperature change which serves as a proxy for the LLJ initiation.

The variability in model configurations induced strong variability in wind power production during the time period of this LLJ. Although the LLJ winds were frequently strong enough to enter the flat region of the power curve, where wind speed forecast errors do not affect power production estimates, the difference in LLJ onset timing introduced variations in power production on the order of 1.6 MWh or 5% of the production during this time period. This variability underscores the need for more observational datasets in coastal and offshore regions to validate model simulations (Archer et al. 2014), for both atmospheric studies into coastal dynamics as well as the wind industry's need for reliable annual energy production estimate improvements.

Chapter 3

Validation of Mesoscale Simulations of Coastal Low-Level Jet with High Resolution Doppler Lidar

3.1 Introduction

The increased use of remote sensing equipment such as doppler lidar and radar systems to probe the lower boundary layer (Newsom et al. 2008; Berg et al. 2015a) and to study the wind turbines (Aitken et al. 2012; Iungo et al. 2013; Aitken and Lundquist 2014; Iungo and Porté-Agel 2014; Banta et al. 2015) and wind farms, have yielded much high fidelity observations and insights into atmospheric interactions. These observational datasets have been also used to validate computational models (Mirocha et al. 2015; Vanderwende et al. 2015). Such simulations and observations can work in concert to assess, predict and improve wind modelling for the wind energy industry.

Similar to Chapter 2, the WRF model is used to simulate coastal LLJs observed in the New England coastal region. In addition to rawinsonder soundings, high-resolution scanning lidar data from the New England Air Quality Study - Intercontinental Transport and Chemical Transformation (NEAQS-ITCT) campaign is also used to evaluate the mesoscale simulations.

I address the following questions in this chapter:

1. How sensitive are the timing of LLJ formation, height of the LLJ core and the strength of the LLJ core to the model parameters such as choice of initial/boundary conditions, planetary boundary layer schemes and sea-surface temperature forcing?
2. How can I quantify the influence that the model parameters have on the simulated jet?
3. How does the model performance differ in different LLJ regimes?

In Section 2, I describe the simulation domain set-up and the different model parameters tested in this case study. In Section 3, I present the results of the simulations and evaluate the performance of the simulations. Lastly, I conclude with a discussion on the implication of the choice of model parameters.

3.2 Domain Description and Simulation Set-up

The New England Air Quality Study - Intercontinental Transport and Chemical Transformation (NEAQS-ITCT) campaign was carried out over the period of June to August 2004. It was carried out under the International Consortium for Atmospheric Research on Transport and Transformation (ICARTT) umbrella, part of an international collaboration to perform air quality studies over North America, North Atlantic Ocean and western Europe (Fehsenfeld et al. 2006). NEAQS-ITCT concentrates on the Northeastern region of the United States and deployed multiple measurement platforms, including airborne measurements, land-based systems and shipboard instruments for the study. The shipboard data provides an invaluable look at offshore coastal conditions that are not readily accessible.

The NOAA Research Vessel *Ronald H. Brown* was a key platform used in NEAQS-ITCT. It was equipped two remote and one in situ sensors during this mission to measure wind profiles (Wolfe et al. 2007). A radar wind profiler measured hourly profiles at 60 m and 100 m vertical resolutions up to 3-5 km above ground level. Four to six rawindsondes were launched daily to assess profiles of winds, pressure, humidity and temperature. A high-resolution Doppler lidar (HRDL; Grund et al. (2001)) was also used to continuously monitor low-level wind profiles (Pichugina et al. 2008,

2012; Banta et al. 2015). The HRDL reported horizontal wind speed and direction profiles extending from the surface to about 2000 m ASL at approximately 15 minutes interval with vertical resolutions ranging from 1 m near the surface to 15 m at the upper end of the altitude range.

3.2.1 Case Selection

The baroclinicity that arises from the land-sea temperature contrast and sloping marine boundary layer inversion provide favourable conditions for LLJ events (Colle and Novak 2010; Helmis et al. 2013; Mahrt et al. 2014). In addition, LLJs may also form over land at night and advect offshore (Angevine et al. 2006). Frequent coastal LLJ events occurred during the field study. It is not feasible to perform nested LES studies for each of the LLJs observed due to computational constraints. Hence the following criteria were set to narrow the focus for case selection. The time period should:

1. feature different LLJ events;
2. have continuous HRDL profiles for at least 2 hours in duration featuring the LLJs;
3. have continuous HRDL profiles up to 500 m in height during the 2-hour period;
4. have no major synoptic events such as fronts within the domain.

The continuous requirement of HRDL profiles in duration and height is because the resulting mesoscale simulations will be used for nested LES simulations in the following chapter; therefore these HRDL profiles will be needed for the validation study to be as complete as possible. As described in Chapter 2, different representation of fronts will adversely impact the variability of the simulated LLJs, hence the requirement for a relatively quiescent synoptic condition. Moreover, in this study, I am also interested in the coastal LLJ formation due to inertial oscillation and baroclinicity arising from land-sea thermal gradient contrast of coastal regions (Burk and Thompson 1996; Mahrt et al. 2014). Hence, the period of 30 July 2004 0000 UTC to 01 August 2004 0000 UTC was identified to best fit the stated criteria.

3.2.2 Case Description - ASOS

I first review at the available surface observations from ASOS stations in the neighboring states of Massachusetts, New Hampshire and Maine. The ASOS data provide a general view of the mesoscale conditions during the selected time periods. I used only the surface temperature, wind speed and wind direction variables in this study. The ASOS surface observations are available in approximately 15-minute intervals and hence I bin the variables into the closest 00/15/30/45 minute.

I can see the diurnal cycle from the surface temperature observations (Fig. 3.1). The average surface temperatures between 30 July 2004 0000 UTC to 1200 UTC were lower than the corresponding period on 31 July 2004. The diurnal temperature ranges observed at coastal locations such as KRKD, KPVC, KHYA, KCQX and KMOV were less than those observed at onshore locations. The stations in the northern half of the domain experienced warmer conditions in the time period between 00 UTC and 12 UTC on the second day compared to the same time period on the first day.

From the surface wind speed observations (Fig. 3.2), no diurnal trend can be discerned. However, the average wind speeds were much higher for the period of 31 July 2004 1200 UTC to 0000 UTC compared to the same period on 30 July 2004. The highest wind speeds ($\sim 9.5 \text{ m s}^{-1}$) observed are at coastal stations KBOS and KEWB. This period of high speeds is also seen in offshore HRDL measurements, shown in the later sections.

From the surface wind direction observations (Fig. 3.3), no diurnal trend can be discerned. For stations close to the coast, no wind reversal trend characteristic of a land-sea breeze system appears. The winds fluctuate between the westerly and southerly directions. The wind directions were more variable on the first day (30 July 2004) compared to the second day. The period with variable wind directions also correlate with weaker wind speeds. On the second day (31 July 2004), most of the stations reported stronger south-westerly winds.

In summary, the ASOS data shows that the region was not dominated by any frontal passages during this time period and the surface winds were light and variable on the first day and strengthened in the south-westerly direction on the second day.

ASOS Stations Location

Figure 3.1: Time series of surface temperature for ASOS stations from 30 July 2004 0000 UTC (1900 LST) to 01 August 2004 0000 UTC (1900 LST)

ASOS Stations Location

Figure 3.2: Time series of surface wind speed for ASOS stations from 30 July 2004 0000 UTC (1900 LST) to 01 August 2004 0000 UTC (1900 LST)

ASOS Stations Location

Figure 3.3: Time series of surface wind direction for ASOS stations from 30 July 2004 0000 UTC (1900 LST) to 01 August 2004 0000 UTC (1900 LST))

3.2.3 Case Description - HRDL Profiles

The profiles in Figs.3.4 and 3.5 are HRDL lidar measurements interpolated to 10 m height intervals from 30 July 2004 0000 UTC to 01 August 2004 0000 UTC. The blank spaces in the profiles indicate the lack of HRDL measurements. Two distinct LLJ events occur during this time period. A weak jet develops between 30 July 0000 UTC and 30 July 0500 UTC that reaches a maximum speed of 11 m s^{-1} at 60 m and a strong jet is seen between 30 July 1800 UTC till 01 August 0000 UTC with a maximum speed of 19 m s^{-1} at 100 m. The jet profiles are predominantly south-southwesterly near the surface and veering to a more west-southwesterly direction above 300 m. The hours between the 2 jet events are marked by weak winds that are predominantly west-southwesterly and west-northwesterly aloft.

The selected 2-day period offers a chance to test the ability of the WRF model in simulating LLJ events of varying strength and the ability to model the transition in between these two jets.

3.2.4 WRF Model Domain Configuration

The domain consists of three nests, such that the innermost nest encapsulates the ship track over the 2 day duration in its center. Fig. 3.6 shows the domain and Fig. 3.7 shows the innermost domain, d03 with the ship track in colorscale. The horizontal resolutions from the outermost domain, d01 to the innermost domain, d03 are 31.25 km, 6.25 km and 1.25 km respectively. The domain sizes ($N_x \times N_y \times N_z$) are 100x100x60, 300x300x60 and 300x300x60 respectively. The 60 vertical model levels are concentrated mainly within the lower boundary layer and progressively stretched upwards. The average vertical resolution within the lowest 150 m is ~ 15 m and within the lowest 1000 m is ~ 35 m. The time-steps used are 100 s, 20 s and 10 s respectively. All simulations used the same NOAH Land surface model, short-wave Dudhia (Dudhia 1989) and longwave RRTM (Mlawer et al. 1997) radiation schemes; the Kain-Fritsch cumulus scheme is switched off for d03; the WSM 3-class microphysics scheme Hong et al. (2004) is used. All simulations ran from 29 July 2004 1200 UTC (1900 LST) to 01 August 2004 0000 UTC (1900 LST) with the first 12 hours discarded for model spin-up.

For the sensitivity studies, I have chosen to vary the initial and boundary condi-

Figure 3.4: Time series of horizontal wind speed profile along ship track from 29 July 2004 1800 UTC (1300 LST) to 01 August 2004 0000 UTC (1900 LST)

Figure 3.5: Time series of horizontal wind direction profile along ship track from 29 July 2004 1800 UTC (1300 LST) to 01 August 2004 0000 UTC (1900 LST)

tions (IBC), planetary boundary layer (PBL) schemes and sea-surface temperature (SST) forcings.

Figure 3.6: WRF domain set-up showing 3 nested domains. D01 is the outermost domain, d02 is the nested intermediate domain and d03 is the innermost domain nested within d02.

3.2.5 Model Configuration Options

Boundary Conditions

In addition to the two IBC datasets (ERA-Interim and NARR) used in Section 2.2.2, the NCEP FNL Operational Global Analysis dataset is also tested, hereby referred to as GFS. Unlike the other two IBCs, GFS is not a reanalysis product. Instead, it is a derived product from the Global Data Assimilation System which collects observational data from multiple global sources for analysis. It uses the same model as the Global Forecast System (NCEP et al. 2000). It contains global atmospheric

Figure 3.7: Map of NEAQS ship track from 30 July 2004 0000 UTC (1900 LST) to 01 August 2004 0000 UTC (1900 LST)

and surface parameters at 1-degree resolution on 26 vertical levels. The fields are available at 6-hour intervals.

PBL Scheme

The PBL schemes used in this study are the MYNN and YSU scheme described in Section 2

SST Field

As described in Section 2.2.2, the standard WRF-ARW model configuration lacks a coupled ocean model and requires SST field to be used as input boundary condition and updated in time as necessary. In addition to using the SST field associated with each IBC: ERA-Interim, NARR and GFS, I also perform two additional model simulations with modified SST field to assess the importance of SST compared to other atmospheric fields in LLJ variability. First, ENT_MYNN uses the ERA-Interim IBC for all the atmospheric fields except the surface skin temperature field. Instead, the surface skin temperature field (over both land and sea) is initialized using NARR surface skin temperature field. The purpose of ENT_MYNN is to

Table 3.1: Names and summary of all the model configurations in the study

	Boundary Condition	PBL Scheme	SST Forcing
ERA_MYNN	ERA-Interim	MYNN	ERA-Interim
ERA_YSU	ERA-Interim	YSU	ERA-Interim
NARR_MYNN	NARR	MYNN	NARR
ENT_MYNN	ERA-Interim	MYNN	NARR-Interim
GFS_MYNN	GFS	MYNN	GFS
GFS_SST	GFS	MYNN	GFS +2 K in d03 SST

test to what extent the skin temperature field (and SST) affects the LLJ formation and evolution. Second, GFS_SST uses the same GFS dataset for initialization and boundary conditions except the SST. The SST in d03 of GFS_SST is increased by 2 K, reducing the overall land-sea temperature contrast and test the sensitivity of the LLJs to this new forcing.

Table 3.1 summarizes the model configurations used in this study.

3.3 Results

3.3.1 Model Wind Speed Profiles

LLJ Criteria

Vanderwende et al. (2015) suggests a LLJ criteria developed by Bonner (1968) and modified by Whiteman et al. (1997) and Song et al. (2005) to identify and class LLJs into different intensities. For a LLJ to be identified in a profile, there needs to be (1) a wind maximum above a certain threshold, (2) this wind maximum must be below 2 km and (3) a decrease of wind speed above another certain threshold above this wind maximum. The LLJ intensities will be classed according the thresholds specified. While, this method is a robust identification of LLJ classes, however, this is not suitable for use because in this study, the HRDL profiles do not always

extend above 1 km and are most complete up to 500m. The imposed shear criterion of at least 5 m s^{-1} would reject most of the HRDL profiles containing a positive LLJ, because the lack of vertical extent in the HRDL profiles would not show the winds reducing back to geostrophic speeds. Given the limited extent of the HRDL profiles, I prefer to use the criteria proposed by Banta et al. (2002) which requires a 1.5 m s^{-1} wind speed difference above and below a jet maximum above a 10 m s^{-1} threshold for a positive LLJ identification. Fig. 3.8 shows the identification of LLJ event based on this criteria.

LLJ Timing

Broadly speaking, all simulations were able to model both LLJ events seen in the HRDL profiles as shown in Figs. 3.9 and 3.10. However, the timing of both LLJs are poorly represented in the simulations. The first LLJ observed between 30 July 2004 0000 UTC and 0600 UTC has a wind maximum occur about 0300 UTC before subsiding into a quiescent state. All model configurations had the LLJ peak around 0000 UTC indicating a 3-hour lead in the simulations. On the other hand, for the second LLJ observed, the LLJ initiated around 1800 UTC compared to initiation of the simulated LLJ at around 2100 UTC. The choice of IBC influenced the Day 2 LLJ initiation time: GFS-based simulations showed an earlier jet initiation compared to ERA-based and NARR-based simulations. The choice of PBL and SST changes did not influenced the LLJ initiation. This points towards larger scale effects than local conditions in being responsible for the LLJ formation in this case-study.

LLJ Height

All the simulations overestimated the height of LLJ core. The first LLJ event has an average LLJ height of 55 m while the simulated LLJs average heights are in the range of 115 m. There is little difference between the LLJ simulated by the 2 simulations, NARR_MYNN and GFS_SST, that had a positive LLJ identification. The second LLJ event has an average LLJ height of 100m while the simulated LLJs average heights are in the range of 180 m to 200 m.

Figure 3.8: Plot of LLJ identification using the criteria of a 1.5 m s^{-1} wind speed difference above and below a threshold speed equal to 10 m s^{-1} . Gray: Negative LLJ; Colors: Positive LLJ, color-coded according to the IBC used, blue (ERA), red (NARR), magenta (ENT), yellow (GFS) and green (HRDL).

LLJ Strength

All the simulations performed similarly in the prediction of the LLJ strength. For the first LLJ, all simulations under-predicted the strength of the LLJ. For the second LLJ, simulations underestimated the strength of the LLJ during the onset while intensifying and over-predicting the strength of the LLJ in the final hours of the selected time period.

Although the simulations were able to simulate broad features of the LLJ events, the results differed from observations and variability between model configurations are also observed. I thus seek to objectively validate the simulation results against observations and explain the possible sources of variability in the following sections.

Figure 3.9: Time series of horizontal wind speed profile along ship track from 30 July 2004 0000 UTC (1900 LST) to 01 August 2004 0000 UTC (1900 LST)

Figure 3.10: Time series of horizontal wind direction profile along ship track from 30 July 2004 0000 UTC (1900 LST) to 01 August 2004 0000 UTC (1900 LST)

3.3.2 Comparison to ASOS data

In this section, I evaluate the performance of the simulations against ASOS data. As described in Section 3.2.2, the ASOS data provides a glimpse of the atmospheric conditions occurring in the domain within the time-period. The ASOS data with a much wider spatial coverage compared to the sounding and HRDL profiles will allow us to examine possible sources in variability due to the IBC dataset used.

The model generates outputs in 15-minute intervals, to match the temporal frequency of the ASOS observations. The time-series for both temperature at 2 m and wind speed at 10 m of the closest grid-point to each ASOS station in the simulation domain, d03 are extracted. I then perform the diagnostic suite on these multi-station pairs, similar to the procedure performed in Chapter 2.

I evaluate the performance of the simulations for the full duration from 30 July

2004 0000 UTC (1900 LST) to 01 August 2004 0000 UTC (1900 LST). The diagnostic metrics for T2 and surface wind speed are plotted out in Fig. 3.11.

Figure 3.11: Error decomposition diagram of surface temperature, T2, (top) and surface wind speed (bottom) for all model configurations from 30 July 2004 0000 UTC (1900 LST) to 01 August 2004 0000 UTC (1900 LST). NRMSE, δ , is the radial distance along the bottom of the plot. NPE, $\sigma = \sqrt{\alpha}$, is the upward vertical distance from the bottom axis and bound by $\sqrt{2}$. NBIAS, μ , is labeled for each radial line. The bold radial lines mark $\mu = \pm 0.3$ denoting bias values between that contributes negligibly to total error. Optimal model with $\delta = 0$ is marked with the green circle. Each marker represents the diagnostic metric for each ASOS station.

In the top panel of Fig. 3.11, all simulations exhibit significant ($NBIAS < -0.3$) cold bias over the entire time period. The NBIAS for all the simulations lie mainly between -0.5 and -2.0. NARR_MYNN exhibit the higher NBIAS compared to the other configurations. It is also noted that despite increasing SST by 2 K in GFS IBC, there is little change in overall NBIAS pattern in GFS_SST. The simulations generally showed low NPE which indicates good skill in variance prediction. Majority of the stations in all the simulations gave low NRMSE scores ($\delta < 0.5$) with a few outliers with NRMSE score greater than 1.

