On the application of a numerical model to simulate the coastal boundary layer

Nissen, Jesper Nielsen

Publication date: 2008

Document Version
Publisher's PDF, also known as Version of record

Link back to DTU Orbit

Citation (APA):
PhD thesis
Jesper Nielsen Nissen

On the application of a numerical model to simulate the coastal boundary layer

Risø-PhD-39(EN)

Academic advisor: Aksel Walløe Hansen
Submitted: 21/07/08
This work aims to study the seasonal difference in normalized wind speed above the surface layer as it is observed at the 160 m high mast at the coastal site Høvsøre at winds from the sea (westerly).

Normalized wind speeds above the surface layer are observed to be 20 to 50% larger in the winter/spring seasons compared to the summer/autumn seasons at winds from west within the same atmospheric stability class.

A method combining the mesoscale model, COAMPS, and observations of the surface stability of the marine boundary layer is presented. The objective of the method is to reconstruct the seasonal signal in wind speed and identify the physical process behind. The method proved reasonably successful in capturing the relative difference in wind speed between seasons, indicating that the simulated physical processes are likely candidates to the observed seasonal signal in normalized wind speed.

The lower part of the seasonal normalized wind speed profiles were also captured reasonably well. However, the method consistently over-predict the absolute values of the normalized wind speeds at the upper part of the profile and suggestions to improve the skills of the method in this region are discussed.

The winds from west at Høvsøre also showed an increased upper level variance of the wind speed during spring and winter when compared to summer and autumn.

It is shown that excess in temperature over England relative to the North Sea is a player because it triggers wind speed oscillations in the boundary layer over the North Sea. The oscillations were found to introduce up to 20% departure in the simulated normalized wind speed at 100 m height, compared to simulations where no upstream land was accounted for or situations where the upstream land was colder than the North Sea.

The signal was found to be stronger during spring and winter as compared to summer and autumn and serves as indicator for the more complex nature of boundary layer processes during winter and spring compared to summer and autumn.
1 Theory and Background 5
1.1 Introduction 5
1.2 Atmospheric internal boundary layers 6
1.2.1 Mathematical and physical considerations on boundary layer growth 7
1.3 Marine boundary layer meteorology 13
1.4 Coastal internal boundary layer 18
1.4.1 The conceptual model for the CIBL 19
1.5 Overview of IBL models 20
1.5.1 Empirical models 20
1.5.2 Similarity models 22
1.5.3 Physical slab models 22
1.5.4 NWP model 27

2 Measurements and site 35
2.1 Høvsøre 35
2.1.1 Data extraction 36
2.2 Horns Rev 37
2.3 Stability 38
2.4 Phi function at Høvsøre 42
2.4.1 Discussion on $\phi_{\text{west}}$ versus $\phi_{\text{east}}$ 43
2.5 Seasonality in over- and under- speeding of the wind 45
2.6 Seasonal change of the normalized wind profiles at Høvsøre 47
2.7 Seasonal change in the upper variance 48
2.8 Logarithmic shape of the mean profiles 49

3 Numerical model experiment 50
3.1 Model details shared for all simulations 52
3.2 Marine atmospheric boundary layer over a homogeneous sea surface 52
3.2.1 Model setup 53
3.2.2 Choice of initial conditions 53
3.2.3 Model results 54
3.3 Boundary layer structure: Wind speed 54
3.3.1 Boundary layer structure: Air and sea temperature difference 57
3.3.2 Boundary layer structure: Evolution in time 57
3.3.3 Summary: MABL simulation over a homogeneous sea surface 58
3.4 Boundary layer structure and an upstream coastline 60
3.4.1 Model setup 61
3.4.2 Choice of initial conditions 62
3.4.3 Spring runs 63
3.4.4 Boundary layer processes over a cold North Sea 66
3.4.5 Boundary layer processes over a cold North Sea simulated 67
3.4.6 Cross section comparison for spring runs 68
3.4.7 Profile comparison of TKE, wind, and lapse rate, spring case 70
3.4.8 Summer Spring Runs 71
3.4.9 Autumn runs 72
3.4.10 Cross section comparison for autumns runs 74
3.4.11 Profile comparison of TKE, wind, and lapse rate, Autumn case 75
3.5 Simulating costal Internal boundary layer builds up at Høvsøre. 79
3.5.1 Introduction 79
<table>
<thead>
<tr>
<th>Section</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>3.5.2 Simplification and method</td>
<td>80</td>
</tr>
<tr>
<td>3.5.3 Seasonal weighting</td>
<td>81</td>
</tr>
<tr>
<td>3.5.4 CIBL at Høvsøre</td>
<td>83</td>
</tr>
<tr>
<td>3.5.5 CIBL Simulation details</td>
<td>84</td>
</tr>
<tr>
<td>3.5.6 Model skills: Wind profile prediction</td>
<td>86</td>
</tr>
<tr>
<td>3.5.7 Observed and simulated height of the internal boundary layer</td>
<td>91</td>
</tr>
<tr>
<td>3.5.8 Model Skill: TKE profile prediction</td>
<td>92</td>
</tr>
<tr>
<td>4 Conclusions, outlook and future improvements</td>
<td>94</td>
</tr>
<tr>
<td>5 Acknowledgement</td>
<td>96</td>
</tr>
<tr>
<td>6 Some acronyms and definitions</td>
<td>97</td>
</tr>
<tr>
<td>7 References</td>
<td>98</td>
</tr>
<tr>
<td>8 Appendix</td>
<td>103</td>
</tr>
<tr>
<td>8.1 Observations winter</td>
<td>103</td>
</tr>
<tr>
<td>8.2 Observations spring</td>
<td>105</td>
</tr>
<tr>
<td>8.3 Observations summer</td>
<td>107</td>
</tr>
<tr>
<td>8.4 Observations autumn</td>
<td>109</td>
</tr>
</tbody>
</table>
1 Theory and Background

1.1 Introduction
An increasing demand for the ability to estimate the wind climate above the surface layer in coastal regions has been generated by the wind power industry because of the steady increase of modern wind turbines with hub heights situated well above the surface layer height. To choose the optimal placement for a wind turbine, information about the wind energy potential needs to be properly assessed. The wind energy potential is proportional to cube of the wind speed at hub height, implying that even a small error in estimating the wind speed can have a large impact on the wind energy potential assessment. Conventional methods for assessing wind speed at heights of up to 200 m, such as that of Troen and Lundtang (1989), are derived from wind measurement taken close to the surface and extrapolated to greater heights by assuming a logarithmic increase of the wind speed with height, with a correction for stability effects.

\[ U(z) = u_* \left[ \frac{\ln(z/z_0)}{k} + \psi \left( \frac{z}{L} \right) \right] \]

where
- \( z \) = height, \( z_0 \) = roughness length, \( u_* \) = friction velocity
- \( \kappa \) = von Karman constant = 0.4
- \( \psi \) = empirical stability function following Businger (1973); Dyer (1974) and \( L \) = Monin-Obukov length

The derivation of the logarithmic wind profile assumes homogeneous surface properties and a constant momentum flux and is therefore confined to the surface layer. Little is known about boundary layer wind profiles above the surface layer, but recent attempts to address this problem have been put forward by Högström et al. (2006) and Gryning et al. (2007), where extension to the Monin-Obukhov surface similarity scaling are suggested with some success.

In a coastal zone, the description of the vertical wind shear is complicated by non-homogenous upstream surface conditions, orography, change of roughness and heat capacity, all of which influence the vertical wind shear and are difficult to describe using simple, universal formulas. An alternative approach is adapted in this study where a well-validated mesoscale model is used to calculate the wind profile in the coastal region, resolving what is considered to be the main upstream physical processes. An observational study is performed in order to quantify how the upstream conditions influence the vertical wind shear close to the coastline, and the numerical simulations and observations are brought together to test the ability of the model to predict the vertical wind shear up to a height of 160 m at a distance of 1500 m downstream from a coastline.

The observational basis for this study comes from the Høvsøre test station for large wind turbines, located on the west coast of Jutland, Denmark, in order to explore the forcing in the upper boundary layer that a modern wind turbine is exposed to. The focus is on winds from the sea where a systematic over-speeding, compared to the traditional logarithmic wind profile, is observed at the top of the surface layer in the winter and spring months and likewise, a systematic under-speeding is found during summer and autumn. As this flow pattern is seasonally dependant, it is considered that heat exchange between the North Sea and the lower part of the
atmospheric boundary layer, together with a low inversion height and possibly numerous low-level jets, play a major role in the physical processes behind the observed structure of the wind profile. These processes are not accounted for in conventional wind energy potential assessment methods and the question posed here is:

Is a numerical model setup with a reasonable vertical resolution in the boundary layer able to reproduce the observed wind profiles and their seasonal variability, when only a limited number of physical surface processes are accounted for?

1.2 Atmospheric internal boundary layers

The concept of internal boundary layers is chosen to address the problem of how the vertical profile of wind is modified in the littoral region.

In the atmosphere, an internal boundary layer is generated when the air column is advected over a step change in surface properties. Downstream from the change, mechanically and thermally driven turbulence generates a new boundary layer inside the existing one and is therefore called internal.

The basic concept of an internal boundary layer is given in a schematic way in Figure 1, where the incoming air is assumed to be neutrally stratified and confined to the surface layer, making the associated wind profiles logarithmic. The situation outlined in Figure 1 is related to the situation on a coastline where the wind is blowing onshore from the smooth water surface to the rougher land surface. After the air column has passed over the new surface, the wind speed at the lowest level decreases due to the increased frictional force created by the new and higher level of momentum flux divergence, while the wind speed above the height \( \delta \) is unchanged, where \( \delta \) denotes the height of the internal boundary layer at the fetch distance \( x \).