All simulations mostly exhibit significant positive NBIAS in wind speed prediction (bottom panel of Fig. 3.11). The magnitude of the positive NBIAS is overall greater than the negative NBIAS for the simulations and ranges between 0.3 and 1.5. The simulations show good skill in variance prediction for wind speeds with overall low NPE scores. Majority of the stations in all simulations gave low NRMSE scores ($0.25 < \delta < 1$).

There is no clear trend that indicates which model performed the best. All simulations showed similarly good skill in simulating surface temperature and wind speed for the entire time period. The simulations overall performed better in predicting surface temperature compared surface winds as evidenced by the overall lower NRMSE scores. However, the biases are more significant in surface temperature prediction due to higher overall NBIAS scores.

Next, I divide the full time period equally into 2 24-hour time periods: 30 July 2004 0000 UTC to 31 July 2004 0000 UTC (Day 1) and 31 July 2004 0000 UTC to 01 August 2004 0000 UTC (Day 2). I then examine the performance of the simulations for each individual day. The error decomposition diagrams are shown in Figs. 3.12 and 3.13 respectively.

I see that the simulations have lower NBIAS scores for temperature and wind speeds on Day 1 compared to Day 2. Similar to the full time period, the simulations show higher significant biases for temperature compared to wind speeds on both days. Also, the overall NBIAS for the simulations increased from Day 1 to Day 2 for both temperature and wind speeds.

The simulations' skill in variance prediction are similar on both days for temperature with majority of the NPE scores between 0.2 and 0.75. On the other hand, the simulations' skill in variance prediction is better on Day 2 ($0.25 < \sigma < 0.75$)

Figure 3.12: Error decomposition diagram of surface temperature, T2, (top) and surface wind speed (bottom) for all model configurations from 30 July 2004 0000 UTC (1900 LST) to 31 July 2004 0000 UTC (1900 LST). NRMSE, δ , is the radial distance along the bottom of the plot. NPE, $\sigma = \sqrt{\alpha}$, is the upward vertical distance from the bottom axis and bound by $\sqrt{2}$. NBIAS, μ , is labeled for each radial line. The bold radial lines mark $\mu = \pm 0.3$ denoting bias values between that contributes negligibly to total error. Optimal model with $\delta = 0$ is marked with the green circle. Each marker represents the diagnostic metric for each ASOS station.

compared to Day 1 ($0.5 < \sigma < 1$) for wind speeds. The simulations also shows slightly better NPE scores for temperature compared to wind speeds on both days,

Figure 3.13: Error decomposition diagram of surface temperature, T2, (top) and surface wind speed (bottom) for all model configurations from 31 July 2004 0000 UTC (1900 LST) to 01 August 2004 0000 UTC (1900 LST). NRMSE, δ , is the radial distance along the bottom of the plot. NPE, $\sigma = \sqrt{\alpha}$, is the upward vertical distance from the bottom axis and bound by $\sqrt{2}$. NBIAS, μ , is labeled for each radial line. The bold radial lines mark $\mu = \pm 0.3$ denoting bias values between that contributes negligibly to total error. Optimal model with $\delta = 0$ is marked with the green circle. Each marker represents the diagnostic metric for each ASOS station.

however the spread of NPE scores is greater in temperature than wind speeds.

Overall, based on NRMSE scores the simulations performed better for tempera-

ture on Day 1 and for wind speeds on Day 2. Once again, no clear trend to indicate if any model configuration is favored on either days.

Next, I look at the spatial distribution of the NBIAS (Fig.3.14) for the entire full duration. There is no common distinct region (coastal or inland stations) in each of the simulation domains that exhibit strong cold NBIAS. As noted previously, the differences in NBIAS prediction between GFS_MYNN and GFS_SST are marginal for coastal stations despite having a 2 K warmer SST field. Also, the magnitudes of NBIAS for ENT_MYNN are more similar to ERA_MYNN compared to NARR_MYNN. This suggests that the surface skin temperature is not a big modifier in model performance for T2 predictions and points to other surface variables at play such as surface fluxes and soil conditions.

3.3.3 Comparison to NEAQS sounding profiles

Wind profiles (Wolfe et al. 2007) and rawinsonde soundings launched from the ship deck are available during this period at approximately 6-hour intervals. I present 3 of the 8 sounding profiles available during this time period. The first (Fig. 3.15) of the three soundings is the only available sounding that features the weak LLJ. The other two soundings (Figs. 3.16 and 3.17, 12 hours apart, feature the strong LLJ during daytime conditions. These three soundings are representative of the conditions when a nocturnal LLJ and a daytime LLJ occur offshore.

The first available sounding during the first observed LLJ event in Day 1 is on 30 July 2004 0500 UTC (Fig. 3.15). The observed virtual temperature profile (top left panel of Fig. 3.15) shows a shallow slightly stable surface-based layer and topped by an almost neutral layer above 200 m. The simulations are able to replicate features of this temperature profile, however, the shallow stable layer in the simulations are much more stable than observed. All the model profiles also show an overall cold bias compared to the observed sounding. Despite the similar temperature profiles, the wind profiles are much more different. There is still an evidence of a pronounced jet structure in the observations however all the simulations show weaker winds within the first 500 m. NARR and GFS simulations have a slight jet-like structure within 200 m while absent in ERA simulations. The discrepancy between the simulations and observations is also much more distinct in the hodograph (bottom right of Fig. 3.15) at this time. The observed winds are south-westerly near the surface and

Figure 3.14: Spatial distribution for T2 NBIAS for each ASOS station. The marker area is scaled to the magnitude of each station's NBIAS. Blue markers indicate cold NBIAS, red markers indicate warm NBIAS and green markers indicate insignificant NBIAS.

steadily veer westerly aloft before showing signs of backing in the top 200 m of the profile. The simulations are only able to model aspects of the wind veer but they do not agree well with the observations. GFS simulations showed good agreement near the surface, with south-westerly winds that veers aloft however, the winds veer westerly at 200 m and continues to veer north-westerly before backing to a westerly direction. NARR_MYNN on the other hand, while its wind veer is confined within the south-west quadrant, starts out too southerly near the surface. ERA simulations

Figure 3.15: Profile on 30 July 2004 0500 UTC (0000 LST) for wind speed (top left), virtual potential temperature (top right), hodograph (bottom left) and map of domain with position of sounding marked with an asterisk (bottom right). The height of the diagnosed PBLH of each model is marked on the wind speed profile (top left). The enlarged markers in the hodograph (bottom left) indicate the wind velocity at the lowest level (50 m) and each subsequent marker indicate wind velocity at the next higher level in 50 m interval.

seem to perform the worst, both ERA_MYNN and ERA_YSU start out too westerly and veers north-westerly within a shallow 300 m layer before backing aloft. The ship was positioned close to shore at the time of sounding and hence the differences in predicted wind direction would indicate different sources of the warm air advection which can lead to the variability in the wind profiles.

The next sounding profile, 31 July 2004 1130 UTC features a Day 2 jet profile (Fig. 3.16). I observe a statically stable atmosphere (top left panel of Fig. 3.16) with three distinct layers of decreasing stability. A strong shallow stable layer exists

Figure 3.16: Profile on 31 July 2004 1130 UTC (0630 LST) for wind speed (top left), virtual potential temperature (top right), hodograph (bottom left) and map of domain with position of sounding marked with an asterisk (bottom right). The height of the diagnosed PBLH of each model is marked on the wind speed profile (top left). The enlarged markers in the hodograph (bottom left) indicate the wind velocity at the lowest level (50 m) and each subsequent marker indicate wind velocity at the next higher level in 50 m interval.

between 50 to 100 m followed by an layer of weaker stability approximately 400 m thick and capped off by a layer that is almost statically neutral. All the simulations exhibited cold bias, especially aloft compared to the sounding. Unlike the previous sounding (Fig.3.15), the simulations are unable to capture features of this temperature sounding. All the simulations that used the MYNN scheme though featuring three distinct layers of varying stability but do not represent the observed conditions well. All four simulations, ERA_MYNN, NARR_MYNN, ENT_MYNN, GFS_MYNN and GFS_SST, are able to simulate a stable layer near surface but it

is deeper than observed and less stable. Above this stable layer, instead of having a less stable layer, these simulations simulated a almost neutral layer of varying thickness, depending on the choice of the IBC. The final layer of these simulations is a statically stable layer that is more stable than observed. Curiously, the ERA_YSU model shows an almost single stable layer that is of the same strength as the intermediate layer in the observed profile. This behavior can be attributed to the differences between the PBL scheme and likely due to the stronger mixing in MYNN than YSU during this regime.

The differences in the temperature profiles are also reflected in the wind speed profiles (top right panel of Fig. 3.16). The observed wind profile features 2 local maxima at the interface between the layers of different stability. The simulations' wind profiles reflected the inability of the simulation to capture the appropriate temperature profiles. The modelled wind profiles show a shear layer near the surface till the top of the surface inversion and within the near-neutral layer, the wind shear rapidly decrease. The surface-based shear layer extends to the approximately the height of the inversion, with the exception for ERA_YSU, which does not have a distinct surface-based inversion. The wind veering aloft is also not predicted well. The observation indicate veering till height of the second maximum and then backing aloft. All the simulations are able to capture the veering within the surface shear layer but are, predictably, unable to feature the appropriate wind backing aloft. ERA simulations and NARR_MYNN showed elements of backing aloft but at heights higher than observed.

Lastly, I compare the results to the final available sounding (Fig. 3.17) during the period of this study, on 31 July 2004 2300 UTC. Similar to the earlier profiles, there is a strong, shallow stable layer at the surface. The intermediate layer of less stability is now shallower compared to the previous sounding and an almost statically neutral layer caps the top of this observed profile. The simulations all showed a cold bias of about 3 K within the first 600 m, the surface inversion in the simulations are all deeper than observed with the ERA_YSU being the deepest at approximately 200 m. The stability of the layer aloft is overestimated in all the simulations and the temperature gradient tends to greater stability with increasing height.

A strong jet profile is seen in the observed wind sounding with a maximum of

Figure 3.17: Profile on 31 July 2004 2300 UTC (1800 LST) for wind speed (top left), virtual potential temperature (top right), hodograph (bottom left) and map of domain with position of sounding marked with an asterisk (bottom right). The height of the diagnosed PBLH of each model is marked on the wind speed profile (top left). The enlarged markers in the hodograph (bottom left) indicate the wind velocity at the lowest level (50 m) and each subsequent marker indicate wind velocity at the next higher level in 50 m interval.

19 m s^{-1} at 150 m before decreasing aloft and increasing to another maximum of 18 m s^{-1} at 800 m. All the simulations overestimated the strength and height of the LLJ. ERA_MYNN severely overestimated the height of the LLJ at 500 m. All the simulations are able to model, to varying degree, the decrease and increase in wind speeds aloft. The observed winds at the surface are south-southwesterly and veers southwesterly aloft. All the simulations replicated this feature quite well, although simulated winds near the surface are slightly more southerly when compared to the observed winds.

In summary, the simulations tend to exhibit cold biases, similar to surface temperature comparisons on land. The modelled temperature profiles are most dissimilar on 31 July 2004 1130 UTC, the choice of IBC greatly affects the prediction of the boundary layer. GFS simulations show a much deeper and developed layer above the surface inversion compared to ERA and NARR simulations. The choice of PBL scheme is also a factor, the YSU scheme predicts a stable layer that does not have a developing layer aloft compared to the MYNN scheme. The simulations fail to predict a LLJ structure on Day 1 and overestimated the strength and height of the LLJ on Day 2. Based on these 3 comparisons, it is difficult to objectively decide which model configuration is able to simulate the varying conditions observed over the 2-day period of study, although the GFS_MYNN seems to be performing the best. Hence, in the following subsection, I utilize the high resolution HRDL to guide the model validation process.

3.3.4 Comparison to HRDL profiles

I now apply the diagnostic tool to analyse the simulated horizontal wind velocity against HRDL profiles within the first 500 m. Unlike the ASOS surface wind speed analysis in the previous section, I analyse the horizontal wind as a vector quantity instead of a scalar quantity. As outlined in (Koh et al. 2012), this gives further insight into the simulation of the LLJ. I also included the familiar mean bias in wind speed in this analysis to provide a more physical meaning to the bias interpretation. I perform the analysis at each 10 m interval for a specified time period to obtain the diagnostic profiles shown in Fig. 3.18 and subsequently in Fig. 3.19 and Fig. 3.20.

I evaluate the performance of the simulations for the full duration from 30 July 2004 0000 UTC (1900 LST) to 01 August 2004 0000 UTC (1900 LST). It is clear from Fig. 3.18 that all the simulations under-predict the mean wind speed especially within the LLJ layer. The maximum mean wind speed biases occur within the 50 m to 100 m height, the LLJ core height layer. GFS simulations have less bias compared to NARR and ERA simulations and have almost no bias aloft ($> 300m$). Comparing ERA_MYNN with ERA_YSU, the effect of the PBL choice is seen within the 50 m to 400 m. the use of the YSU scheme reduced the bias by approximately 0.2 to 0.4 $m s^{-1}$ within this layer however, the improvements are not as significant compared to the choice of IBC. When the skin temperature of ERA_MYNN is changed to that of

Figure 3.18: Diagnostic metrics profile: mean wind speed bias (top left), NBIAS (top right), ρ (middle left), η (middle right), α (bottom left) and NRMSE (bottom right). Evaluated for FULL time period from 30 July 2004 0000 UTC (1900 LST) to 01 August 2004 0000 UTC (1900 LST)

NARR_MYNN, ENT_MYNN wind speed bias above the surface remains relatively unchanged (top left panel of Fig. 3.18), ENT_MYNN bias profile better resembles ERA_MYNN compared to NARR_MYNN above 40 m. This trend is similar when comparing the bias profiles of GFS_MYNN and GFS_SST, the effect of warming the sea surface by 2 K did not significantly affect the wind speed biases either.

When I examine the NBIAS metric, I see similar results whereby the largest NBIAS are within the LLJ layer between 50 m to 100 m and that GFS simulations performed the best. However, the NBIAS metric (for vector variables is always positive) offers an interesting insight to the behavior of the simulated winds that is not apparent in the mean wind speed bias plot. The GFS simulations have negligible biases aloft but has similar NBIAS compared to the other simulations. This implies that although the average wind speeds are well predicted in GFS simulations but the directions are not well predicted, resulting in non-zero NBIAS profiles.

I compare the skill in variance prediction and find that all the simulations performed similarly well. I find that all the simulations have relatively poor ρ within the first 100 m especially within the first 40 m. However, ρ improved rapidly aloft. The poor correlation points to poor phase agreement within this layer. Aloft, the correlation metric shows little differences between simulations with the same IBC with little influence from the choice of PBL scheme as well as skin temperature. This is expected since these configurations should have little impact on the temporal variation of wind velocities. As I look to the η metric, all the simulations performed well in predicting the amplitude of the variation in winds with $\eta \approx 1$ within the first 100 m. This points to overall good prediction of variance amplitude even within the LLJ layer. Due to overall skill in predicting variance amplitude and phase, all simulations scored well on the combined α metric, with a maximum of $\alpha \approx 0.3$ within the first 50 m.

For the overall error metric, NRMSE which combines both bias and variance error prediction, all the simulations performed similarly well. The simulations have the highest NRMSE within the first 100 m and improved performance aloft. GFS simulations performed the best within the LLJ layer compared to the other simulations; NARR_MYNN outperforms both GFS simulations slightly aloft however taking into account the behavior near the surface, GFS is clearly the choice model. I see some subtle differences in NRMSE within the first 40 m between ERA_MYNN

and ERA_YSU, because of better NBIAS scores by ERA_MYNN in this layer. It is difficult to isolate which aspect of the MYNN scheme is contributing to this slightly better performance. However, the overall differences in NRMSE is minor compared to the choice of using GFS IBC.

Similar to the diagnosis of the ASOS data, I separately analyse the performance of the simulations in each of the days. The results are plotted in Fig. 3.19 and Fig. 3.20 for Day 1 and 2 respectively.

From the mean wind speed bias plot for Day 1, I find that all simulations underestimated the wind speeds Fig. 3.19 (top left). The maximum underestimation at around 50 m, the average height of the LLJ on Day 1. On Day 2, I see similar underestimation profiles within the LLJ layer in Fig. 3.20 (top left). However, the simulations performed differently aloft. I see that ERA- and NARR-based simulations have little mean wind speed bias while GFS-based simulations overestimated the wind speed by an average of 2 m s^{-1} . Also, it is clear that GFS-based simulations underestimated and overestimated the winds aloft on Day 1 and 2 respectively that resulted in 0 bias aloft for the FULL duration. As pointed out earlier, the mean wind speed bias plots do not account well for wind vector biases. Hence the NBIAS profiles are analyzed to further evaluate the simulations performance.

GFS simulations, having a consistently lower NBIAS throughout the evaluated height, performed the best for Day 1 compared to the other simulations. The effect of warming the SST is not shown when comparing the profiles of GFS_MYNN and GFS_SST. This is similarly reflected in comparing the profiles of ERA_MYNN and ENT_MYNN. The choice of PBL did not significantly improve the NBIAS of ERA_YSU compared to ERA_MYNN despite giving a smaller mean wind speed bias aloft.

For the NBIAS evaluation on Day 2, the effect of non-IBC choices becomes much more apparent. All the simulations performed similarly within the first 80 m, with low NBIAS ≈ 0.3 at the surface and increasing rapidly to NBIAS ≈ 1.5 above 80 m, the average height of observed LLJ over this duration. The similarly low NBIAS within the first 40 m indicates that the simulations biases likely do not originate from IBC conditions and it is a function of surface-based parameterization scheme.

The NBIAS profiles significant diverge above 80 m. The divergence due to IBC is within expectation, however the poor performance of GFS simulations aloft is not.

Figure 3.19: Diagnostic metrics profile: mean wind speed bias (top left), NBIAS (top right), ρ (middle left), η (middle right), α (bottom left) and NRMSE (bottom right). Evaluated for WEAK time period from 30 July 2004 0000 UTC (1900 LST) to 31 July 2004 0000 UTC (1900 LST)

GFS simulations performed much better in Day 1 and when evaluated throughout the FULL duration. The YSU scheme is shown to perform better aloft than the MYNN scheme on this day when comparing ERA_MYNN against ERA_YSU.

The effect of changing SST is now apparent on Day 2. Comparing GFS_MYNN against GFS_SST, it can be seen that warming the SST by 2 K resulted in difference biases aloft. The maximum in NBIAS is stronger and occurs at a lower height in GFS_SST. At the same height for the mean wind speed bias, there is little difference between GFS_MYNN and GFS_SST, which implies that the warmer sea surface had an effect on the wind directions aloft resulting in differences in NBIAS. Similarly when ERA_MYNN, ENT_MYNN and NARR_MYNN are compared, the NBIAS profiles ENT_MYNN tracks closely with NARR_MYNN and is very different compared to ERA_MYNN. This implies that the sea surface temperature is a strong influence in the NBIAS metric during Day 2.

Next, I evaluate the skill in variance prediction for the simulations on each of the 2 days. I find that all the simulations performed well on Day 1, with the α score between 0.1 and 0.3 throughout the height. However, when I examine ρ and η , the components of α , I find that the source of each simulations' strength lies in different aspect of variance prediction. For ERA simulations (ERA_MYNN and ERA_YSU), they score higher in ρ and η compared to the other simulations near the surface however their performance degraded aloft. NARR on the other hand, has poor η scores compared to the other simulations but is able to make up for that with good ρ scores, especially aloft. GFS performed the worst near the surface for ρ but is able to make up with good η scores throughout the profile. When I compare the performance of PBL schemes, I find that the YSU scheme confers an advantage in variance prediction compared to MYNN. ERA_YSU consistently shows better correlation and variance similarity compared to ERA_MYNN in Day 1. I also do not find an appreciable difference in variance prediction due to differences in SST.

Unlike the NBIAS analysis for Day 2, the simulations performance in variance prediction did not diverge significantly on Day 2 compared to Day 1. I find that all the simulations performed similarly overall. The maximum error in variance prediction occurs near the surface for all the simulations at around 20 m before decreasing aloft. I also note that α aloft for all simulations are similar in magnitude on both Days 1 and 2. Looking to the ρ and η metric, I see that all the simulations have poor

Figure 3.20: Diagnostic metrics profile: mean wind speed bias (top left), NBIAS (top right), ρ (middle left), η (middle right), α (bottom left) and NRMSE (bottom right). Evaluated for STRONG time period from 31 July 2004 0000 UTC (1900 LST) to 01 August 2004 0000 UTC (1900 LST)

correlation near the surface ($\rho < 0.5$) and relatively poor η and that contributed to the overall poor variance prediction. The η metric, which measures the error in variance amplitude prediction, shows similar trend near surface for all the simulations. It decreases sharply till 30 m before increasing sharply and this behavior is not seen in Day 1, where near surface wind variance amplitude is well predicted. I find that the simulations under-predict the variance in winds near the surface by approximately 3 times. The performance of the MYNN and YSU schemes are similar in this surface layer. The effect of different SST is represented in η as well. GFS_SST, having a higher SST, has higher variance compared to GFS_MYNN in this layer. While ENT_MYNN is closer in magnitude to NARR_MYNN compared to ERA_MYNN. This likely points to poor turbulence representation near the surface on Day 2 due to surface-based parameterization.

Overall, using the NRMSE metric, I find that the simulations performed better on Day 1 than on Day 2. The effect of IBC is more dominant on Day 1, resulting in NRMSE profiles that are similar within each class of IBC. The effects of PBL scheme and SST are less discernible. For Day 2, the choice of IBC does influence the overall NRMSE, however the profiles of all the simulations are similar to each other in terms of shape and magnitude, especially aloft. This indicates that either the IBCs are more similar in skill on Day 2 or that the dynamical parameterization schemes are able to compensate for differences in mesoscale conditions in IBCs resulting in similarly skilled forecasts. Further and indepth analysis of mesoscale conditions of IBC is outside the scope of this present study.

I pick GFS_MYNN as the most optimal model configuration that will be used in the later LES study. It is the best performing model over the FULL duration evaluation. Although NARR_MYNN outperforms GFS_MYNN in some aspects, such as smaller NBIAS aloft on Day 2, smaller NRMSE aloft on Day 2, GFS_MYNN is the overall better model especially within the first 100 m on both days.

3.4 Summary and Discussion

I simulated two LLJ events observed over two days during the NEAQS-ITCT 2004 campaign and aimed to characterize the variability in the LLJ timing, height and speed of the LLJs due to the different model configurations. I varied the choice of

IBC dataset, PBL schemes and SST.

The simulations behave similarly and are able to simulate most of the prominent features of the LLJs observed. The phase of the LLJs is poorly predicted. All the modelled weak LLJ leads the observed LLJ while all the modelled strong LLJ lags the observed LLJ. The heights of both LLJs are overpredicted while the peak speeds of the weak (strong) LLJ is underpredicted (overpredicted), resulting in underprediction of the average wind shear in the subjet layer. The overprediction in LLJ heights might be improved with even finer vertical resolution than used in this study (~ 15 m within the lowest 150 m), however, the mesoscale simulation will be used for the WRF-LES nested study and increasing the vertical resolution will be computationally prohibitive.

Analysis of the ASOS surface data showed average cold biases (~ 2 K) in the surface temperature, with no distinguishable spatial pattern of this bias. None of the simulations distinguished themselves in terms of overall NPE and NRMSE scores. This result is indicative of the performance of the surface layer model used and shared by all the simulations.

The cold bias observed in the surface temperature is also extended aloft, when the modelled temperature profiles are compared to available rawinsonde profiles. Although the simulations are able to replicate the wind veer relatively well, the wind directions tends to be too southerly, indicating that winds in the simulations travelled a longer fetch distance across the sea surface compared to the observed winds. It is plausible that with the increase in travel time over the cool marine boundary layer, the daytime land-based convective turbulence decays further, leading to increased acceleration of the winds aloft (Fig. 3.17).

The differences due to PBL scheme and SST forcing are much more prominent in the soundings for Day 2 than Day 1. This can largely be attributed to the less stable marine boundary layer in the simulations. The second sounding profile (Fig. 3.16) illuminated the differences in PBL scheme. The layer aloft the weakly stable marine layer inversion is well mixed in simulations using MYNN scheme, likely due to quick development of a convective layer on land just after sunrise that got advected offshore. ERA_YSU which uses the YSU scheme, instead showed a stable layer at similar heights. The differences in PBL schemes, particularly in PBL height prediction, are discussed later.

Analysis of the modelled wind profiles against HRDL shows overall good prediction by all simulations. All model configurations show similar error behavior for all the diagnostic metrics used. However, the simulations behave relatively poorer within the LLJ layer (< 150 m) compared to aloft. This is due to a combination of strong NBIAS and poor ρ for the wind velocity prediction.

The average NBIAS is strongest at 100 m height for all the simulations. On average, the underestimated wind speed in this layer is due in part to the simulations overpredicting the height of the LLJ core. GFS_MYNN showed the least mean bias and NBIAS for this evaluation. The relatively poor performance of $\rho = 0.7$ for all simulations near the surface indicates poor phase prediction of the wind velocity variance. This performance improves rapidly aloft, above 100 m. This is an indication at the failure of surface parameterization related to poor stability prediction as seen in temperature profiles.

Decomposing the evaluation of the WRF simulations into 2 separate 24-h period, I find varying behavior between the different model configurations. The choice of IBC is much more significant on Day 1 as compared to Day 2. GFS simulations consistent have a lower NBIAS compared to other simulations on Day 1. Variable performance for variance prediction for the simulations

On Day 2, all the simulations performed similarly overall, with little dependence on model configurations in the subjet layer. The contrast in model performance between Day 1 and Day 2 can be attributed to the different mesoscale conditions. Day 1 encompasses various transitions, a weak LLJ decaying into a quiescent period followed by a strong LLJ initiation, while a strong LLJ condition persists throughout Day 2. This indicates that synoptic conditions are a strong driver of such transition periods. However, once the LLJ has been initiated, the WRF model does a comparable job of modelling aspects of the LLJ dynamics. IBC controls the biases further aloft while biases in the subjet layer are similar across model configurations. The subjet layer is driven by model dynamics despite IBC differences, the differences in synoptic conditions manifests in the free atmosphere aloft. The LLJ on Day 2 is an advection based phenomenon similar to Smedman et al. (1995), the differences in wind velocities aloft is likely related to differences in IBC conditions such as the position and strength of the high pressure system over the North Atlantic.

Chapter 4

Nested WRF-LES Real Simulations of Coastal Low-Level Jets - An NEAQS-ITCT 2004 Case Study

4.1 Introduction

Since the pioneering work of Deardorff (1970), large-eddy simulation (LES) has been used to study many fundamental aspects of the atmospheric boundary, such as the convective boundary layer (Moeng and Wyngaard 1988; Mason 1989; Nieuwstadt et al. 1993; Khanna and Brasseur 1997), neutral boundary layer (Moeng and Sullivan 1994; Stoll and Porté-Agel 2006) and the stable boundary layer (Mason and Derbyshire 1990; Kosovic and Curry 2000; Basu et al. 2006; Beare et al. 2006; Holtslag 2006). In more recent years, there is an increasing interest in using LES for wind turbine and wind farm modelling Churchfield et al. (2010); Porté-Agel et al. (2011); Churchfield et al. (2012a); Yang et al. (2014) and have yielded invaluable insights into turbine wake interactions (Aitken et al. 2014; Dörenkämper et al. 2015; Mirocha et al. 2015) and wind farm alignment (Stevens et al. 2014). However, these studies used mainly idealized initial and/or boundary conditions with simplified atmospheric physics that do not fully take into account synoptic nor mesoscale processes.

The capability to couple LES to real regional weather conditions is particularly relevant to wind energy forecasting. Multi-scale models have been proposed to fulfil this need and have had some success (Liu et al. 2011; Talbot et al. 2012; Zhou and Chow 2014). One of the chief challenges facing multi-scale modelling is the transition from modelled turbulence in mesoscale domain to 3D resolved turbulence within LES domain. In this chapter, I address this challenge through use of a turbulence spin-up method within the LES domain, known as Cell-Perturbation Method (Muñoz-Esparza et al. 2015). The high temporal and vertical resolution of the HRDL profiles from the NEAQS-ITCT 2004 case study provides us the unique opportunity to validate real nested LES results of two low-level jet (LLJ) episodes. I address the following questions:

1. Are the predicted height of the LLJ core and the strength of the LLJ core improved through the use of LES?
2. Does the use of the Cell-Perturbation Method improve LES performance in the LLJ predictions?
3. How robustly does the Cell-Perturbation Method perform under various different LLJ conditions?

In Section 2, I describe the LLJ cases selection, LES domain numerical set-up and the parameters used the Cell-Perturbation Method. In Section 3, I present the results of the simulations and evaluate the performance of the simulations against HRDL profiles and the spectral analysis of turbulence. Lastly, I conclude with a discussion on the performance of LES and the use of the Cell-Perturbation Method in the two selected LLJ cases.

4.1.1 Cell-Perturbation Method

There are multiple methods that have been proposed to aid in turbulence spin-up within a LES domain. Four such methods will be briefly described in this section. The simplest method is known as the precursor method where an auxiliary computation is first run using periodic lateral boundary conditions, then imposing the solution onto the domain of interest. The imposed flow will contain turbulence that satisfies the equations of motion thus reducing a need for fetch distances. However,

this method can introduce artificial periodicity and would be computationally intensive due to the need of an auxiliary computation. Moreover, running such auxiliary computations to spin up turbulence cannot be done for real case studies.

The cyclic method, also known as rescaling and recycling method, serves to alleviate some of the issues associated with the prepared method. This method creates a buffer zone at the inflow boundary. The solution along a vertical plane some distance from the inlet is extracted and then reimposed at the inlet, creating this buffer zone. The flow is thereby being recycled in this zone, allowing turbulence to develop fully (Churchfield et al. 2012a). This reduces the computational costs and fetch distances normally required. However, this method is not practical for real atmospheric flows where flow direction changes in time, requiring multiple buffer zones to be set-up. The cyclic method also suffers from periodicity contamination that arise from the recycling period/distance used.

Synthetic turbulence methods generate a synthetic turbulent velocity field using information using the flow condition under study. The method relies on digital filter to generate turbulence. This method is attractive when only limited turbulence statistical data of the flow regime is available. This method is found to be highly efficient and has been used to simulate atmospheric boundary layer flow. Specific implementation of various digital filters can be found in (Klein et al. 2003; Batten et al. 2004; Di Mare et al. 2006; Xie and Castro 2008)

The common disadvantage of the previous methods is a need for *a priori* turbulence information about the flow, which are largely unknown. Mirocha et al. (2014b) aimed to address this shortcoming by introducing a new approach to generate turbulence. Their approach is based on adding sinusoidal perturbations to the temperature and horizontal momentum fields at the inlet. They showed that turbulence spin-up was easily achieved with dynamic models. Muñoz-Esparza et al. (2014) expanded on their approach and demonstrated that the Cell-Perturbation Method (CPM) gave the most efficient and effective performance. The CPM is based on imposing stochastic perturbations in the potential temperature field. The perturbations are imposed in square cells of variable sizes (cells are a group of grid points). The amplitude and frequency of perturbation was initially determined empirically. Muñoz-Esparza et al. (2015) developed further guidance on the use of the CPM. The Eckert number, a ratio of the flow's kinetic energy to enthalpy applied by

the perturbations, was proposed as a suitable parameter that governs the transition of turbulence. $E_c \approx 0.2$ proved to be the optimal number for convective, neutral and weakly stable flow.

4.2 Domain Description and Simulation Set-up

4.2.1 Case Selection

In Chapter 3, I set the basic criteria for the LLJ to be simulated in this LES study. The time period should:

1. feature different LLJ events;
2. have continuous HRDL profiles of at least 2 hours in duration featuring the LLJs;
3. have continuous HRDL profiles up to 500 m in height during the 2-hour period;
4. have no major synoptic events such as fronts within the domain.

The location of the HRDL profiles changes in time, */sim* 3-4 km between each profile, hence it would require a large LES domain to cover the geographical span of the entire HRDL track during this 48-h period. Moreover, it will be computational expensive to run a LES domain for 48 hours. Thus, a 2-h period for each of the two LLJ events are specified, as a compromise between computational restrictions and to have as long a timespan for reasonable analysis. I identified two 2-h HRDL periods for the LES study and they are 30 July 2004 0245 UTC to 0445 UTC for the WEAK LLJ and 31 July 2004 2045 UTC to 2245 UTC for the STRONG LLJ. These identified periods will also guide the design of the LES domains in the following subsection.

4.2.2 WRF Model Domain Configuration

The domain consists of two nests, an outer mesoscale nest (d04) and an inner LES nest (d05) with horizontal resolutions of 250 m and 50 m respectively. The domain sizes ($N_x \times N_y \times N_z$) are 300x300x60 and 600x800x60 respectively. The 60 vertical

model levels are concentrated mainly within the lower boundary layer and progressively stretched upwards. The average vertical resolution within the lowest 150 m is ~ 15 m and within the lowest 1000 m is ~ 35 m. The time-steps for d04 and d05 are 2.5 s and 0.5 s respectively. The inner LES domain is designed to encompass the positions of the ship over the identified 2-h period. The ship track is confined within the top half of d05 as the winds are predominantly southerly and south-south-westerly in both WEAK and STRONG periods. The outer mesoscale domain is designed such that the LES domain is kept within the north-eastern region of the domain. The distance of d05 from the southern and western boundary of d04 is kept the same for both WEAK and STRONG. The initial and boundary conditions for d04 are obtained from domain d03 of GFS_MYNN using the ndown procedure in WRFV3.6.1 (Skamarock et al. 2008) at 15-minute intervals. Both Figs. 4.1 and 4.1 show the domain set-up for both WEAK and STRONG periods within d03. Table 4.1 summarizes the model configurations used in this study. Table 4.2 provides an estimate of the computational load required for each simulation run.

4.2.3 Cell-Perturbation Method Setup

Muñoz-Esparza et al. (2015) provided guidance on optimal CPM set-up for effective turbulence spin-up. They identified Eckert number, $Ec \approx 0.2$ as the governing parameter for introducing turbulence in an ideal neutrally stratified boundary layer. I use that to guide the study and set the required CPM parameters, There are three parameters to be determined, they are amplitude of the θ , potential temperature perturbation, height of θ perturbation to be applied and time scale of perturbation.

The amplitude of the θ , perturbation is determined using the following equation,

$$Ec = \frac{U_g^2}{c_p \theta_{CPM}}, \quad (4.1)$$

where U_g is the geostrophic forcing wind speed, θ_{CPM} is the maximum potential temperature perturbation amplitude applied, $c_p = 1004.6 \text{ J kg}^{-1} \text{ K}^{-1}$ is the specific heat capacity at constant pressure. In this case-study, U_g is defined as the wind speed (U_{PBLH}) at height of the planetary boundary layer (PBLH), as determined by the PBL scheme in the WRF model. I average U_{PBLH} along the southern boundary of the LES domain and determine θ_{CPM} when $Ec \approx 0.2$.

Figure 4.1: WRF domain set-up showing 2 nested domains for WEAK LLJ time period within d03 (1250 m), the outermost domain generated from mesoscale case-study in Chapter 3. D04 (250 m) is the intermediate mesoscale domain. D05 (50 m) is the inner-most LES domain nested within d04. Black markers indicate the HRDL track between 30 July 2004 0245 UTC (2145 LST) to 30 July 2004 0445 UTC (2345 LST)

Figure 4.2: WRF domain set-up showing 2 nested domains for STRONG LLJ time period within d03 (1250 m), the outermost domain generated from mesoscale case-study in Chapter 3. D04 (250 m) is the intermediate mesoscale domain. D05 (50 m) is the inner-most LES domain nested within d04. Black markers indicate the HRDL track between 31 July 2004 2045 UTC (1545 LST) to 31 July 2004 2245 UTC (1745 LST).

Table 4.1: Names and summary of all the model configurations in this chapter

	HRDL	Dura-	Simulation	Du-	Cell-
	tion		ration		Perturbation
					Method
WEAK_NO	30 July 2004 0245 UTC to 0445 UTC	30 July 2004 0200 UTC to 0600 UTC	No		
WEAK_CP	30 July 2004 0245 UTC to 0445 UTC	30 July 2004 0200 UTC to 0600 UTC	Yes		
STRONG_NO	31 July 2004 2045 UTC to 2245 UTC	31 July 2004 2000 UTC to 0000 UTC	No		
STRONG_CP	31 July 2004 2045 UTC to 2245 UTC	31 July 2004 2000 UTC to 0000 UTC	Yes		

Maximum height till which θ perturbation is applied is at 2/3 of the PBLH to prevent unrealistic perturbations at the inversion layer (Muñoz-Esparza et al. 2015). I then determine the closest eta level that corresponds to 2/3 of the average PBLH along the southern boundary.

Lastly, the time-scale for the perturbation is determined using the following equation,

$$\Gamma = \frac{t_p U_1}{d_c}, \quad (4.2)$$

where Γ is a dimensionless time scale, t_p is the perturbation time period, U_1 is the wind speed at the first vertical point perturbed and $d_c \approx 70$ m is the diagonal distance of the grid resolution. $\Gamma = 1$ is recommended by Muñoz-Esparza et al. (2015). In this study, $\Gamma = 0.5$ was used instead in order to perturb at larger mesoscale wavelengths. The first perturbed vertical point is the third vertical point (z3) from the surface instead of the first vertical point (z1). This prevents the imposed potential temperature perturbations from interfering with the surface layer

Table 4.2: Summary of computational load required for each simulation run.