For \( z > \delta \), upstream flow properties are found, while for \( z < \delta \), a distinction is made between the lower equilibrium layer and the upper transition layer. In the equilibrium layer, the atmosphere has adjusted to the new surface and this layer is defined in terms of having a 90% level of adjustment in the momentum flux \( \text{Garratt (1990)} \), while in the transition zone layer the atmosphere is affected by the new surface, but has not yet reached an equilibrium state.

It is common to estimate the height of the internal boundary layer as a function of downstream fetch, and a practical rule of thumb tells that the internal boundary layer grows as a power law of the fetch distance \( \text{Pasquill (1972)}; \text{Smedman and Högström (1978)}; \text{Bergström (1988)}; \text{Rao et al. (1974)} \). The power-law dependence on fetch is the subject of the following sections.
1.2.1 Mathematical and physical considerations on boundary layer growth

The growth of the boundary layer height expressed as a power law of downstream fetch can be derived from the governing equation for the atmospheric flow when the following definition of the boundary layer is made:

The atmospheric boundary layer is the lower region of the atmosphere where the velocity gradient normal to the ground is large and where the associated vertical stress divergence has an important influence on the flow.

Following Schlichting (1968), a power law of downstream fetch can be derived from the equations of motion in the $x$ and $z$ direction for an atmosphere moving initially with a mean horizontal velocity described by a constant velocity profile, $U(z)=U_0$. The moving atmosphere encounters a solid horizontal plate, and deceleration of the mean velocity is caused by an opposing force, denoted as the vertical stress divergence, acting in a shallow area above the plate as outlined in Figure 2. The vertical stress is caused by the internal friction of the atmosphere and dictates that the atmosphere has to come to a stop just above the plate; this is known as the no-slip boundary condition.

The vertical distance to which deceleration of the mean wind speed takes place is the height of the boundary layer according to the definition above. The situation is outlined in figure 2.
Figure 2: Downstream boundary layer development over a solid plate

The equations of motion in x and z direction, when the Coriolis force is neglected, are as follows

$$\frac{\partial u}{\partial t} + u \frac{\partial u}{\partial x} + w \frac{\partial u}{\partial z} = -\frac{\partial p}{\partial x} + \nu \left( \frac{\partial^2 u}{\partial x^2} + \frac{\partial^2 u}{\partial z^2} \right)$$

$$\frac{\partial w}{\partial t} + u \frac{\partial w}{\partial x} + w \frac{\partial w}{\partial z} = -\frac{\partial p}{\partial z} - g + \nu \left( \frac{\partial^2 w}{\partial x^2} + \frac{\partial^2 w}{\partial z^2} \right)$$

where \( t \) is time, \( u \) is wind speed in x direction, \( p \) is pressure, and \( x \) and \( z \) are the horizontal and vertical coordinates, respectively.

The kinematic viscosity, \( \nu \), is defined from the viscosity, \( \mu \), and the density of the fluid, \( \rho \), and denotes the constant of proportionality between the velocity shear and the stress:

$$\nu = \frac{\mu}{\rho}$$

$$\tau_x = \nu \left( \frac{\partial u}{\partial z} + \frac{\partial u}{\partial x} \right)$$

\( \tau_x \) = Stress in x direction

The incompressible version of the equation of continuity reads

$$\frac{\partial u}{\partial x} + \frac{\partial w}{\partial z} = 0$$
Following Schlichting (1968), the equations form the basis of scale analysis of each individual term carried out in the following way. Velocities are normalized by the free atmosphere velocity \( U_{\text{free-atmosphere}} \) and the length dimensions \( x, z \) are normalized by the horizontal length scale \( L_{\text{horizontal}} \) defined in such a way that the dimensionless derivatives in the \( x \) direction do not exceed unity:

\[
\frac{\partial u}{\partial x} : \frac{\partial \left( \frac{u}{U_{\text{free-atmosphere}}} \right)}{\partial \left( \frac{x}{L_{\text{horizontal}}} \right)} \leq 1
\]

\[
\frac{u}{U_{\text{free-atmosphere}}} \frac{\partial u}{\partial x} \cdot \frac{\partial \left( \frac{u}{U_{\text{free-atmosphere}}} \right)}{\partial \left( \frac{x}{L_{\text{horizontal}}} \right)} \leq 1
\]

The height of the boundary layer is the vertical distance up to which the perturbations to the mean velocity that are generated by the plate are felt by the flow; this distance is denoted \( h_{\text{boundary layer}} \). The height of the boundary layer is normalized by the horizontal distance \( L_{\text{horizontal}} \) and this ratio assumed small:

\[
\delta = \frac{h_{\text{boundary layer}}}{L_{\text{horizontal}}} \ll 1
\]