	Domain sizes (Nx x Ny). All domains Nz at 60	CPU-Hours per Simulation- Hour	Total CPU- Hours (including spin- up)
D01, D02, D03	100x100, 300x300, 300x300	40	2400
D04, D05	300x300, 600x800	1200	8640

parameterization and avoids modifications of the surface stability.

The CPM parameters used in this study are summarized in Table 4.3.

	WEAK	STRONG
Amplitude (K)	0.21	1.80
U_{PBLH} (m s^{-1})	6.50	19
PBLH (m)	80	185
Z_{PBLH} (level)	4	9
Time step iteration	11	5
U_{z3} (m s^{-1})	6.36	14

Table 4.3: Summary of CPM parameters used for both WEAK and STRONG case.

4.2.4 PBLH in WRF

In this section, I elaborate on the formulation of the planetary-boundary layer height (PBLH) in MYNN and YSU as implemented in WRF. The role of PBLH in each scheme differs and exists as a diagnosed quantity. The MYNN scheme (Nakanishi and Niino 2004) uses PBLH to determine the mixing length scales for formulation of the eddy diffusivity model. It determines PBLH based on turbulent kinetic energy

(TKE) when the first estimate is below 500 m. In YSU scheme (Hong 2010), PBLH is used to separate the regime between the mixed layer and the free atmosphere diffusion. The formulation for diffusion defers between the mixed layer and the free atmosphere. PBLH is also a proxy for diagnosed stability of the atmosphere. The lower the PBLH, the more stable is the atmosphere in the domain and hence weaker turbulence.

Determination of PBLH in MYNN

MYNN scheme uses a hybrid method to determine the PBLH. It estimates the PBLH based on the profiles of two quantities: potential temperature and TKE. The potential temperature method (Nielsen-Gammon et al. 2008) defines a $PBLH_\theta$ as the height where the virtual potential temperature first exceeds the minimum virtual potential temperature by $\Delta\theta_v$ and is expressed as

$$PBLH_\theta = z(\theta_{v,min} + \Delta\theta_v), \quad (4.3)$$

where z is height, $\theta_{v,min}$ is the minimum virtual potential temperature within the first 500 m, $\Delta\theta_v = 0.75$ for sea surface and 1.25 for land surface.

The potential temperature method does not account for profiles with LLJ, hence a second height based on TKE is used in the MYNN scheme. This method is based on Banta et al. (2006) and Pichugina et al. (2008) where they showed that minimum TKE height corresponds to max LLJ height and this can be used as a measure for PBLH in stable atmosphere featuring a jet. The $PBLH_{TKE}$ is defined as the height where the TKE decays to at least ϵ_{TKE} . ϵ_{TKE} is defined as 5% of the maximum TKE or $0.025 \text{ m}^2\text{s}^{-2}$, whichever is greater. This relation is expressed as

$$PBLH_{TKE} = z(\epsilon_{TKE}), \quad (4.4)$$

where $\epsilon_{TKE} = \max(0.05TKE_{max}, 0.025)$.

MYNN scheme then uses a weighted tanh function to blend both heights to obtain the PBLH. The weighting function, w , is defined as

$$w = 0.5 \tanh\left(\frac{PBLH_\theta - SBL_{lim}}{SBL_{damp}}\right) + 0.5, \quad (4.5)$$

where $SBL_{lim} = 200$ m is set to the typical scale of a stable boundary layer and $SBL_{damp} = 400$ m is set to the transition length for blending between a stable boundary layer and free atmosphere. The final PBLH is weighted as such

$$PBLH = (1 - w)PBLH_{TKE} + wPBLH_{\theta}. \quad (4.6)$$

However, the diagnosed PBLH as shown in some of the figures do not correspond to the height of the LLJ and frequently overpredicts the height. This is related to the limits imposed on the PBLH calculation within the MYNN scheme.

$PBLH_{TKE}$ is limited by $PBLH_{\theta}$, if $PBLH_{TKE}$ is greater than $PBLH_{\theta} + 500$ m, $PBLH_{TKE}$ is set to $PBLH_{\theta} + 500$ m. This limit is imposed such that if strong turbulence exists aloft, $PBLH_{TKE}$ will not dominate and produce unrealistically large PBLH. In this case, there exists time periods when there exists strong turbulence aloft due to advection that result in a local TKE maximum aloft, while being lesser than TKE at the surface. The local minimum that is now between the aloft and surface maxima, might not satisfy the 5% of maximum TKE as required by MYNN to define $PBLH_{TKE}$ at this height. Hence, the $PBLH_{TKE}$ will be defined above the maximum aloft, thereby overestimating the height of the LLJ.

Determination of PBLH in YSU

The YSU scheme determines PBLH in a different manner, it relies on Bulk Richardson number. YSU first estimates the PBLH as the height where Bulk Richardson number exceeds Critical Bulk Richardson number of 0.0. For a stable boundary layer, if the PBLH is less than the z_1 , the height of the first vertical grid point above the surface, the Critical Bulk Richardson number is revised to 0.25 for land surface and for water surface, the revised Critical Bulk Richardson number is calculated as

$$Ri_{B,crit} = \min(0.16(Ro \times 10^{-7})^{-0.18}, 0.3), \quad (4.7)$$

where Ro is the Rossby number based on the wind speed at 10m, defined as

$$Ro = \frac{U_{10}}{Lf}, \quad (4.8)$$

where L is the roughness length associated with the surface and f is the Coriolis parameter $\approx 1 \times 10^{-4}$.

Then a new PBLH is estimated using this revised Critical Bulk Richardson number. PBLH in the YSU scheme separates the boundary layer into 2 regions, below and above the PBLH, where diffusion coefficients are formulated differently. Below the PBLH, the diffusion coefficients for momentum and heat, K_m and K_h are calculated based on stability function approximations and surface characteristic velocity. Above the PBLH, the diffusion coefficients are based on shear and gradient Richardson number. The interested reader can refer to Hong et al. (2006) and Hong (2010) for further details on the YSU formulation.

Implication of PBLH

The diagnosed PBLH should reflect the height of the boundary layer, regardless of the stability conditions. However, as seen in Figs. 3.16 and 3.17, the diagnosed PBLH can be erroneous. The PBLH is too high compared to the height of the LLJ core, which is often taken to be height of the boundary layer. While this error does not impact the choice of model configurations, as the PBL schemes do not defer in overall performance shown in Fig. 3.18, it is relevant in the subsequent nested LES study. PBLH will be used to determine the vertical extent of the turbulence spin-up method.

4.3 Results

4.3.1 WEAK LLJ

4.3.1.1 Comparison to HRDL profiles

In Chapter 3, I have seen that D03 run underestimated the speed and height of the WEAK LLJ. I investigate if the use of higher resolution LES would correct these biases. I compare the performance of the LES simulations to GFS_MYNN and to HRDL during this 2 hour time period.

From Fig. 4.3 and Fig. 4.4, I see that both LES simulations, WEAK_CP and WEAK do not appear to be qualitatively different from GFS_MYNN. However, I see that the coherent smooth jet seen in the GFS_MYNN and HRDL profiles at ≈ 80 m is lacking in the LES profiles. The vertical profiles for the LES are also in general more variable. This can be attributed to the 3D turbulence mixing in LES

Figure 4.3: Time series of horizontal wind speed profile along ship track from 30 July 2004 0245 UTC (2145 LST) to 30 July 2004 0445 UTC (2345 LST)

leading to such spatial variations.

Fig. 4.5 shows the 2-h averaged profile of the wind speed for the WEAK LLJ case. The maximum LLJ speed in the LES profiles is further reduced compared to GFS_MYNN and a larger underestimate of the HRDL maximum LLJ speed observed. However, the LES profiles also show a lowered LLJ height and increased wind speed in the subjet layer. The wind shear in the subjet layer of the LES profiles is still too weak compared to the HRDL profile.

Fig. 4.6 shows the 2-h averaged profile of the wind direction for the WEAK LLJ case. The LES profiles are similar to GFS_MYNN and shows that the wind directions are still poorly represented. The LES profiles are too southerly below 250 m and too westerly above 250 m. The averaged wind veer throughout the 500 m is overestimated in the LES models at $\approx 70^\circ$ compared to $\approx 50^\circ$ observed in the

Figure 4.4: Time series of horizontal wind direction profile along ship track from 30 July 2004 0245 UTC (2145 LST) to 30 July 2004 0445 UTC (2345 LST)

HRDL profile.

I use the same set of diagnostic metrics in Chapter 3 to evaluate the performance of the models. From the BIAS and NBIAS profile, it is clear that the LES runs are able to reduce the BIAS in wind speed and NBIAS in wind velocity in the LLJ layer below 50 m compared to GFS_MYNN. Above this height to 150 m, the BIAS of the LES profiles increased compared to GFS_MYNN, this is attributed to overestimate in LLJ nose height in GFS_MYNN as shown in Fig. 4.5. The increase in wind speed in GFS_MYNN at this height, compared to the decreasing wind speeds aloft the LES simulated LLJ nose resulted in this perceived degradation in wind speed performance. The NBIAS profiles, a measure of significance in bias, corroborates with this analysis. The NBIAS profiles between 50 m and 150 m are similar in magnitude for GFS_MYNN and LES simulations. Conversely, above 150 m, the

Figure 4.5: Averaged wind speed profile for WEAK time period from 30 July 2004 0245 UTC (2145 LST) to 30 July 2004 0445 UTC (2345 LST)

Figure 4.6: Averaged wind direction profile for WEAK time period from 30 July 2004 0245 UTC (2145 LST) to 30 July 2004 0445 UTC (2345 LST)

BIAS of GFS_MYNN and the LES profiles are similar but NBIAS shows that the LES are not performing as well.

The LES runs are able to reduce both BIAS and NBIAS within the LLJ region (< 100 m) but do not perform as well aloft, compared to GFS_MYNN. There is little difference in BIAS/NBIAS performance between WEAK_CP and WEAK_NO.

The skill in variance prediction does diverge amongst the models. Both LES have degraded performance in correlation, ρ compared to GFS_MYNN within the first 150 m. WEAK_CP is the worst performing model in this layer. Poor ρ can be attributed to the additional variance introduced by increased turbulence within the LLJ layer in the LES models. This is evident in the η metric, which measures the amplitude of the variance prediction.

GFS_MYNN underestimates the variance of the wind velocity (not shown) compared to HRDL, the improvement in η is due to increased variance that matches that of HRDL. Both LES performed better than GFS_MYNN within 150 m. Although there is an improvement in η in the LES, the variance prediction is still poor in the LLJ layer. In addition, I find that CP performed worse than NOCP within the LLJ layer despite the use of Cell-Perturbation method to spin up turbulence.

Taking into account both variance phase and amplitude prediction skill in the α metric, I find that both LES performed similar to GFS_MYNN in the first 50 m, near the surface and then diverges in performance aloft. WEAK_CP is the worst performing model between 50 m and 150 m. Above this layer (> 150 m), both LES performed better than GFS_MYNN and this is in part due to better variance prediction. The LES performed similar to each other, with little discernible impact of Cell-Perturbation.

The increasingly poor variance prediction performance aloft is in contrast to the performance evaluated over 48-h and both 24-h periods (in Chapter 3). This might suggest that the WRF has poor performance in short time period predictions but the more likely implication is that more frequent observations will be needed to fully evaluate such predictions.

Despite the poor performance of LES in variance prediction, the improvements in NBIAS compensated enough that NRMSE in LES are lesser than GFS_MYNN within the subjet layer (< 50 m). Within the LLJ nose layer, approximately between 50 m and 150 m, the poor α of LES models resulted in a decrease in

Figure 4.7: Profile for diagnostic metrics: mean wind speed bias (top left), NBIAS (top right), ρ (middle left), η (middle right), α (bottom left) and NRMSE (bottom right). Evaluated for WEAK time period from 30 July 2004 0245 UTC (2145 LST) to 30 July 2004 0445 UTC (2345 LST)).

NRMSE performance compared to GFS_MYNN. Above 300 m, aloft of the LLJ, the performances of the models are similar and have decreasing performance with increasing height, unlike in the profiles for Chapter 3 where NRMSE were steady. This is a direct result of poor α . As α tends to 1, this means that performance aloft is as good as a random forecast, as evinced by the poor correlation score. And in such cases, η is a better skill predictor as argued for by Koh et al. (2012). Koh et al. (2012) proposed an alternate metric, $\alpha^* = \alpha - \eta$ and this is similar to a metric proposed in Taylor (2000). However, this metric will not be evaluated further in this thesis and I leave this for further work.

In summary, I find that LES improved the simulated LLJ within the subject layer by decreasing the overall biases in this layer. I also find that the performance evaluated over this 2-h time period (9 profiles) differ significantly from the 48-h and 2 24-h analysis done in Chapter 4.

4.3.1.2 Spectral analysis

The instantaneous velocity fields at 30 July 2004 0245 UTC for both WEAK_CP and WEAK_NO are shown in Fig. 4.8. There appears to be little qualitative difference between WEAK_CP and WEAK_NO in turbulence spin-up. In the bottom row of Fig. 4.8 (w-velocity field), in the region between 200 to 400 ΔX and 0 to 100 ΔY , small turbulence features can be seen in WEAK_CP. However, these turbulence cannot be readily discerned with increased fetch.

For a more quantitative analysis of the effectiveness of the Cell-Perturbation method (CPM) in generating turbulence, I examine the spectra for the 3 velocity fields in both North-South and East-West directions. The spectra are calculated along the North-South (East-West) direction over a distance of 700 (500) points starting 50 points from the western (southern) boundary at every 100 points interval. The spatial spectra are Hamming windowed and averaged in time over a 4-h period with a sampling frequency of 60 s.

I first examine the East-West spectra of the horizontal velocity fields as they evolve from the South to the North. In Figs. 4.9 and 4.10, I see that there is little difference between the WEAK_CP and WEAK_NO spectra. The differences between WEAK_CP and WEAK_NO are mainly seen in the spectra near the southern boundary. The southernmost spectra for WEAK_CP at higher wavenumbers

Figure 4.8: Contours of instantaneous velocity fields at 30 July 2004 0245 UTC (2145 LST) for WEAK LLJ interpolated to 80 m. Left column: WEAK_CP. Right column: WEAK_NO. Top, middle and bottom rows are u -, v - and w - velocity fields respectively.

is more energetic than WEAK_NO within the similar range of wavenumbers. Further north into the domain, the spectra within each LES configuration tend towards self-similarity. A slight peak appears to be develop in the northernmost spectra in the $k = 10^{-2} \text{ rad m}^{-1}$ range and this peak is also seen in WEAK_NO, it is unlikely that this peak in spectral energy is due to CPM. This wavenumber ($k = 10^{-2} \text{ rad m}^{-1}$) also corresponds to a large-scale eddy wavelength $\approx 400 \text{ m}$. This length scale is close to the effective resolution of WRF ($6\Delta x$), after which the energy spectra rapidly dissipates, suggests that the horizontal resolution chosen for this stability regime is too coarse to resolve an inertial range.

Next, I see similar effects of CPM (Figs. 4.11 and 4.12) in the North-South spectra of the horizontal velocity fields in their evolution from the West to the East. The spectra at the western inflow boundary contain more energy in WEAK_CP than WEAK_NO. However, with increasing fetch distances, the effect of CPM becomes muted. The spectral peak observed in Fig. 4.10 is also seen in both u - and v -velocity fields with and without CPM. The East-West spectra content is larger than North-South spectra which points to a certain degree of anisotropy in this LES stability case.

Lastly, the vertical velocity spectra for WEAK LLJ are shown in Figs. 4.13 and 4.14 and is indicative of the potential temperature variations induced by CPM. The spectra do not exhibit the typical highest energy at the lowest wavenumber that decays rapidly at the subsequent higher wavenumber, as seen in the horizontal velocities spectra. The maximum is a characteristic of horizontal mesoscale forcing nested within the LES domain. Unlike the East-West spectra for horizontal velocities, turbulence at low wavenumbers are steadily increasing with increasing fetch, regardless of CPM and are self similar half way into the domain. This indicates turbulence at these larger scales are effectively being spun up in the domain. In contrast, the North-South spectra shows a different behavior of turbulence. Turbulence at low wavenumbers decrease further east into the domain and turbulence scales about $k_y \approx 10^{-2} \text{ rad m}^{-1}$ peak quickly. This peak is seen in both WEAK_CP and WEAK_NO indicating that the production length scale $O(500) \text{ m}$ is a feature of this WEAK LLJ system. The use of CPM in this WEAK LLJ can be seen at the higher wavenumbers range in spectra near the inflow boundaries. CPM is able to inject energy within $4 \times 10^{-3} - 2 \times 10^{-2} \text{ rad m}^{-1}$ range. However, it does not

Figure 4.9: Top panel: Development of East-West u velocity spectra in y -direction at 80 m from 30 July 2004 0200 UTC to 30 July 2004 0600 UTC. Lines are colored by distance (indicated in numbers within spectra plots) from southern boundary in units of Δy . Bottom panel: Time-averaged contours of u velocity at 80m from 30 July 2004 0200 UTC to 30 July 2004 0600 UTC.

Figure 4.10: Top panel: Development of East-West v velocity spectra in y -direction at 80 m from 30 July 2004 0200 UTC to 30 July 2004 0600 UTC. Lines are colored by distance (indicated in numbers within spectra plots) from southern boundary in units of Δy . Bottom panel: Time-averaged contours of v velocity at 80m from 30 July 2004 0200 UTC to 30 July 2004 0600 UTC.

Figure 4.11: Top panel: Development of North-South u velocity spectra in x -direction at 80 m from 30 July 2004 0200 UTC to 30 July 2004 0600 UTC. Lines are colored by distance (indicated in numbers within spectra plots) from western boundary in units of Δx . Bottom panel: Time-averaged contours of u velocity at 80m from 30 July 2004 0200 UTC to 30 July 2004 0600 UTC.

Figure 4.12: Top panel: Development of North-South v velocity spectra in x -direction at 80 m from 30 July 2004 0200 UTC to 30 July 2004 0600 UTC. Lines are colored by distance (indicated in numbers within spectra plots) from western boundary in units of Δx . Bottom panel: Time-averaged contours of v velocity at 80m from 30 July 2004 0200 UTC to 30 July 2004 0600 UTC.