Schlichting completed a scale analysis of the individual terms in the normalized Navier-Stokes equations for a boundary layer, starting with the incompressible version of the equation of continuity. The first term is equal to one according to the definition of \( L_{\text{horizontal}} \) and accordingly the magnitude of the second term must be one also in order to satisfy the equation of continuity.

\[
\frac{\partial u_{\text{normalized}}}{\partial x_{\text{normalized}}} + \frac{\partial w_{\text{normalized}}}{\partial \delta} = 0
\]

Temporal changes to the wind speed are assumed to be of the order of one, ruling out situations with large accelerations, which are uncommon in the atmosphere.

\[
\frac{\partial u_{\text{normalized}}}{\partial t} \approx 1
\]

Schlichting argued from the equation of continuity that the second horizontal derivative of \( u \) must be close to unity also:
The vertical shear of the normalized $u$ can be estimated, recalling that $u$ is equal to zero at the ground, is equal to one above the boundary layer, and from the argumentation from Schlichting the second derivative follows as

\[
\frac{\partial u_{\text{normalized}}}{\partial \delta} \leq 1 \Rightarrow \frac{\partial^2 u_{\text{normalized}}}{\partial \delta^2} \leq 1
\]

The order of magnitude of the individual terms in the governing equations can, based on the above considerations, be estimated as follows

\[
\frac{\partial u_{\text{normalized}}}{\partial t} + u_{\text{normalized}} \frac{\partial u_{\text{normalized}}}{\partial x} + v_{\text{normalized}} \frac{\partial u_{\text{normalized}}}{\partial z} = \frac{1}{\delta}
\]

\[
\frac{\partial p}{\partial x} + \frac{1}{R_e} \left( \frac{\partial^2 u_{\text{normalized}}}{\partial x^2} + \frac{\partial^2 u_{\text{normalized}}}{\partial z^2} \right) \approx \frac{1}{\delta^2}
\]

The Reynolds number for atmospheric flow is defined as

\[
R_e = \frac{U_{\text{Free-atmosphere}}}{\nu}
\]

The Reynolds number for atmospheric flow is very large due to the low value of kinematic viscosity for atmospheric air, and the pressure gradient in the $x$ direction is assumed to be very small, leaving 1.5 with a left-hand side with terms of the order of 1. In order to balance the acceleration and advection terms on the left-hand side, the right-hand side must be of the same order of magnitude:

\[
\frac{1}{R_e} \left( \frac{\partial^2 u_{\text{normalized}}}{\partial x^2} + \frac{\partial^2 u_{\text{normalized}}}{\partial z^2} \right) \approx 1
\]

\[
\frac{1}{R_e} \frac{\partial^2 u_{\text{normalized}}}{\partial x^2} + \frac{1}{R_e} \frac{\partial^2 u_{\text{normalized}}}{\partial z^2} \approx 1 \Rightarrow \frac{1}{R_e} \frac{\partial^2 u_{\text{normalized}}}{\partial \delta^2} \approx 1
\]

where the small product

\[
\frac{1}{R_e} \frac{\partial^2 u_{\text{normalized}}}{\partial x^2} \ll 1 \approx 0
\]

The internal boundary layer considerations above lead to a criterion for the size of the Reynolds number in terms of the normalized vertical height, and a formula for the internal boundary layer height can be derived accordingly:
It was found after a scale analysis of terms that the vertical stress divergence term can only become of the same order of magnitude as the acceleration and advection terms for Reynolds numbers equal to the square of the dimensionless boundary layer height. These considerations lead to a diagnostic formula for the boundary layer and then also the internal boundary layer height, as a power law dependency of the fetch distance and kinematic viscosity, assuming that the boundary layer is the vertical part of the atmosphere where the vertical stress divergence generated by viscosity balances the advection and acceleration terms on the right-hand side.

The situation discussed is somewhat artificial and oversimplified and can only be expected to have a limited resemblance with real-life internal boundary layers. Nevertheless, the discussion leads to a power law dependency on downwind fetch distance that numerous observational studies have validated, indicating that the major physical player is accounted for.

However, when one applies representative values for normal atmospheric conditions into equation 1.6

\[ \nu = 1.4607 \times 10^{-5} \, \text{m}^2 / \text{s} \]

\[ U_{\text{free-atmosphere}} = 10 \, \text{m} / \text{s} \]

\[ L_{\text{horizontal}} = 10000 \, \text{m} \]

an internal boundary layer height of 10 cm is found. 1.6 dramatically under-predicts the observed internal boundary layer height, but is a reasonable estimate of the micro layer, which is the shallow layer between the horizontal surface and the atmosphere where the stress generated by viscosity dominates the flow.