Figure 4.13: Top panel: Development of East-West w velocity spectra in y -direction at 80 m from 30 July 2004 0200 UTC to 30 July 2004 0600 UTC. Lines are colored by distance (indicated in numbers within spectra plots) from southern boundary in units of Δy . Bottom panel: Time-averaged contours of w velocity at 80m from 30 July 2004 0200 UTC to 30 July 2004 0600 UTC.

Figure 4.14: Top panel: Development of North-South w velocity spectra in x -direction at 80 m from 30 July 2004 0200 UTC to 30 July 2004 0600 UTC. Lines are colored by distance (indicated in numbers within spectra plots) from western boundary in units of Δx . Bottom panel: Time-averaged contours of w velocity at 80m from 30 July 2004 0200 UTC to 30 July 2004 0600 UTC.

significantly decrease the fetch distance required.

4.3.2 STRONG LLJ

4.3.2.1 Comparison to HRDL profiles

Having compared the performance of LES in simulating a weak LLJ in a more stable atmosphere, I now investigate LES performance in simulating a strong LLJ in a less stable atmosphere. The strong jet occurred during daytime (1545 LST to 1745 LST) which points to warm air advection from land to sea as a mechanism, rather than just radiative cooling mechanism due to the diurnal cycle.

Figure 4.15: Time series of horizontal wind speed profile along ship track from 31 July 2004 2045 UTC (1545 LST) to 31 July 2004 2245 UTC (1745 LST).

Similar to the WEAK case, the LES runs are not significantly qualitatively different from GFS_MYNN for both wind speed (Fig. 4.15) and wind direction

Figure 4.16: Time series of horizontal wind direction profile along ship track from 31 July 2004 2045 UTC (1545 LST) to 31 July 2004 2245 UTC (1745 LST).

(Fig. 4.16). I also see a more noisy profiles of the LLJ in the LES runs, compared to the smooth LLJ structure in GFS_MYNN. Fig. 4.17 shows that the average LLJ heights are similar to GFS_MYNN and that they are overestimated compared to HRDL. Unlike in the WEAK case, the LES overestimated the height and the nose speed of the LLJ. The simulated LLJ profiles have a less pronounced jet structure above the nose as the wind shear above and below the LLJ nose is weaker in the simulations compared to HRDL.

The averaged wind direction profiles for this time period is shown in Fig. 4.18. The LES profiles are too southerly throughout the 500 m height by approximately 5° . The overall wind veer of ($\approx 30^\circ$) in HRDL is well represented by the LES simulations, unlike that in WEAK.

There is little difference separating the LES simulations seen in these average

Figure 4.17: Averaged wind speed profile for STRONG time period from 31 July 2004 2045 UTC (1545 LST) to 31 July 2004 2245 UTC (1745 LST).

Figure 4.18: Averaged wind direction profile for STRONG time period from from 31 July 2004 2045 UTC (1545 LST) to 31 July 2004 2245 UTC (1745 LST).

profiles. The Cell-Perturbation method does not seem to offer an advantage in this STRONG case as well. However, using the diagnostic metrics, I find that Cell-Perturbation method do offer some advantage over the regular LES.

The mean wind speed bias is shown in the top left panel of Fig. 4.19. It shows that the mean speed is predominantly overestimated, except at the LLJ height. Within the subjet layer (< 100 m), the LES models have reduced the bias compared to GFS_MYNN. The NBIAS plot (top right panel of Fig. 4.19) further illuminates the use of LES compared to GFS_MYNN. The LES models consistently outperforms GFS_MYNN within the profile evaluated. This is in contrast with the WEAK LLJ. The effect of Cell-Perturbation is seen in the NBIAS metric, STRONG_CP has reduced the NBIAS minima at 30 m from 0.9 in STRONG_NO to 0.6. Moreover, the improvement in NBIAS through the use of LES is more substantial in STRONG than in WEAK. This is likely due to overall better representation of turbulence in this stability regime than in WEAK. The use of Cell-Perturbation further accentuate the positive effects of improved turbulence representation.

I then compare the skill in variance prediction of the LES. Similar to WEAK, I find an overall degradation in variance prediction when LES is used. Unlike in WEAK, the correlation of the LES models are worse than GFS_MYNN at all vertical levels. There is a peak in ρ at about 100 m for both LES models that is comparable to GFS_MYNN performance. The models have poor correlation to the HRDL dataset that is in contrast to overall 48-h time period evaluations.

The increase in turbulence within LES models result in overall better η scores than GFS_MYNN, except near the surface and at the LLJ layer. The decrease in η between 50 m and 150 m is due to overestimation of the LLJ height and the increase in turbulence in LES. Within this layer, the LLJ in HRDL is at its peak and the variance is low. However, in the LES models, this is subjet region where variance is high, coupled with increased turbulence in LES models, this resulted in a poorer η score compared to GFS_MYNN.

The degradation in ρ is due to the increased turbulence modelled in LES, leading to poorer phase correlation. The poor ρ scores contribute heavily to the poor α scores of the LES models. The minimum in α scores for both LES is just below 100 m, the region where the LES models have good ρ and η scores. The LES consistently scored larger α throughout the 500 m profile compared to GFS_MYNN, unlike the

performance in WEAK.

Overall, despite poor α scores, the NBIAS improvements in the LES models are significant enough to have improved NRMSE scores over the GFS_MYNN model. Cell-Perturbation method is able to improve NRMSE scores from STRONG within 150 m, the simulated subjet region. There is a slight degradation in performance of STRONG_CP compared to NOCP within the 150 m to 250 m heights and that is due to increase in bias. Likely source of this bias is the increased turbulence in STRONG_CP that mixed higher momentum aloft the LLJ nose.

Figure 4.19: Profile for diagnostic metrics: mean wind speed bias (top left), NBIAS (top right), ρ (middle left), η (middle right), α (bottom left) and NRMSE (bottom right). Evaluated for STRONG time period from 31 July 2004 2045 UTC (1545 LST) to 31 July 2004 2245 UTC (1745 LST).

4.3.2.2 Spectral Analysis

The instantaneous velocity fields at 31 July 2004 2045 UTC for both STRONG_CP and STRONG_NO are shown in Fig. 4.20. Unlike the WEAK case, there is a significant qualitative difference between the velocity fields in STRONG_CP and STRONG_NO. Small-scale turbulence features are readily seen throughout the domain in all three velocity fields. The STRONG_NO remains relatively smooth in comparison, although small wave-like structures can be seen in the region between 400 to 600 ΔX and 400 to 600 ΔY . Strong vertical velocity perturbations are seen in near the western and southern inflow boundaries in the bottom left panel of Fig. 4.20 that is due to CPM, the strength of these perturbations dissipated halfway into the domain.

I apply the same spectral analysis methodology as in the WEAK case. The spectral analysis is carried out at the height of 150m. Examining the spectra for STRONG case, I see evidence of CPM enhancing turbulence that results in appreciable difference from the LES without CPM.

In Fig. 4.21, I find an underdeveloped East-West spectra for STRONG. Near the southern boundary, there is a peak at $k_x \approx 4 \times 10^{-3} \text{ rad m}^{-1}$ which corresponds to 1.5 km that serves to generate turbulence. This peak disappears with increased fetch distance and appears to have distributed its energy content to higher wavenumbers. A peak in spectra energy is seen at the lower wavenumbers near the northern boundary of the domain at $k_x \approx 6 \times 10^{-4} \text{ rad m}^{-1}$ that would correspond to a wavelength of $\approx 9 \text{ km}$. The peak is due to the north-south oriented wave visible in each of the time-averaged plot of the velocity fields.

The spectral plot for STRONG_CP reveals a different behaviour of the turbulence in contrast to STRONG_NO. The spectra near the southern boundary show developed turbulence between $10^{-3} - 10^{-2} \text{ rad m}^{-1}$. CPM is able to enhance turbulence within this range close to the inflow boundary. Further north into the domain, the spectra energy are reduced at the higher wavenumbers and are self-similar with fetch ≥ 450 from the inflow. The excess turbulence introduced by CPM got reduced. Conversely, at the smaller wavenumbers, there is an increase in spectra energy due to the wave seen. The peak at this range is similar to STRONG. The spectra of the v -velocity field show similar trends.

The spectral analysis of the horizontal fields along North-South direction (Figs.

Figure 4.20: Contours of instantaneous velocity fields at 31 July 2004 2045 UTC (1545 LST) for STRONG LLJ interpolated to 150 m. Left column: STRONG_CP. Right column: STRONG_NO. Top, middle and bottom rows are u -, v - and w -velocity fields respectively.

Figure 4.21: Top panel: Development of East-West u velocity spectra in y -direction at 150 m from 31 July 2004 2000 UTC to 1 August 2004 0000 UTC. Lines are colored by distance (indicated in numbers within spectra plots) from southern boundary in units of Δy . Bottom panel: Time-averaged contours of u velocity at 150 m from 31 July 2004 2000 UTC to 1 August 2004 0000 UTC.

Figure 4.22: Top panel: Development of East-West v velocity spectra in y -direction at 150 m from 31 July 2004 2000 UTC to 1 August 2004 0000 UTC. Lines are colored by distance (indicated in numbers within spectra plots) from southern boundary in units of Δy . Bottom panel: Time-averaged contours of v velocity at 150 m from 31 July 2004 2000 UTC to 1 August 2004 0000 UTC.

Figure 4.23: Top panel: Development of North-South u velocity spectra in x -direction at 150 m from 31 July 2004 2000 UTC to 1 August 2004 0000 UTC. Lines are colored by distance (indicated in numbers within spectra plots) from western boundary in units of Δx . Bottom panel: Time-averaged contours of u velocity at 150 m from 31 July 2004 2000 UTC to 1 August 2004 0000 UTC.

Figure 4.24: Top panel: Development of North-South v velocity spectra in x -direction at 150 m from 31 July 2004 2000 UTC to 1 August 2004 0000 UTC. Lines are colored by distance (indicated in numbers within spectra plots) from western boundary in units of Δx . Bottom panel: Time-averaged contours of v velocity at 150 m from 31 July 2004 2000 UTC to 1 August 2004 0000 UTC.

Figure 4.25: Top panel: Development of East-West w velocity spectra in y -direction at 150 m from 31 July 2004 2000 UTC to 1 August 2004 0000 UTC. Lines are colored by distance (indicated in numbers within spectra plots) from southern boundary in units of Δy . Bottom panel: Time-averaged contours of w velocity at 150 m from 31 July 2004 2000 UTC to 1 August 2004 0000 UTC.

Figure 4.26: Top panel: Development of North-South w velocity spectra in x -direction at 150 m from 31 July 2004 2000 UTC to 1 August 2004 0000 UTC. Lines are colored by distance (indicated in numbers within spectra plots) from western boundary in units of Δx . Bottom panel: Time-averaged contours of w velocity at 150 m from 31 July 2004 2000 UTC to 1 August 2004 0000 UTC.

4.23 and 4.24 reveal similar behavior as the East-West spectra. In STRONG_NO, there is a peak at the same wavenumber $k_y \approx 4 \times 10^{-3} \text{ rad m}^{-1}$ that is the source of turbulence production. As the flow develops further towards the eastern region of the domain, the energy spectra content increases at most wavenumbers and the peak remains. When CPM is applied, the westernmost spectrum shows elevated energy at higher wavenumbers. With increasing fetch, the spectra content decreases and converge at $\text{fetch} \geq 250$. The peak at $k_y \approx 4 \times 10^{-3} \text{ rad m}^{-1}$ is much more apparent for the v -velocity field in the eastern portion of the domain, $\text{fetch} \geq 450$ in STRONG_CP. The energy of this peak is similar in magnitude to STRONG_NO at the same fetch distances. This indicates that it is a dynamical feature of this case-study and it is not smeared out by the CPM. It also appears that the evolution of the STRONG_NO spectra is tending towards that of STRONG_CP. There is a little change in energy for the STRONG_CP spectra between $5 \times 10^{-4} - 5 \times 10^{-3} \text{ rad m}^{-1}$.

The energy of East-West spectra and the North-South spectra in this STRONG case are similar in magnitude as compared to the difference in energy between the spectra in WEAK indicates a lower level of anisotropy as expected in this stability regime. There is a short inertial subrange in the STRONG_CP spectra and energy decayed quickly at higher wavenumbers. The uptick tail at the highest wavenumbers suggest that increased grid resolution is needed to model the sub-grid scale turbulence more accurately.

The vertical velocity spectra for STRONG LLJ are shown in Figs. 4.25 and 4.26. The effect on the spectra using CPM in this STRONG LLJ case is much more significant than in WEAK LLJ case. In both figures, I see that CPM is able to spin-up turbulence in the wavenumber ranges $10^{-3} - 10^{-2} \text{ rad m}^{-1}$. The spectra closest to the inflow boundaries (southern and western in Figs. 4.25 and 4.26 respectively) exhibited enhanced energy due to CPM but quickly diminished with increased fetch. The spectra for STRONG_CP converges halfway into the domain and a short inertial subrange can be seen in both figures. Spectra without CPM do not exhibit these characteristics. It is clear that CPM helps to enhance the quality of the turbulence characteristics in this STRONG LLJ by introducing turbulence scales that have yet to be developed in the STRONG_NO domain.

4.4 Summary and Discussion

In this study, I investigate the use of high-resolution LES to simulate real LLJ events and the use of CPM in improving turbulence in nested WRF-LES domains. I use two LLJ events of different strengths to study the performance of LES and CPM under different stability regimes. I use a set of diagnostic metrics to assess simulated LLJ profiles against observed HRDL profiles and quantify the performance of the simulations. I also assess the quality of turbulence introduced by CPM against standard WRF-LES through spectral analysis.

LES did not result in significant differences in the mean LLJ characteristics. The main source of error remains that of the forcing used. However, there are quantifiable improvements to biases in horizontal wind velocity prediction, particularly in the LLJ subject layer. The increase in turbulence in this layer resulted in appropriate dissipation of momentum. The use of CPM did not improve biases for WEAK but was able to further reduce biases for STRONG.

LES did not improve variance prediction skill score, α . LES increased the overall variance in wind velocities which improved the η score, a prediction of variance amplitude. However, the increase in variance reduced the ρ score of the LES results against HRDL, which led to the decrease in α score.

LES is able to reduce overall NRMSE compared to GFS_MYNN, the mesoscale simulation of the LLJ events. This is an encouraging outcome for high-resolution modelling of real data cases. The NRMSE improvement is greater in STRONG than in WEAK, especially when CPM is applied.

The use of CPM resulted in better turbulence spectra in STRONG than WEAK. CPM is able to spin up turbulence quickly and effectively in STRONG LLJ regime as compared to WEAK. In the WEAK case, the spectra peak at higher wavenumber, $k \approx 10^{-2} \text{ rad m}^{-1}$ is close to $6\Delta x$, the effective resolution of WRF-LES (Skamarock et al. 2008). The resulting inertial subrange is not well resolved in WEAK. As CPM is meant to be applied at a spatial scale that corresponds to a fully resolved inertial range, the lack of turbulence spin-up in WEAK is not indicative of CPM failure. In STRONG, the production range is in $k \approx 5 \times 10^{-3} \text{ rad m}^{-1}$. CPM is able to perturb scales between the production range and the dissipation range resulting in a greater spectra content with decreased fetch distances.

Preliminary resolution sensitivity test (see Section 4.A) showed that while higher

resolution (27 m) simulation was able to better resolve turbulence at higher wavenumbers as expected, there is little qualitative difference separating the performance of 50 m and 27 m LES resolution in WEAK LLJ condition with CPM applied. Hence, the choice of horizontal resolution, while being a factor to be considered, does not invalidate the current study and that the 50 m resolution chosen is sufficiently accurate.

4.A Supplementary Material

In this section, I present findings from the resolution sensitivity study of the LES domain. While these results are preliminary, they support the conclusions made in the previous section.

LES Resolution Sensitivity Study - 10 m

The domain consists of three nests, an outer mesoscale nest (d04), an intermediate LES nest (d05) and an innermost LES nest (d06) with horizontal resolutions of 250 m, 50 m and 10 m respectively. The domain sizes ($N_x \times N_y \times N_z$) are 300x300x60, 600x600x60 and 300x300x60 respectively. The 60 vertical model levels are concentrated mainly within the lower boundary layer and progressively stretched upwards. The average vertical resolution within the lowest 150 m is ~ 15 m and within the lowest 1000 m is ~ 35 m. The intermediate LES domain is designed to encompass the positions of the ship over an identified 1-h period. The ship track is confined within the top half of d05 as the winds are predominantly southerly and south-south-westerly over this period. The outer mesoscale domain is designed such that the LES domain is kept within the north-eastern region of the domain. The distance of d05 from the southern and western boundary of d04 is kept the same as both WEAK and STRONG. D06 is centered about a single HRDL profile. The initial and boundary conditions for d04 are obtained from domain d03 of GFS_MYNN using the ndown procedure in WRFV3.6.1 (Skamarock et al. 2008) at 15-minute intervals. Both Figs. 4.27 and 4.28 show the domain set-up for both LES10 and LES10E6 periods within d03. Table 4.4 summarizes the model configurations used in this study.

First, the sensitivity of the 10 m domain to a 50 m parent domain with and without CPM applied were tested. Fig. 4.29 shows the instantaneous fields of d06 of both LES10_CP and LES10_NO_CP. I can see unrealistic coherent structures in the d06 domain in all velocity fields. These structures are deemed unrealistic due to strong variations of $\sim 5 \text{ m s}^{-1}$ from ridge to trough over a short spatial span of 300 m in both u - and v - velocity fields, moreover, these waves are not seen in d05 (50 m) parent domain (not shown).

Fig. 4.30 shows the spectral analysis of both LES10_CP and LES10_NO_CP.

Figure 4.27: WRF domain set-up showing 2 nested domains for LES10 LLJ time period within d04 (250 m), the intermediate mesoscale domain nested with D03 generated from mesoscale case-study in Chapter 3. D05 (50 m) is the LES nested domain. D06 (10 m) is the inner-most LES domain nested within d05. Black markers indicate the HRDL track between 31 July 2004 0400 UTC (2300 LST) to 31 July 2004 0500 UTC (0000 LST)

Figure 4.28: WRF domain set-up showing 2 nested domains for LES10E6 LLJ time period within d04 (250 m). Similar configuration to 4.27, except D06 (10m) is a 600x600 domain instead of 300x300.

Figure 4.29: Contours of instantaneous velocity fields at 31 July 2004 0400 UTC (2300 LST) interpolated to 100 m. Left: LES10_CP. Right: LES10_NO_CP. Top, middle and bottom rows are u –, v – and w – velocity fields respectively.

Table 4.4: Names and summary of all the model configurations in this section

	HRDL	Dura-	Simulation	Duration	CPM	CPM
	tion				(D05)	(D06)
LES10_CP	31 July 2004 0400 UTC to 0500 UTC	31 July 2004 0200 UTC to 0600 UTC			Yes	No
LES10_NO_CP	31 July 2004 0400 UTC to 0500 UTC	31 July 2004 0200 UTC to 0600 UTC			No	No
LES10E6	31 July 2004 0400 UTC to 0500 UTC	31 July 2004 2000 UTC to 0000 UTC			Yes	No

There is a clear peak at ~ 300 m in both simulations, corresponding to the wavelength of the coherent structures seen in Fig.4.29, regardless the choice of parent domain's Cell-Perturbation Method (CPM) configuration. Hence, it can be concluded that these structures observed are not an artefact introduced by CPM but is instead native to the real WRF-LES implementation. While only the East-West u -velocity spectra in y-direction is presented, the spectra analysis for the other velocity fields show similar findings that there is little difference between LES10_CP and LES10_NO_CP at 10 m resolution.

It is possible that the structures seen in Fig. 4.29 is due to a limited domain size of d06. I enlarged the 10 m domain of LES10 from 300x300 to 600x600 (see Fig. 4.28) to test this hypothesis. In Fig. 4.31, I see that these structures appear regardless of the LES domain size. These structures are likely to be numerical in nature rather than physical.

The origin of these structures are tracked to the start of the LES simulation. In Fig. 4.32, I can see wave-like disturbances from the western and eastern boundaries propagating in time. After 1 second of simulation time from the LES initiation (top right panel of Fig. 4.32), I can see waves extending $\sim 30 \Delta X$ into the domain from both western and eastern boundaries and in 2 seconds (middle left panel), these

Figure 4.30: Top panel: Development of East-West u velocity spectra in y -direction at 100 m from 31 July 2004 0400 UTC to 31 July 2004 0500 UTC. Lines are colored by distance (indicated in numbers within spectra plots) from southern boundary in units of Δy . Bottom panel: Time-averaged contours of u velocity at 100m from 31 July 2004 0400 UTC to 31 July 2004 0500 UTC. The vertical lines demarcate the wavenumbers corresponding to the length of CPM perturbation applied ($8\Delta_{50m}$), WRF effective resolution of parent domain, d05 ($6\Delta_{50m}$) and WRF effective resolution of d06 ($6\Delta_{10m}$).

waves have extended to $\sim 60 \Delta X$, and finally in 15 seconds, these structures appear to extend throughout the domain. Since the average inflow speed is $\sim 3 \text{ m s}^{-1}$, these waves are not due to physical conditions within flow. Moreover, the speed of this disturbance propagation is $\sim 300 \text{ m s}^{-1}$, suggests that it could be acoustic related. It is also likely specified boundary conditions necessary for a nested real WRF-LES simulation might be too numerically stiff for a stable boundary layer, resulting in these unrealistic structures. However, further investigations will be required to determine the exact cause and to arrive at a solution.

Figure 4.31: Contours of instantaneous velocity fields at 31 July 2004 0400 UTC (2300 LST) interpolated to 100 m. Left: LES10_CP, 300x300 domain. Right: LES10E6, 600x600 domain, cropped to show the same 300x300 area in the 300x300 domain. Top, middle and bottom rows are u –, v – and w – velocity fields respectively.

Figure 4.32: Contours of instantaneous u -velocity field at 31 July 2004 0000 UTC at z_4 . From left to right, top to bottom: Start, 1-second, 2-second, 3-second, 4-second, 15-second of simulation time.

LES Resolution Sensitivity Study - 27 m

Given the issues with a nested 10 m LES domain highlighted in the previous section, another set of higher resolution simulation of the WEAK LLJ was carried. Instead of nesting a 10 m LES within a 50 m LES domain, the number of domains is kept the same at 5 and the innermost domain, d05, is downscaled from the 250 m intermediate domain with a grid ratio of 9, improving the resolution by 1.8 times of the original WEAK_CP case while keeping the same spatial domain of 30 km x 40 km. Figures 4.33 and 4.34 shows the computational domains for WEAK_CP and WEAK27_CP respectively.

In order to use more CPUs for this numerically larger LES domain, d05, the intermediate domain, d04, needed to be doubled in both the north and east-wards direction. There are technical issues associated with parallelisation in WRF, when the number of CPUs used is more than the number of horizontal grid points. Since, the flow from d04 into d05 is south-westerly throughout the LES simulation duration, this domain extension should have no impact on the boundary conditions ingested by d05.

Table 4.5: Names and summary of all the model configurations in this section

	Domains	Domain Sizes	Simulation Duration	CPM ap- plied
WEAK_CP	D04, D05	300x300x60, 600x800x60	30 July 2004 0200 UTC to 0600 UTC	Yes
WEAK27_CP	D04, D05	600x600x60, 1080x1440x60,	30 July 2004 0200 UTC to 0600 UTC	Yes

Figure 4.33: Original WRF domain set-up showing 2 nested domains for WEAK LLJ time period within d03 (1250 m), the outermost domain generated from mesoscale case-study in Chapter 3. D04 (250 m) is the intermediate mesoscale domain. D05 (50 m) is the inner-most LES domain nested within d04. Black markers indicate the HRDL track between 30 July 2004 0245 UTC (2145 LST) to 30 July 2004 0445 UTC (2345 LST)

Figure 4.34: Similar to Fig. 4.33, except D04 (250 m) is 600x600 instead of 300x300 and D05 has grid resolution of 27.8 m instead of 50 m.

Figure 4.35: Contours of instantaneous velocity fields at 30 July 2004 0245 UTC (2145 LST) for WEAK LLJ interpolated to 80 m. Left column: WEAK27_CP. Right column: WEAK_CP. Top, middle and bottom rows are u -, v - and w -velocity fields respectively.

The instantaneous velocity fields for both WEAK27_CP and WEAK_CP are shown in Fig. 4.35. Both simulations are qualitatively similar in representing the large scales variation, as expected due to the same initial and boundary conditions. There are slight differences in the small scale structures in the north-eastern portion of the domain. As expected, the higher resolution of WEAK27_CP is able to represent smaller scale structures compared to WEAK_CP and these smaller scale structures are able to break down some of the coherent larger scale structures.

Next I perform the same spectral analysis of WEAK27_CP and compare against WEAK_CP. The plots are shown in Figs. 4.36- 4.39. Only the East-West spectra of the all three velocity fields and the North-South spectra of u are presented.

In Fig. 4.36, I see that while the spectral power at higher wavenumbers are increased in WEAK27_CP near the inflow boundaries, the spectral curves past the the middle of the domain (yellow curves and higher in Figs. 4.36-4.38 and light blue curves and higher in Figs 4.39) are qualitatively similar to WEAK_CP. The use of the higher resolution in WEAK27_CP does result in better resolved turbulence at finer scales, as expected. However, this comparison also shows that the performance between WEAK27_CP and WEAK_CP are not significantly different. Until I are able to solve the numerical issues associated with the 10 m resolution simulation, I posit that the choice of grid horizontal resolution does not have a significant impact on the results and the 50 m is sufficiently accurate for the original study.

Figure 4.36: Top panel: Development of East-West u velocity spectra in y-direction at 80 m from 30 July 2004 0200 UTC to 30 July 2004 0600 UTC. Lines are colored by distance (indicated in numbers within spectra plots) from southern boundary in units of Δy . Bottom panel: Time-averaged contours of u velocity at 80m from 30 July 2004 0200 UTC to 30 July 2004 0600 UTC. The vertical lines demarcate the wavenumbers corresponding to the length of CPM perturbation applied ($8\Delta_r$), WRF effective resolution of the domain, d05 ($6\Delta_r$), where r is the resolution of LES domain.

Figure 4.37: Top panel: Development of East-West v velocity spectra in y -direction at 80 m from 30 July 2004 0200 UTC to 30 July 2004 0600 UTC. Lines are colored by distance (indicated in numbers within spectra plots) from southern boundary in units of Δy . Bottom panel: Time-averaged contours of v velocity at 80m from 30 July 2004 0200 UTC to 30 July 2004 0600 UTC. The vertical lines demarcate the wavenumbers corresponding to the length of CPM perturbation applied ($8\Delta_r$), WRF effective resolution of the domain, d05 ($6\Delta_r$), where r is the resolution of LES domain.

Figure 4.38: Top panel: Development of East-West w velocity spectra in y-direction at 80 m from 30 July 2004 0200 UTC to 30 July 2004 0600 UTC. Lines are colored by distance (indicated in numbers within spectra plots) from southern boundary in units of Δy . Bottom panel: Time-averaged contours of w velocity at 80m from 30 July 2004 0200 UTC to 30 July 2004 0600 UTC. The vertical lines demarcate the wavenumbers corresponding to the length of CPM perturbation applied ($8\Delta_r$), WRF effective resolution of the domain, d05 ($6\Delta_r$), where r is the resolution of LES domain.

Figure 4.39: Top panel: Development of North-South u velocity spectra in x -direction at 80 m from 30 July 2004 0200 UTC to 30 July 2004 0600 UTC. Lines are colored by distance (indicated in numbers within spectra plots) from western boundary in units of Δx . Bottom panel: Time-averaged contours of u velocity at 80m from 30 July 2004 0200 UTC to 30 July 2004 0600 UTC. The vertical lines demarcate the wavenumbers corresponding to the length of CPM perturbation applied ($8\Delta_r$), WRF effective resolution of the domain, d05 ($6\Delta_r$), where r is the resolution of LES domain.

Chapter 5

Conclusions

In this thesis, Large-Eddy Simulation (LES) is used for high-resolution modelling of offshore wind conditions. The aim for increased accuracy and modelling low-level jet (LLJ) events offshore in a coastal region. Such LLJ events are a prevalent condition that occur offshore in coastal regions due in part to the land-sea temperature contrast, especially during summer in the mid-latitudes. LLJ, while being characterized as high wind speeds close to the surface, can substantially boost wind power generation. In addition, the occurrence of LLJ may adversely affect structural loading on wind turbines. The mechanisms of LLJ formation are an interaction of inertial oscillation and diurnal heating over sloping terrain, as studied by Blackadar (1957) and Holton (1967) respectively. The common thread between both mechanisms is that the decay of turbulence aloft a stable surface layer results in accelerated jet-like winds. It is thus important to properly model turbulence during these LLJ events. LES is employed to study if improved turbulence representation leads to more accurate LLJ predictions.

5.1 Thesis Summary

Prior to the LES study, the variability in selected model configurations in mesoscale simulations needed to be quantified. I selected CBLAST-LOW 2001 campaign as a first test case. CBLAST-LOW 2001 is a one of the few field campaigns that observed coastal LLJ and it was carried out in New England region of North America, a region of interest for wind energy development. The initial/boundary conditions (IBC), planetary boundary layer (PBL) scheme, sea surface temperature (SST) con-

ditions and vertical resolution was varied. I found that there were significant biases in the predicted wind profile compared with the available rawinsonde soundings that featured the LLJ. However, a single sounding was not sufficient to evaluate the simulations' performance hence I also use ASOS data for further analysis (see Chapter 2). The observed LLJ arose due to a warm frontal passage through the region during the time of simulation. Hence, the surface temperature readings from ASOS provide a reasonable proxy of the broader mesoscale conditions during the LLJ event.

Based on the ASOS data analysis, it was determined that ERA_MYNN gave the best performance among all the mesoscale model configurations tested. ERA_MYNN was thus used as a baseline control and performed inter-model comparisons to quantify the variability introduced by IBC, PBL, SST and vertical resolution. Similar to other modelling studies, IBC remains the greatest source of mesoscale variability. In CBLAST-LOW 2001, the frontal passage in NARR lags ERA, which contributed to large NRMSE offshore. Other configurations such as PBL, SST and vertical resolution did not modify variance errors as much and were manifested mainly as biases instead. This study indicates that while the LLJ formation is largely dictated by mesoscale boundary conditions and that PBL and vertical resolution still matters in affecting the local vertical structure. SST did not appear to play a role in triggering the LLJ which is largely attributed to the frontal passage as the dominant mesoscale forcing.

After the preliminary study using CBLAST-LOW 2001 campaign, a similar methodology was used to evaluate the NEAQS-ITCT 2004 campaign, which measured offshore conditions in the same geographical region. Learning from CBLAST-LOW 2001, a case-study duration that featured LLJ events but did not experience anomalous mesoscale conditions (i.e. frontal passage) was selected. Once again, IBC, PBL and SST sensitivity tests were carried out. A vertical resolution study was neglected due to computational constraints and other studies have found that the increase in vertical resolution did not significantly alter the vertical structure compared to the choice of PBL schemes.

In the NEAQS-ITCT 2004 case study, I analyzed a 48-h period during which 2 different LLJ events were observed - a WEAK LLJ ($\sim 10 \text{ m s}^{-1}$) and a STRONG LLJ ($\sim 18 \text{ m s}^{-1}$) (see Chapter 3). Overall, the simulations were able to replicate the broad features of the LLJs fairly well. The WEAK LLJ was underpredicted

and earlier than the observed LLJ whereas the STRONG LLJ was overestimated and lagged the observed LLJ. The surface temperature model predictions were also compared to ASOS data. In all simulations, the land surface had an average cold bias with no discernible geographical pattern within the domain of interest. All simulations had similar NRMSE scores for surface temperature and surface wind speeds.

Analysis of modelled wind profiles against lidar measurements from the field campaign differentiated the simulations' performance. The simulations scored similar in α but with differing NBIAS scores throughout the profile. The diagnostic tool showed the need for evaluating winds as a vector quantity than a set of decomposed scalars (wind speed and direction). GFS showed almost zero mean bias aloft over the 48-h period but it was clear that a directional bias exists from qualitative comparisons. NBIAS was able to reveal this discrepancy easily and showed that despite seemingly different performances by the simulations, their NBIAS are similar aloft. The choice of IBC was still the significant source of variability and contributed to the overall NRMSE. GFS_MYNN was selected based on its overall lowest NBIAS and NRMSE scores, especially within the LLJ layer (< 150 m).

For a 24-h period featuring a range of conditions such as on 30 July 2004 (WEAK LLJ), the choice of IBC is the greatest source of variability, as seen in the NBIAS and NRMSE profiles. Without accurate representation of synoptic conditions, WRF was unable to model the transition between regimes well, this led to the difference in LLJ timings between the simulations. GFS out performed ERA and NARR in both NBIAS and NRMSE scores. The low α score profiles ($\alpha < 0.3$) showed that variance prediction performance were comparable amongst the simulations.

On the other hand, for the period of strong forcing on Day 2 (STRONG LLJ), the choice of IBC had little impact in the subjet layer, shown in the similar NBIAS scores for all the model configurations. The choice of IBC mattered in NBIAS aloft. This points to inherent model physics and dynamics as the source of biases rather than IBC within the turbulent subjet layer. The choice of PBL scheme and SST condition also manifested as NBIAS aloft. The α scores for the simulations were more varied and with poorer performance within the subjet layer ($\alpha \sim 0.5$) compared to during WEAK LLJ. The overall NRMSE scores indicated that GFS_MYNN was the best performing configuration.

Finally, the role of SST in LLJ formation within the WRF model was tested. The SST in GFS_SST was increased uniformly by 2 K from that of GFS_MYNN. The reduced land-sea temperature contrast did not result in significant differences in NBIAS and NRMSE from that of GFS_MYNN. This likely points to model errors, where the increase in SST did not alter the bulk stability within the marine boundary layer, likely due to poor surface parameterization interaction with PBL schemes. The modeled marine boundary layer tended to be cooler and more statically stable compared to the observed temperature soundings.

Having selected GFS_MYNN as the parent mesoscale domain within which to test the nesting of LES, I designed 2 sets of simulations, categorized by the strength of the LLJ and referred to as WEAK and STRONG. In each set of LES simulations, a 50 m LES domain is one-way nested within a 250 m intermediate mesoscale domain. Each LES ran for 6 hours with the initial 2 hours discarded as spin-up. In each LLJ condition, the effectiveness of the Cell-Perturbation Method (CPM) in generating turbulence scales within the nested LES domain was tested. CPM is a method that introduces potential temperature perturbations at inflow boundaries with the purpose of aiding turbulence generation.

LES did not significantly alter the mean characteristics of the predicted wind speed profiles compared to the parent mesoscale domain. Upon further evaluation, NBIAS and NRMSE were reduced for both cases examined, WEAK and STRONG, particularly within the subjet and jet core layer. There was no clear advantage in using CPM to reduce NBIAS and NRMSE for the WEAK case. In contrast, CPM has overall improved NBIAS and NRMSE scores in the STRONG case compared to the control simulation without CPM. This likely points to the importance of better turbulence representation when modelling SBL and LLJ.

The quality of turbulence generated with and without CPM was investigated in order to quantify the performance of CPM in different stability regimes. Spectral analysis of the u and v fields at height of the simulated LLJ during each respective regimes (~ 80 m for WEAK, ~ 150 m for STRONG) was performed.

In the WEAK case, there was no discernible inertial range that fits the $-5/3$ decay with and without CPM. It is likely that due to the stronger stability regime in WEAK, a finer horizontal resolution simulation would be needed to resolve more of the eddies within the production range before I can conclude on the effectiveness

of CPM in this forcing.

In the STRONG case, I found that CPM was able to reduce the fetch distance required much more effectively. The turbulence spectra were much improved compared to STRONG_NO. The spectra reached equilibrium approximately halfway into the domain from the inflow, which is not seen in STRONG_NO. CPM was able to inject the missing scales with the transition from between 250 m to 50 m and thus improved the spectrum. CPM was quite effective in this regime with the current configuration.

5.2 Significant Contributions

Traditional mesoscale models are often too coarse in resolution to account for the wide range of turbulent scales present in the atmospheric boundary layer while idealized LES simulations do not often take into account varying atmospheric conditions over the length of the simulation. In this work, the aim was to demonstrate the suitability of using high-resolution LES nested within traditional mesoscale NWP to more accurately model turbulence conditions within the atmospheric boundary layer. In this section, the achievements in improving the understanding of using such modelling tools are listed.

5.2.1 WRF Evaluation Methodology

A diagnostic suite suitable for quantifying model performance was used and yielded insight into WRF performance for LLJ modelling. The α score is useful to evaluate LLJ. LLJ is a phenomenon that evolves in space and time, hence the variance prediction during its evolution is equally as important as the biases in wind speed when comparing model performances. Furthermore, α can be broken down to both ρ and η which measures phase and amplitude of variance prediction. This level of diagnostic detail allows the source of errors in the model configurations to be discerned. In the CBLAST-LOW 2001 case-study, the α and ρ metrics pinpoint that the errors in LLJ are due to a phase difference in the frontal passage representation between ERA and NARR, rather than a simple difference in spatial representation. In the NEAQS-ITCT 2004 case-study, high-resolution lidar measurements from HRDL were used in the model evaluation of both mesoscale and LES simulations. It was shown that

while amplitude of variance in the subject layer is well predicted, the errors stem from poor ρ , phase prediction. This allows further in-depth study into improving this aspect of model behavior.