The atmospheric internal boundary layer is found to be several orders of magnitude larger than predicted by 1.6. The reason for this is that internal boundary layer flows are characterised by turbulent motions giving rise to another stress term, denoted the Reynolds stress and defined as

\[ \frac{\tau_{x, \text{sublayer}}}{\rho} = \overline{u'w'} = u_*^2 \]

where it has been assumed that the flow can be characterised by a mean part and a fluctuating part

\[ u = \bar{u} + u' \quad w = \bar{w} + w' \]

\( u_*^2 \) is the vertical momentum flux found as the statistical covariance between \( u' \) and \( w' \).

In analogy with the definition of kinematic viscosity in 1.3, an eddy viscosity is defined in 1.9 using a first-order closure assumption.

\[ \frac{\tau_{x, \text{sublayer}}}{\rho} = \overline{u'w'} \approx K_m \frac{\partial \bar{u}}{\partial z} \]
The eddy viscosity is a property of the flow rather than a property of the fluid. The vertical divergence of the Reynolds stress dominates by far in boundary layer flows, as it is 4-5 orders of magnitude larger than the kinematic viscosity everywhere except from the lowest 10 cm, as previously described. The equations of motion 1.2 can now be rewritten neglecting the influence of stress generated by kinematic viscosity and assuming that the eddy viscosity defined in 1.9 is constant with height.

\[
\frac{\partial u}{\partial t} + u \frac{\partial u}{\partial x} + w \frac{\partial u}{\partial z} = -\frac{\partial p}{\partial x} + K_m \left( \frac{\partial^2 u}{\partial x^2} + \frac{\partial^2 u}{\partial z^2} \right)
\]

\[
\frac{\partial w}{\partial t} + u \frac{\partial w}{\partial x} + w \frac{\partial w}{\partial z} = -\frac{\partial p}{\partial z} - g + K_m \left( \frac{\partial^2 w}{\partial x^2} + \frac{\partial^2 w}{\partial z^2} \right)
\]

Repeating the scale argumentations from Schlichting (1968) described above and yielding equilibrium between the advection and acceleration terms on the left-hand side, and the Reynolds stress divergence on the right-hand side, leads to:

\[
\frac{U_{\text{free-atmosphere}} \cdot L_{\text{horizontal}}}{K_m} = \delta^2 \Rightarrow h_{\text{boundary layer}} = \sqrt{\frac{K_m \cdot L_{\text{horizontal}}}{U_{\text{free-atmosphere}}}}
\]

The application of equation 1.11 yields predictions of the internal boundary layer height in the 10 m to 1000 m range when inserting representative values for atmospheric flow, underlining that turbulence is the main process behind the generation of the vertical stress divergence and therefore for the growth of the internal boundary layer.

In real atmospheric flow situations, inland generation of internal boundary layers are taking place everywhere all the time, when air flows over a land surface with frequently changing landscapes and therefore surface properties. For example, surface properties representative for a city, a forest and rural landscape generate separate sets of surface fluxes of momentum, heat and moisture and therefore separate associated internal boundary layers, as depicted in Figure 3.

Figure 3: Boundary layers over a complex surface
Also minor changes in land use cause changes to the surface properties and associated heat and momentum fluxes, leading to the development of new internal boundary layers. The heterogeneous nature of inland surface properties makes most inland boundary layers a complex set of numerous internal boundary layers inside each other, reflecting upstream heterogeneity.

The coastline is a somewhat simpler place for the study of internal boundary layer, as the landscape upstream is, to a first order approximation, homogeneous and the changes of the turbulence and flow characteristics downstream from the coastline therefore take place inside a homogeneous marine boundary layer.

As will be discussed in Section 3.2, the North Sea can be considered as a semi-enclosed body of water over which air is advected from the UK in westerly wind directions, resulting in the downstream development of a Marine Internal Boundary Layer (MIBL); the air is then advected towards land over the coastline of Jutland, giving rise to a Coastal Internal Boundary Layer (CIBL) inside the marine boundary layer. Each internal boundary layer has its own meteorology; these can be quite different and are therefore described separately in the following. The objective is to describe the physical processes that are expected to be of importance.

In the case of an internal boundary layer build-up downstream from a coastline, the incoming marine boundary layer is perturbed from below due to the presence of a coastline. This perturbation or development of an internal boundary layer is reported to be sensitive to the upstream stability, as the new turbulent layer grows into an environment that either suppresses vertical motion and turbulence (stable upstream conditions) or a layer that reinforces turbulence and vertical motion (convective upstream layer).

The upstream state of the marine boundary layer, especially in terms of the mean profiles of wind and temperature Melas (1992) Gryning and Batchvarova (1990), is therefore considered of importance in the description of the coastal internal boundary layer. A separate series of numerical simulations is devoted to describing the marine layer and the parameters that are considered to be of importance. This is further described in chapter 3.

The present description will, however, start with an overview of what is considered to be important aspects of the marine boundary layer meteorology as it is observed and brought into the context of internal boundary layers.