5.2.2 Nested WRF-LES

The one-way nesting capability of the WRF code was used to nest a high-resolution LES domain within a mesoscale domain ran with real IBC. The nested WRF-LES improved upon the prediction of LLJs, especially within the subject layer when compared to a traditional mesoscale model. By using higher resolution modelling and improved turbulence representation, I am able to better predict LLJ behavior. The metrics indicate that while LES has improved η scores however ρ scores have degraded. This reinforces the finding in the mesoscale study that phase prediction of the WRF model needs to be improved.

5.2.3 Cell Perturbation Method

The suitability of CPM to generate turbulence in realistic WRF-LES simulations was demonstrated. WRF-LES using CPM is able to improve upon biases in horizontal wind velocity prediction, particularly within the subject layer of WRF-LES LLJ predictions. CPM improved the quality of turbulence within the nested LES domain especially when the forcing was strong as evinced by the spectral analysis of STRONG_CP. CPM was also able to efficiently reduce the fetch distances required for real nested WRF-LES, reducing the need for large computational domain and hence reduction in computational costs. This will aid in the feasibility of carrying out more detailed and varied numerical WRF-LES experiments in the future.

5.3 Future Work

While having advanced some insights into using LES to model LLJ, there are still other avenues of research to pursue towards a greater understanding of LLJ modelling. I propose the following areas of study that can be extended from this work.

5.3.1 Role of SST and surface model in WRF

It was found that during strong LLJ event, the simulations showed similar bias profiles within the subjet layer despite having differing IBC. There was little difference between GFS_MYNN and GFS_SST, despite having a difference in 2 K in SST. This result seemingly contradicts the idea that the land-sea temperature contrast modulates LLJ evolution. It is suspected that the surface model did not interact well with the artificially imposed SST increase. Further study into this anomalous behavior is warranted. The phase prediction of variance is poor, especially within the subjet layer. The source of this error was not explored in the current study but it is possibly linked to the turbulence representation near the surface.

5.3.2 Further Cell-Perturbation Method Testing

The ability of CPM in generating turbulence was demonstrated in this case-study, especially during strong LLJ forcing. I validated that the general guidance of CPM parameters in Muñoz-Esparza et al. (2015) is indeed suitable. However, in both WEAK and STRONG cases, the stability regime and wind conditions, while not periodic, were fairly constant. The next step to test the robustness of CPM parameters, will be to apply it to a LES study with varying stability regime, such as the transition from quiescent winds to LLJ or vice-versa. The interaction of CPM with different sub-grid stress closures was not explored in this study. Future sensitivity studies using different LES closures such as the dynamic reconstruction model (Chow et al. 2005), locally averaged scale-dependent dynamic model (Basu and Porté-Agel 2005) and the nonlinear-backscatter anisotropic (Kosović 1997; Mirocha et al. 2010) should be carried out to gauge the effectiveness of CPM used in concert with these closure models for real nested WRF-LES simulations. This can lend greater confidence in incorporating such a turbulence spin-up method for future WRF-LES nested studies.

I note that by definition LES needs to be in the inertial subrange to be able to resolve turbulence. Given the resolution of the simulations carried out, the $-5/3$ trend cannot be classified as the traditional Kolmogorov inertial subrange, since a stable boundary layer has a typical dominant eddy size of 10 m. The $-5/3$ trend has been shown to exist outside of the traditional scaling arguments as theorized by Lindborg (2006) in a stable boundary layer. When a coupled WRF-LES is able to capture the energy cascade from larger to smaller scales, especially with the aid of CPM, is worth investigating further.

5.3.3 Wind Farm Modelling Study

The viability of using high-resolution nested WRF-LES is shown in this work. Although the current resolution used (50 m) is still far too coarse for detailed wind turbine modelling studies, the nested WRF-LES simulations can be further down-scaled and coupled with a Computational Fluid Dynamics (CFD) toolbox such as OpenFOAM (Churchfield et al. 2010, 2012b) to perform wind farm level studies. WRF-LES would be able to provide boundary conditions with more realistic turbulence and atmospheric dynamics. In addition, the Generalized Actuator Model (GAD) implemented in WRF (Mirocha et al. 2014a; Aitken et al. 2014) has shown promise in modelling wind farm wake dynamics, can also be used in combination with these real WRF-LES simulations. This can further enable wind farm wake studies (Churchfield et al. 2012a; Dörenkämper et al. 2015; Lundquist et al. 2015) that rely on high resolution simulations and accurate atmospheric conditions to model accurately. Such wake studies can be used to improve optimal wind turbine configurations within wind farms, thereby improving wind harvesting and reducing the cost of wind energy.

5.3.4 Increased Model-Data Collaboration

The availability of the high-resolution HRDL lidar dataset from the NEAQS-ITCT 2004 campaign was useful in aiding the evaluation of simulations in this study. Only horizontal wind profiles were available for evaluation. The full extent of the LES simulations, such as high temporal resolution turbulence quantities could not be evaluated due to limitations of the measurement datasets. However, with advance-

ments in remote sensing sampling techniques, turbulence profiles, for example, can now be measured (Klein et al. 2015; Lundquist and Bariteau 2015). Comprehensive field campaigns such as the recent U.S. Dept. of Energy sponsored the eXperimental Planetary boundary layer Instrumentation Assessment (XPIA) campaign at the Boulder Atmospheric Observatory (Lundquist et al. 2016; NREL Technical Report, number pending). This used a multitude of remote sensing systems to characterize boundary layer processes, was also designed to measure relevant atmospheric data that could be used for subsequent numerical evaluations and simulations. The capabilities of these state-of-the-art systems can provide not only a more robust and nuanced view of the atmosphere but can be used for model validation and improvements. Closer collaborations between the experimental and modelling communities is particularly relevant for coastal and offshore regions where more observation datasets are needed for numerical model validation, for both atmospheric studies into coastal dynamics as well as the wind industry’s need for reliable annual energy production estimate improvements.

References

- Aitken, M. L., B. Kosović, J. D. Mirocha, and J. K. Lundquist, 2014: Large-eddy simulation of wind turbine wake dynamics in the stable boundary layer using the Weather Research and Forecasting model. *J. Renewable and Sustainable Energy*, **6** (3), 033 137.
- Aitken, M. L., and J. K. Lundquist, 2014: Utility-scale wind turbine wake characterization using nacelle-based long-range scanning lidar. *J. Atmos. Oceanic Technol.*, **31** (7), 1529–1539.
- Aitken, M. L., M. E. Rhodes, and J. K. Lundquist, 2012: Performance of a wind-profiling lidar in the region of wind turbine rotor disks. *J. Atmos. Oceanic Technol.*, **29** (3), 347–355.
- Angevine, W. M., M. Tjernström, and M. Žagar, 2006: Modeling of the coastal boundary layer and pollutant transport in New England. *J. Appl. Meteor. Climatol.*, **45** (1), 137–154.
- Archer, C. L., and Coauthors, 2014: Meteorology for coastal/offshore wind energy in the United States: recommendations and research needs for the next 10 years. *Bull. Amer. Meteor. Soc.*
- Badger, J., H. Frank, A. N. Hahmann, and G. Giebel, 2014: Wind-climate estimation based on mesoscale and microscale modeling: Statistical-dynamical downscaling for wind energy applications. *J. Appl. Meteor. Climatol.*, **53** (8), 1901–1919.
- Banta, R., R. Newsom, J. Lundquist, Y. Pichugina, R. Coulter, and L. Mahrt, 2002: Nocturnal low-level jet characteristics over Kansas during CASES-99. *Boundary-Layer Meteorology*, **105** (2), 221–252.

- Banta, R. M., Y. L. Pichugina, and W. A. Brewer, 2006: Turbulent velocity-variance profiles in the stable boundary layer generated by a nocturnal low-level jet. *J. Atmos. Sci.*, **63** (11), 2700–2719.
- Banta, R. M., and Coauthors, 2015: 3d volumetric analysis of wind turbine wake properties in the atmosphere using high-resolution doppler lidar. *Journal of Atmospheric and Oceanic Technology*, **32** (5), 904–914.
- Bardina, J., J. H. Ferziger, and W. Reynolds, 1980: Improved subgrid-scale models for large-eddy simulation. *American Institute of Aeronautics and Astronautics, Fluid and Plasma Dynamics Conference, 13th, Snowmass, Colo., July 14-16, 1980, 10 p.*, Vol. 1.
- Barry, R. G., and R. J. Chorley, 2009: *Atmosphere, weather and climate*. Routledge.
- Basu, S., and F. Porté-Agel, 2005: Large-eddy simulation of stably stratified atmospheric boundary layer turbulence: a scale-dependent dynamic modeling approach. *J. Atmos. Sci.*
- Basu, S., F. Porté-Agel, E. Foufoula-Georgiou, J.-F. Vinuesa, and M. Pahlow, 2006: Revisiting the local scaling hypothesis in stably stratified atmospheric boundary-layer turbulence: an integration of field and laboratory measurements with large-eddy simulations. *Boundary-Layer Meteorology*, **119** (3), 473–500.
- Batten, P., U. Goldberg, and S. Chakravarthy, 2004: Interfacing statistical turbulence closures with large-eddy simulation. *AIAA journal*, **42** (3), 485–492.
- Beare, R. J., and Coauthors, 2006: An intercomparison of large-eddy simulations of the stable boundary layer. *Boundary-Layer Meteorology*, **118** (2), 247–272.
- Beaucage, P., M. C. Brower, and J. Tensen, 2014: Evaluation of four numerical wind flow models for wind resource mapping. *Wind Energy*, **17** (2), 197–208.
- Berg, J., N. Vasiljević, M. Kelly, G. Lea, and M. Courtney, 2015a: Addressing spatial variability of surface-layer wind with long-range WindScanners. *Journal of Atmospheric and Oceanic Technology*, **32** (3), 518–527.

- Berg, L. K., L. D. Riihimaki, Y. Qian, H. Yan, and M. Huang, 2015b: The low-level jet over the Southern Great Plains determined from observations and reanalyses and its impact on moisture transport. *J. Climate*, **(2015)**.
- Bernier, N. B., and S. Blair, 2012: High horizontal and vertical resolution limited-area model: Near-surface and wind energy forecast applications. *J. Appl. Meteor. Climatol.*, **51 (6)**, 1061–1078.
- Blackadar, A. K., 1957: Boundary layer wind maxima and their significance for the growth of nocturnal inversions. *Bull. Amer. Meteor. Soc.*, **38 (5)**, 283–290.
- Bonner, W. D., 1968: Climatology of the low level jet. *Mon. Wea. Rev.*, **96 (12)**, 833–850.
- Bonner, W. D., and J. Paegle, 1970: Diurnal variations in boundary layer winds over the south-central United States in summer. *Mon. Wea. Rev.*, **98 (10)**, 735–744.
- Brower, M., 2012: *Wind resource assessment: a practical guide to developing a wind project*. John Wiley & Sons.
- Burk, S. D., and W. T. Thompson, 1996: The summertime low-level jet and marine boundary layer structure along the california coast. *Mon. Wea. Rev.*, **124 (4)**, 668.
- Businger, J. A., J. C. Wyngaard, Y. Izumi, and E. F. Bradley, 1971: Flux-profile relationships in the atmospheric surface layer. *J. Atmos. Sci.*, **28 (2)**, 181–189.
- Cape Wind, cited 2015: Cape Wind Project. [Available online at <http://www.capewind.org/>].
- Carvalho, D., A. Rocha, M. Gómez-Gesteira, and C. Santos, 2012: A sensitivity study of the wrf model in wind simulation for an area of high wind energy. *Environmental Modelling & Software*, **33**, 23–34.
- Chow, F. K., R. L. Street, M. Xue, and J. H. Ferziger, 2005: Explicit filtering and reconstruction turbulence modeling for large-eddy simulation of neutral boundary layer flow. *J. Atmos. Sci.*, **62 (7)**, 2058–2077.

- Churchfield, M., P. Moriarty, G. Vijayakumar, and J. Brasseur, 2010: Wind energy-related atmospheric boundary-layer large-eddy simulation using OpenFOAM. *National Renewable Energy Laboratory, Golden, CO, Report No. NREL/CP-500-48905*.
- Churchfield, M. J., S. Lee, J. Michalakes, and P. J. Moriarty, 2012a: A numerical study of the effects of atmospheric and wake turbulence on wind turbine dynamics. *Journal of turbulence*, **(13)**, N14.
- Churchfield, M. J., S. Lee, P. J. Moriarty, L. A. Martinez, S. Leonardi, G. Vijayakumar, and J. G. Brasseur, 2012b: A large-eddy simulation of wind-plant aerodynamics. *AIAA*.
- Colle, B. A., and D. R. Novak, 2010: The New York Bight jet: climatology and dynamical evolution. *Mon. Wea. Rev.*, **138 (6)**, 2385–2404.
- Davidson, L., 2011: Fluid mechanics, turbulent flow and turbulence modeling. *Chalmers University of Technology, Goteborg, Sweden (Nov 2011)*.
- Deardorff, J. W., 1970: A numerical study of three-dimensional turbulent channel flow at large reynolds numbers. *Journal of Fluid Mechanics*, **41 (02)**, 453–480.
- Dee, D., and Coauthors, 2011: The ERA Interim reanalysis: Configuration and performance of the data assimilation system. *Quart. J. Roy. Meteor. Soc.*, **137 (656)**, 553–597.
- Di Mare, L., M. Klein, W. Jones, and J. Janicka, 2006: Synthetic turbulence inflow conditions for large-eddy simulation. *Physics of Fluids (1994-present)*, **18 (2)**, 025 107.
- Dörenkämper, M., B. Witha, G. Steinfeld, D. Heinemann, and M. Kühn, 2015: The impact of stable atmospheric boundary layers on wind-turbine wakes within offshore wind farms. *Journal of Wind Engineering and Industrial Aerodynamics*, **144**, 146–153.
- Doyle, J. D., and T. T. Warner, 1993: A three-dimensional numerical investigation of a Carolina coastal low-level jet during GALE IOP 2. *Mon. Wea. Rev.*, **121 (4)**, 1030–1047.

- Du, Y., Y.-L. Chen, and Q. Zhang, 2015: Numerical simulations of the boundary layer jet off the southeastern coast of China. *Mon. Wea. Rev.*, **(2015)**.
- Du, Y., and R. Rotunno, 2014: A simple analytical model of the nocturnal low-level jet over the Great Plains of the United States. *J. Atmos. Sci.*, **71 (10)**, 3674–3683.
- Dudhia, J., 1989: Numerical study of convection observed during the winter monsoon experiment using a mesoscale two-dimensional model. *J. Atmos. Sci.*, **46 (20)**, 3077–3107.
- Dvorak, M. J., C. L. Archer, and M. Z. Jacobson, 2010: California offshore wind energy potential. *Renewable Energy*, **35 (6)**, 1244–1254.
- Edson, J., and Coauthors, 2007: The coupled boundary layers and airsea transfer experiment in low winds. *Bull. Amer. Meteor. Soc.*
- Fehsenfeld, F., and Coauthors, 2006: International consortium for atmospheric research on transport and transformation (ICARTT): North America to Europe overview of the 2004 summer field study. *Journal of Geophysical Research: Atmospheres*, **111 (D23)**.
- Fureby, C., G. Tabor, H. Weller, and A. Gosman, 1997: A comparative study of sub-grid scale models in homogeneous isotropic turbulence. *Physics of Fluids (1994-present)*, **9 (5)**, 1416–1429.
- Germano, M., U. Piomelli, P. Moin, and W. H. Cabot, 1991: A dynamic subgrid-scale eddy viscosity model. *Physics of Fluids A: Fluid Dynamics (1989-1993)*, **3 (7)**, 1760–1765.
- Ghosal, S., T. S. Lund, P. Moin, and K. Akselvoll, 1995: A dynamic localization model for large-eddy simulation of turbulent flows. *Journal of Fluid Mechanics*, **286**, 229–255.
- Gopalan, H., C. Gundling, K. Brown, B. Roget, J. Sitaraman, J. D. Mirocha, and W. O. Miller, 2014: A coupled mesoscale-microscale framework for wind resource estimation and farm aerodynamics. *J. Wind Eng. Ind. Aerodyn.*, **132**, 13–26.

- Grund, C. J., R. M. Banta, J. L. George, J. N. Howell, M. J. Post, R. A. Richter, and A. M. Weickmann, 2001: High-resolution doppler lidar for boundary layer and cloud research. *Journal of Atmospheric and Oceanic Technology*, **18** (3), 376–393.
- Gryning, S.-E., J. Badger, A. N. Hahmann, and E. Batchvarova, 2014: Current status and challenges in wind energy assessment. *Weather Matters for Energy*, Springer, 275–293.
- Hahmann, A. N., C. L. Vincent, A. Peña, J. Lange, and C. B. Hasager, 2014: Wind climate estimation using WRF model output: method and model sensitivities over the sea. *Int. J. Climatol.*
- Helmis, C., G. Sgouros, and Q. Wang, 2015: On the vertical structure and spectral characteristics of the marine low-level jet. *Atmos. Res.*, **152**, 74–81.
- Helmis, C., Q. Wang, G. Sgouros, S. Wang, and C. Halios, 2013: Investigating the summertime low-level jet over the east coast of the USA: A case study. *Boundary-layer meteorology*, **149** (2), 259–276.
- Högström, U., 1988: Non-dimensional wind and temperature profiles in the atmospheric surface layer: A re-evaluation. *Topics in Micrometeorology. A Festschrift for Arch Dyer*, Springer, 55–78.
- Holton, J. R., 1967: The diurnal boundary layer wind oscillation above sloping terrain. *Tellus*, **19** (2), 199–205.
- Holtslag, B., 2006: Preface: GEWEX atmospheric boundary-layer study (GABLS) on stable boundary layers. *Boundary-Layer Meteorology*, **118** (2), 243–246.
- Hong, S., 2010: A new stable boundarylayer mixing scheme and its impact on the simulated East Asian summer monsoon. *Quart. J. Roy. Meteor. Soc.*, **136** (651), 1481–1496.
- Hong, S.-Y., J. Dudhia, and S.-H. Chen, 2004: A revised approach to ice microphysical processes for the bulk parameterization of clouds and precipitation. *Mon. Wea. Rev.*, **132** (1), 103–120.
- Hong, S.-Y., Y. Noh, and J. Dudhia, 2006: A new vertical diffusion package with an explicit treatment of entrainment processes. *Mon. Wea. Rev.*, **134** (9), 2318–2341.