1.3 Marine boundary layer meteorology

The marine boundary layer differs from the boundary layer over land mainly due to the significant lower aerodynamic roughness of the sea surface as compared to that of land surfaces. Moreover, the roughness is, contrary to the roughness over land, not constant with wind speeds. The roughness is a function mainly of wind speed, but wave characteristics such as wave steepness, Taylor and Yelland (2001) and wave age Johnson et al. (1998) are also known to affect the surface roughness. Most commonly used in models is the Charnock (1955) formulation in which the sea-surface roughness is a function of friction velocity. It is, however, beyond the scope of this report to investigate how the MABL structure depends on a variable roughness and all simulation are therefore done with a fixed sea-surface roughness. The sea-surface temperatures in the numerical simulations are also kept constant in time and have no spatial variability in the MABL simulation. The simulations of the MABL are related to the surface heat exchange between the ocean and the atmosphere and are considered to be an important physical driver in the MABL generation over the North Sea.

Simulation with no spatial heterogeneity in SST is chosen to reduce the complexity of the analysis, but as can be seen on the AVHRR satellite pictures in Figures 4 and 5, some degree of spatial heterogeneity can be observed in the SST. A better agreement with observations from a coastal site might possibly be achieved if this
heterogeneity were taken into account in the simulations, or it might appear that this it not a relevant forcing when describing temporal variability on a seasonal scale as is the subject of this study. For now this question will be left open.

The spatial variability of the sea-surface temperature, as observed by AVHRR (Advanced Very High Resolution Radiometer) instruments on the NOAA polar orbiter satellite, can be seen in 1 km pixel resolution on Figures 4 and 5. In the winter, the western half is seen to be warmest, while the eastern half is 4 degrees colder. The picture is reversed in the summer, as the eastern half is warmest with the western half, 3-4 degree colder for the year 1990. The monthly averaged temperature gradient across the North Sea is of the order of 0.004 Celsius/km and the temperature can therefore to first order approximation be considered constant.

Annual variability is reported by Corten and Kamp (1996), highlighting that this may be an area where further accuracy in the analysis and better correspondence with observations could be achieved by allowing for a heterogenic SST in the MABL simulation; however, this falls beyond the scope of the present report.

The diurnal cycle in the offshore part of the MABL is to a good approximation absent mainly due to an efficient mixing in the upper part of the ocean and due to the larger heat capacity of the surface water compared to that of land. Diurnal cycles have been reported in nearly-calm conditions and shallow ocean depths, but are not representative of the general picture.

Figure 4: Average SST for February, showing the average spatial SST heterogeneity in North Sea. The data are from KNMI, Netherlands and based on the average of weekly mean charts

The physical influence from surface heat exchange on the structure of the MABL as the fetch increases is investigated in Section 3.2. The MABL is found to approach a quasi-stationary equilibrium over time and distance, characterized by a near constant potential temperature profile in the lower part of the boundary layer. The temperature will gradually approach the SST, and the MABL will be capped by a temperature inversion of increasing strength over time, as discussed and observed by
Csanady (1974) and Smedman et al. (1997) over Lake Ontario and the Baltic Sea, respectively. Important physical properties about the vertical structure of the MABL can therefore be investigated keeping both sea roughness and temperature constant and letting the MABL evolve over time towards a horizontal homogeneous state with a quasi-stationary vertical structure.

The MABL close to a coastline is, however, found to be far from homogeneous and the wind speed may change over short distances, thus making the constant sea roughness questionable in littoral regions. Also the SST is reported to have large spatial variability especially in upwelling areas, as found in the Coastal Ocean Dynamic Experiment (CODE) and reported by Beardsley and Lenz (1987), likewise making the assumption of constant SST questionable in these regions.

It is therefore appropriate at this point to underline that in order to study the coastal induced perturbation in the MABL, which is the main objective of this study, a simplified description of the meteorology in the MABL is chosen with both constant roughness and SST in order to limit the complexity in the study.

Figure 5: Average SST for August showing the average spatial SST heterogeneity in the North Sea. The data are from KNMI, Netherlands and based on the average of weekly mean charts.

Smedman et al. (1997) found that the development of an upper MABL inversion lid can be diagnosed from the fetch time over water $t$, the Coriolis parameter $f$, and the air-sea-temperature difference $\Delta \theta$, formulated as a non-dimensional number:
where \( S > 75 \) creates a well-mixed lower internal boundary layer capped by an upper inversion and \( S < 75 \) results in a shallow and stable internal boundary layer. The formulation is developed for the Baltic Sea and therefore applies to a semi-enclosed body of water where large air-sea temperature differences are common as continental heated air blows offshore.

The North Sea is bounded by Norway to the north, UK to the west and central Europe to the south and therefore meets the semi-enclosed criterion, and similar conditions to those prevailing over the Baltic are expected. This is further discussed in Section 3.4.

Westerly wind directions over the North Sea are, relative to the sea-surface temperature, warmer in the winter and colder in the summer. Internal boundary layers capped by upper inversion lids must therefore be expected to occur during winter and spring with a somewhat similar structure as the observations from Baltic Sea Smedman et al. (1997) even though smaller air-sea temperature differences must be expected, influencing the structure somewhat.

Lange (2003) reported a systematic over-speeding compared to traditional profiles predicted by Monin–Obukov surface layer scaling for the stable surface layer, when an upper inversion was present in an offshore study of the wind profiles close to a coastline. Garratt (1989) found low-level wind maxima underneath the inversion capping the internal boundary layer in cold offshore flow off the coast from Australia. The evolution of a lower mixed layer capped by an elevated inversion lid takes place when warm air is advected over colder water as will be discussed in Section 3.2 and as discussed by Smedman et al. (1997). Frictional decoupling by the increase of stability above the sea surface, reduced eddy viscosity and therefore significantly reduce turbulent vertical stress divergence are known to cause strong perturbations to the downstream wind profile. This can result in inertial oscillation and generates super-geostrophic low-level jets above the stable layer Edgar (2000); Thorpe Guymer (1977), Burk and Thomson (1995), Källstrand (1998). This issue is address for the MABL above the North Sea in Section 3.4.

As the marine internal boundary layer approaches a coastline, an interaction between the marine internal boundary layer and the coastline takes place and may affect the atmosphere both upstream and downstream. The controlling parameters in this flow response are found in the structure of the marine atmosphere and topographical features ashore. A classical example of meteorology generated on a coastline by the interaction between the marine IBL and the coastline is found outside the Californian west coast. It has been investigated in numerous studies, including the Coastal Wave project reported by Brooks et al. (2002). The mechanism at play here is driven by interplay between the topography, the near-neutral marine layer and the strong low-level capping inversion. Internal waves are generated by headland-induced perturbations to the flow in the inversion. Hydraulics jumps and downstream expansion fans with associated wind speed up in the boundary layer with a factor of two have been reported. During the Danish Galathea-3 expedition, similar condition were encountered 150 km west of the coast of Namibia, where high-amplitude internal waves were observed and their influence on the boundary layer flow and momentum budget appeared to be significant, as can be seen in Figure 6 and Figure 7.
Figure 6: The vertical differences in backscatter intensity (scatter ratio per km sr$^{-1}$km$^{-1}$) from the onboard ceilometer. The large backscatter intensities measured are caused by reflection from aerosols in the marine boundary layer while the free atmosphere has low values of backscatter on the particular day with no clouds. The difference in backscatter highlights the inversion zone, whose height is seen to be altered significantly by a train of sinusoidal internal gravity waves observed from 20:00 to around 21:30 local ship time.

\[
\text{Variance}(u) = \left(1 - \frac{1}{n} \sum_{i=1}^{n} U_i^2 \right) - \left(\frac{1}{20*60} \sum_{i=1}^{n} U_i \right)^2
\]

\( n \) is number of measurements in one minute

Figure 7 shows how the one-minute average wind in the boundary layer oscillates with the period of the passing wave, estimated to be approximately 15 minutes. Assuming that the wave is stationary relative to the seafloor, an estimate for the wavelength can be deduced from

\[
\lambda = T \cdot V_{\text{ship}} = 7500m
\]

\( T \) = Period wave

\( V_{\text{ship}} \) = Ship speed relative to the seafloor

The calculated variance plotted in Figure 7 is clearly influenced by the passage of the wave. To establish the source of the observed disturbance, numerical simulations proved to be useful. The numerical simulations for that particular day developed strong showers over the littoral regions of Namibia in the late afternoon. The model simulations produced strong convective precipitation cells and the associated gust front swept from land towards sea, initiated disturbances in the capping inversion.
The gust fronts from showers over the littoral region of Namibia are therefore suspected to be the source of the observed wave train seen in the observations.

Figure 7: Wind speed observations from onboard showing 20-Hertz sonic anemometer observations and the calculated variance after applying a one-minute high-pass filter during the passage of an internal gravity wave train.

This situation is another example of how a disturbance created over land is carried many miles offshore as a consequence of the stratification of the marine atmospheric boundary layer and that the mean wind and dispersion properties are significant influence by the coastal disturbance.

The lapse rate in the marine boundary layer has also been shown to have great importance for the internal boundary layer build-up downstream from the coastline, Källstrand and Smedman (1997).

The first part of the modelling study is therefore to develop a realistic and yet physically comprehensive lapse rate in the marine boundary layer and an associated wind profile. This is done in a controlled manner using the numerical model in a highly idealized setup, which will be described in Section 3

### 1.4 Coastal internal boundary layer

A coastal internal boundary layer (CIBL) is established within the marine boundary layer downstream from the coastline as a consequence of the increase in roughness length and change in surface temperature when air blows from the sea over land as outlined in Figure 1.