- Hu, X., P. M. Klein, and M. Xue, 2013a: Evaluation of the updated YSU planetary boundary layer scheme within WRF for wind resource and air quality assessments. *J. Geophys. Res.*, **118** (18), 10,490–10,505.
- Hu, X.-M., P. M. Klein, M. Xue, J. K. Lundquist, F. Zhang, and Y. Qi, 2013b: Impact of low-level jets on the nocturnal urban heat island intensity in Oklahoma City. *J. Appl. Meteor. Climatol.*, **52** (8), 1779–1802.
- Hu, X.-M., J. W. Nielsen-Gammon, and F. Zhang, 2010: Evaluation of three planetary boundary layer schemes in the WRF model. *J. Appl. Meteor. Climatol.*, **49** (9), 1831–1844.
- Huang, Y., and X. Peng, 2016: Improvement of the mellor–yamada–nakanishi–niino planetary boundary-layer scheme based on observational data in china. *Boundary-Layer Meteorology*, 1–18.
- Iungo, G. V., and F. Porté-Agel, 2014: Volumetric lidar scanning of wind turbine wakes under convective and neutral atmospheric stability regimes. *Journal of Atmospheric and Oceanic Technology*, **31** (10), 2035–2048.
- Iungo, G. V., Y.-T. Wu, and F. Porté-Agel, 2013: Field measurements of wind turbine wakes with lidars. *Journal of Atmospheric and Oceanic Technology*, **30** (2), 274–287.
- Jackson, P., and J. Hunt, 1975: Turbulent wind flow over a low hill. *Quart. J. Roy. Meteor. Soc.*, **101** (430), 929–955.
- Kaimal, J. C., and J. J. Finnigan, 1994: *Atmospheric boundary layer flows: their structure and measurement*. Oxford University Press.
- Kelley, N., M. Shirazi, D. Jager, S. Wilde, J. Adams, M. Buhl, P. Sullivan, and E. Patton, 2004: Lamar low-level jet project interim report. NCAR Tech. Rep. NREL/TP-500-34593.
- Khanna, S., and J. G. Brasseur, 1997: Analysis of Monin-Obukhov similarity from large-eddy simulation. *Journal of Fluid Mechanics*, **345**, 251–286.

- Klein, M., A. Sadiki, and J. Janicka, 2003: A digital filter based generation of inflow data for spatially developing direct numerical or large eddy simulations. *Journal of Computational Physics*, **186** (2), 652–665.
- Klein, P., and Coauthors, 2015: LABLE: a multi-institutional, student-led, atmospheric boundary layer experiment. *Bulletin of the American Meteorological Society*, **96** (10), 1743–1764.
- Koh, T., S. Wang, and B. C. Bhatt, 2012: A diagnostic suite to assess NWP performance. *J. Geophys. Res.*, **117** (D13).
- Kolmogorov, A. N., 1991: The local structure of turbulence in incompressible viscous fluid for very large reynolds numbers. *Proceedings: Mathematical and Physical Sciences*, **434** (1890), 9–13.
- Kosović, B., 1997: Subgrid-scale modelling for the large-eddy simulation of high-reynolds-number boundary layers. *Journal of Fluid Mechanics*, **336**, 151–182.
- Kosovic, B., and J. A. Curry, 2000: A large eddy simulation study of a quasi-steady, stably stratified atmospheric boundary layer. *J. Atmos. Sci.*, **57** (8), 1052–1068.
- Lilly, D. K., 1967: The representation of small scale turbulence in numerical simulation experiments.
- Lindborg, E., 2006: The energy cascade in a strongly stratified fluid. *J. Fluid Mech.*, **550**, 207–242.
- Little, C. G., 1969: Acoustic methods for the remote probing of the lower atmosphere. *Proceedings of the IEEE*, **57** (4), 571–578.
- Liu, Y., and Coauthors, 2011: Simultaneous nested modeling from the synoptic scale to the les scale for wind energy applications. *Journal of Wind Engineering and Industrial Aerodynamics*, **99** (4), 308–319.
- Lorenz, E. N., 1963: Deterministic nonperiodic flow. *J. Atmos. Sci.*, **20** (2), 130–141.
- Lundquist, J., and L. Bariteau, 2015: Dissipation of turbulence in the wake of a wind turbine. *Boundary-Layer Meteorology*, **154** (2), 229–241.

- Lundquist, J., M. Churchfield, S. Lee, and A. Clifton, 2015: Quantifying error of lidar and sodar doppler beam swinging measurements of wind turbine wakes using computational fluid dynamics. *Atmospheric Measurement Techniques*, **8** (2), 907–920.
- Lundquist, J. K., and J. D. Mirocha, 2008: Interaction of nocturnal low-level jets with urban geometries as seen in Joint Urban 2003 data. *J. Appl. Meteor. Climatol.*, **47** (1), 44–58.
- Mahoney, W. P., and Coauthors, 2012: A wind power forecasting system to optimize grid integration. *Sustainable Energy, IEEE Transactions on*, **3** (4), 670–682.
- Mahrt, L., D. Vickers, and E. L. Andreas, 2014: Low-level wind maxima and structure of the stably stratified boundary layer in the coastal zone. *J. Appl. Meteor. Climatol.*, **53** (2), 363–376.
- Mason, P., and S. Derbyshire, 1990: Large-eddy simulation of the stably-stratified atmospheric boundary layer. *Boundary-layer meteorology*, **53** (1-2), 117–162.
- Mason, P. J., 1989: Large-eddy simulation of the convective atmospheric boundary layer. *J. Atmos. Sci.*, **46** (11), 1492–1516.
- McAllister, L. G., J. R. Pollard, A. R. Mahoney, and P. J. Shaw, 1969: Acoustic soundings a new approach to the study of atmospheric structure. *Proceedings of the IEEE*, **57** (4), 579–587.
- Mesinger, F., and Coauthors, 2006: North American Regional Reanalysis. *Bull. Amer. Meteor. Soc.*, **87** (3), 343–360.
- Mirocha, J., B. Kosović, M. Aitken, and J. Lundquist, 2014a: Implementation of a generalized actuator disk wind turbine model into the weather research and forecasting model for large-eddy simulation applications. *J. Renewable and Sustainable Energy*, **6** (1), 013 104.
- Mirocha, J., B. Kosović, B., and G. Kirkil, 2014b: Resolved turbulence characteristics in large-eddy simulations nested within mesoscale simulations using the weather research and forecasting model. *Mon. Wea. Rev.*, **142** (2), 806–831.

- Mirocha, J., J. Lundquist, and B. Kosovic, 2010: Implementation of a nonlinear sub-filter turbulence stress model for large-eddy simulation in the advanced research wrf model. *Mon. Wea. Rev.*, **138** (11), 4212–4228.
- Mirocha, J. D., D. A. Rajewski, N. Marjanovic, J. K. Lundquist, B. Kosović, C. Draxl, and M. J. Churchfield, 2015: Investigating wind turbine impacts on near-wake flow using profiling lidar data and large-eddy simulations with an actuator disk model. *J. Renewable and Sustainable Energy*, **7** (4), 043 143.
- Mlawer, E. J., S. J. Taubman, P. D. Brown, M. J. Iacono, and S. A. Clough, 1997: Radiative transfer for inhomogeneous atmospheres: RRTM, a validated correlated-k model for the longwave. *J. Geophys. Res.*, **102** (D14), 16 663–16 682.
- Moeng, C., J. Dudhia, J. Klemp, and P. Sullivan, 2007: Examining two-way grid nesting for large eddy simulation of the PBL using the WRF model. *Mon. Wea. Rev.*, **135** (6), 2295–2311.
- Moeng, C.-H., and P. P. Sullivan, 1994: A comparison of shear-and buoyancy-driven planetary boundary layer flows. *J. Atmos. Sci.*, **51** (7), 999–1022.
- Moeng, C.-H., and J. C. Wyngaard, 1988: Spectral analysis of large-eddy simulations of the convective boundary layer. *J. Atmos. Sci.*, **45** (23), 3573–3587.
- Monin, A., and A. Obukhov, 1954: Basic laws of turbulent mixing in the surface layer of the atmosphere. *Contrib. Geophys. Inst. Acad. Sci. USSR*, **151**, 163–187.
- Muñoz-Esparza, D., B. Kosović, J. Mirocha, and J. van Beeck, 2014: Bridging the transition from mesoscale to microscale turbulence in numerical weather prediction models. *Boundary-Layer Meteorology*, **153** (3), 409–440.
- Muñoz-Esparza, D., B. Kosović, J. van Beeck, and J. Mirocha, 2015: A stochastic perturbation method to generate inflow turbulence in large-eddy simulation models: Application to neutrally stratified atmospheric boundary layers. *Physics of Fluids (1994-present)*, **27** (3), 035 102.
- Nakanish, M., 2001: Improvement of the mellor–yamada turbulence closure model based on large-eddy simulation data. *Boundary-layer meteorology*, **99** (3), 349–378.

- Nakanishi, M., and H. Niino, 2004: An improved Mellor-Yamada level-3 model with condensation physics: Its design and verification. *Boundary-layer meteorology*, **112** (1), 1–31.
- NCEP, NWS, NOAA, and U.S. Department of Commerce, 2000: NCEP FNL Operational Model Global Tropospheric Analyses, continuing from July 1999. Research Data Archive at the National Center for Atmospheric Research, Computational and Information Systems Laboratory, Boulder CO, URL <http://dx.doi.org/10.5065/D6M043C6>.
- Newsom, R., R. Calhoun, D. Ligon, and J. Allwine, 2008: Linearly organized turbulence structures observed over a suburban area by dual-doppler lidar. *Boundary-layer meteorology*, **127** (1), 111–130.
- Nielsen-Gammon, J. W., X.-M. Hu, F. Zhang, and J. E. Pleim, 2010: Evaluation of planetary boundary layer scheme sensitivities for the purpose of parameter estimation. *Mon. Wea. Rev.*, **138** (9), 3400–3417.
- Nielsen-Gammon, J. W., and Coauthors, 2008: Multisensor estimation of mixing heights over a coastal city. *Journal of Applied Meteorology and Climatology*, **47** (1), 27–43.
- Nieuwstadt, F. T., P. J. Mason, C.-H. Moeng, and U. Schumann, 1993: Large-eddy simulation of the convective boundary layer: A comparison of four computer codes. *Turbulent Shear Flows 8*, Springer, 343–367.
- NOAA, 1998: *Automated Surface Observing System (User Guide)*. Available online at <http://www.nws.noaa.gov/asos/pdfs/aum-toc.pdf>.
- Nunalee, C. G., and S. Basu, 2014: Mesoscale modeling of coastal lowlevel jets: implications for offshore wind resource estimation. *Wind Energy*, **17** (8), 1199–1216.
- Obukhov, A., 1971: Turbulence in an atmosphere with a non-uniform temperature. *Boundary-layer meteorology*, **2** (1), 7–29.
- Olson, J. B., and J. Brown, 2009: A comparison of two MellorYamada-based PBL schemes in simulating a hybrid barrier jet. *23rd Conference on weather anal-*

ysis and forecasting/19th conference on numerical weather prediction, Omaha.
<http://ams.confex.com/ams/pdfpapers/154321.pdf>.

- Ostdiek, V., and W. Blumen, 1997: A dynamic trio: Inertial oscillation, deformation frontogenesis, and the Ekman-Taylor boundary layer. *J. Atmos. Sci.*, **54** (11), 1490–1502.
- Peyret, R., and E. Krause, 2000: *Advanced turbulent flow computations*. Springer.
- Pichugina, Y. L., R. M. Banta, W. A. Brewer, S. P. Sandberg, and R. M. Hardesty, 2012: Doppler lidar-based wind-profile measurement system for offshore wind-energy and other marine boundary layer applications. *Journal of Applied Meteorology and Climatology*, **51** (2), 327–349.
- Pichugina, Y. L., S. C. Tucker, R. M. Banta, W. A. Brewer, N. D. Kelley, B. J. Jonkman, and R. K. Newsom, 2008: Horizontal-velocity and variance measurements in the stable boundary layer using doppler lidar: Sensitivity to averaging procedures. *Journal of Atmospheric and Oceanic Technology*, **25** (8), 1307–1327.
- Piomelli, U., 1999: Large-eddy simulation: achievements and challenges. *Progress in Aerospace Sciences*, **35** (4), 335–362.
- Porté-Agel, F., Y.-T. Wu, H. Lu, and R. J. Conzemius, 2011: Large-eddy simulation of atmospheric boundary layer flow through wind turbines and wind farms. *Journal of Wind Engineering and Industrial Aerodynamics*, **99** (4), 154–168.
- Rareshide, E., A. Tindal, C. Johnson, A. Graves, E. Simpson, J. Bleeg, T. Harris, and D. Schoborg, 2009: Effects of complex wind regimes on turbine performance. *Scientific Proceedings. American Wind Energy Association WINDPOWER Conference*.
- Rife, D. L., J. O. Pinto, A. J. Monaghan, C. A. Davis, and J. R. Hannan, 2010: Global distribution and characteristics of diurnally varying low-level jets. *J. Climate*, **23** (19), 5041–5064.
- Shin, H. H., and S.-Y. Hong, 2011: Intercomparison of planetary boundary-layer parametrizations in the WRF model for a single day from CASES-99. *Boundary-Layer Meteorology*, **139** (2), 261–281.

- Skamarock, W., and Coauthors, 2008: A description of the Advanced Research WRF version 3. NCAR Tech. Report, Note NCAR/TN-4751STR.
- Smagorinsky, J., 1963: General circulation experiments with the primitive equations: I. the basic experiment*. *Mon. Wea. Rev.*, **91** (3), 99–164.
- Smedman, A., H. Bergström, and B. Grisogono, 1997: Evolution of stable internal boundary layers over a cold sea. *J. Geophys. Res.*, **102** (C1), 1091–1099.
- Smedman, A.-S., H. Bergström, and U. Högström, 1995: Spectra, variances and length scales in a marine stable boundary layer dominated by a low level jet. *Boundary-Layer Meteorology*, **76** (3), 211–232.
- Song, J., K. Liao, R. L. Coulter, and B. M. Lesht, 2005: Climatology of the low-level jet at the Southern Great Plains atmospheric boundary layer experiments site. *J. Appl. Meteor.*, **44** (10), 1593–1606.
- Steele, C., S. Dorling, R. Von Glasow, and J. Bacon, 2013: Idealized wrf model sensitivity simulations of sea breeze types and their effects on offshore windfields. *Atmos. Chem. Phys.*, **13** (1), 443–461.
- Stevens, R. J., D. F. Gayme, and C. Meneveau, 2014: Large eddy simulation studies of the effects of alignment and wind farm length. *Journal of renewable and sustainable energy*, **6** (2), 023 105.
- Stoll, R., and F. Porté-Agel, 2006: Dynamic subgrid-scale models for momentum and scalar fluxes in large-eddy simulations of neutrally stratified atmospheric boundary layers over heterogeneous terrain. *Water Resources Research*, **42** (1).
- Storm, B., J. Dudhia, S. Basu, A. Swift, and I. Giammanco, 2009: Evaluation of the weather research and forecasting model on forecasting low-level jets: Implications for wind energy. *Wind Energy*, **12** (1), 81–90.
- Talbot, C., E. Bou-Zeid, and J. Smith, 2012: Nested mesoscale large-eddy simulations with WRF: performance in real test cases. *J. Hydrometeor.*, **13** (5), 1421–1441.

- Taylor, K. E., 2000: *Summarizing multiple aspects of model performance in a single diagram*. Program for Climate Model Diagnosis and Intercomparison, Lawrence Livermore National Laboratory, University of California.
- Tennekes, H., and J. L. Lumley, 1972: *A first course in turbulence*. MIT press.
- Troen, I., and L. Mahrt, 1986: A simple model of the atmospheric boundary layer; sensitivity to surface evaporation. *Boundary-Layer Meteorology*, **37** (1-2), 129–148.
- Van de Wiel, B. J., A. Moene, G. Steeneveld, P. Baas, F. Bosveld, and A. Holtslag, 2010: A conceptual view on inertial oscillations and nocturnal low-level jets. *J. Atmos. Sci.*, **67** (8), 2679–2689.
- Vanderwende, B., and J. K. Lundquist, 2012: The modification of wind turbine performance by statistically distinct atmospheric regimes. *Environ. Res. Lett.*, **7** (3), 034 035.
- Vanderwende, B. J., J. K. Lundquist, M. E. Rhodes, E. S. Takle, and S. L. Irvin, 2015: Observing and simulating the summertime low-level jet in central Iowa. *Mon. Wea. Rev.*, (2015).
- Versteeg, H. K., and W. Malalasekera, 2007: *An introduction to computational fluid dynamics: the finite volume method*. Pearson Education.
- Vestas, cited 2015: V100 2.6 MW. [Available online at <http://nozebra.ipapercms.dk/Vestas/Communication/Productbrochure/V10026MW/V10026MW/>].
- Vincent, C. L., A. N. Hahmann, J. Badger, X. G. Larsén, and S. E. Larsen, 2011: Off-shore mesoscale variability. *EWEA Wind Resource Assessment Technology Workshop*.
- Wagner, R., M. Courtney, J. Gottschall, and P. LindelwMarsden, 2011: Accounting for the speed shear in wind turbine power performance measurement. *Wind Energy*, **14** (8), 993–1004.
- Whiteman, C. D., X. Bian, and S. Zhong, 1997: Low-level jet climatology from enhanced rawinsonde observations at a site in the Southern Great Plains. *J. Appl. Meteor.*, **36** (10), 1363–1376.

- Wilczak, J., and Coauthors, 2014: The wind forecast improvement project (WFIP): A public-private partnership addressing wind energy forecast needs. *Bull. Amer. Meteor. Soc.*
- Wolfe, D., and Coauthors, 2007: Shipboard multisensor merged wind profiles from the New England Air Quality Study 2004. *JGR: Atmospheres (1984–2012)*, **112** (D10).
- Wyngaard, J. C., 2004: Toward numerical modeling in the Terra Incognita. *J. Atmos. Sci.*, **61** (14), 1816–1826.
- Wyngaard, J. C., 2010: *Turbulence in the Atmosphere*. Cambridge University Press.
- Xie, Z.-T., and I. P. Castro, 2008: Efficient generation of inflow conditions for large eddy simulation of street-scale flows. *Flow, turbulence and combustion*, **81** (3), 449–470.
- Yang, D., C. Meneveau, and L. Shen, 2014: Large-eddy simulation of offshore wind farm. *Physics of Fluids (1994-present)*, **26** (2), 025 101.
- Zhou, B., and F. K. Chow, 2014: Nested large-eddy simulations of the intermittently turbulent stable atmospheric boundary layer over real terrain. *J. Atmos. Sci.*, **71** (3), 1021–1039.