The increase of surface roughness gives immediately rise to higher levels of turbulence right above the surface downstream from the coastline, causing a new set of turbulent fluxes for momentum, heat, and moisture. The new fluxes penetrate the marine boundary layer from below and perturb all the mean variables here, but each variable differently. The fluxes for TKE react most effectively to the new surface and their perturbation reaches highest up in the CIBL.

We will therefore use the height to which turbulence is increased by the presence of the coast as the height of the CIBL. The momentum reacts somewhat more slowly to
the new surface and has only in a shallow layer above the surface adjusted completely to the new surface. This layer is called the equilibrium layer. On top of the equilibrium layer is the transition layer. The flow is slowed down by the new friction force caused by the vertical divergence of the new momentum fluxes in the lower part of the transition layer, but are unaffected in the rest of the transition layer. It is of great importance for many practical applications in the littoral regions to be able to estimate the height of the internal boundary layer as an estimate of the shape of the wind profile can be deduced from this and different ways of modelling this task is described in the next section.

1.4.1 The conceptual model for the CIBL
Sempreviva (1990); Jensen and Peterson (1977) suggested a model for linear matching of the respective outer and inner wind profiles downstream from a coastline. The model is widely used and forms the basis for many wind energy applications in coastal environments, including the widely-used WAsP method, Troen and Lundtang (1989). The model assumes neutral condition over both land and water and predicts the wind profile as follows.

\( h \) is the height of the CIBL. It is to good approximation described by a power law of fetch, which is the upwind distance to the coast, and a rough estimate can be diagnosed from a scale analysis of the equation of motion as already discussed; however numerous other diagnostic approaches exist based on different levels of physical consideration. An overview will be given in the following section.

![Figure 8: Wind profile across the internal boundary layer as diagnosed from Sempreviva (1990)](image)

If the height of the internal boundary layer, \( h \), is known, the wind profile can be established accordingly Sempreviva (1990); Jensen and Peterson (1977).
\[ u(z) = \frac{u_z}{k} \ln \left( \frac{z}{z_{0,\text{land}}} \right) \text{ where } z \leq 0.1h \]

\[ u(z) = \frac{u_z}{k} \left( \frac{u(h_2) - u(h_1)}{h_2 - h_1} \right) \text{ where } 0.1h \leq z \leq \frac{h}{3} \]

\[ u(z) = \frac{u_z}{k} \ln \left( \frac{z}{z_{0,\text{water}}} \right) \text{ where } z \geq \frac{h}{3} \]

From 1.14 is noted that the upstream wind profile is diagnosed in the upper 2/3 of the internal boundary layer while the wind profile in equilibrium with the local surface only extend from the surface to 1/10 of the internal boundary layer height.

1.5 Overview of IBL models

Many approaches have been brought forward to estimate the height of the internal boundary layer with a variety of complexity and involving meteorological fields as reviewed and described by Garratt (1990) and Melas (1992).

The objective of this section is to give an overview of the existing methods for diagnosing the depth of the internal boundary layer and highlight the areas where the by far more computationally expensive numerical approach should contribute with more complex results over existing methods. This is done in terms of describing the four fundamental different ways of producing IBL models capable of this task and will lead to a description of the numerical model approached used in this study. The IBL models can be divided into 4 groups listed here with an increasing amount of complexity:

- Empirical IBL models
- Similarity models
- Slab IBL model
- Numerical mesoscale models

1.5.1 Empirical models

As accounted for in section 1.1.1, it is physically reasonable to expect, that the height of the internal boundary layer follows a power-law dependency on the downwind fetch distance \( x \)

\[ Z_{ibl} = ax^b \]

Several authors Pasquill (1972); Smedman and Högström (1978) derive expressions for the stability- and surface-roughness-dependant parameters \( a \) and \( b \) from observations. Also versions with constant \( a \) and \( b \) have been put forward Hsu (1986) who suggested \( a = 1.91 \) and \( b = 0.5 \).

The power-law dependency for the stationary, one-dimensional case can alternatively be derived following Gryning (2005) and, assuming that the growth of the internal boundary layer along the \( x \) axis is due solely to mechanical turbulence production, formulated as
\[ \frac{dh}{dx} \approx u \frac{\partial h}{\partial x} = \frac{B}{Cu} \]

\[ \downarrow \]

\[ \frac{\partial h}{\partial x} = \frac{Bu}{Cu} \]

\[ C = 1.667 \text{ Panofsky and Dutton (1984)} \]

Assuming that the wind at height \( z \) can be found according to the logarithmic expression

\[ u(z) = \frac{u_*}{k} \ln \left( \frac{z}{z_0} \right) \]

and applying to 1.15 leads to

\[ \frac{\partial h}{\partial x} \ln \left( \frac{z}{z_0} \right) = \frac{Bk}{C} \frac{\partial x}{\partial x} \]

\[ \downarrow \]

\[ \int_{z_0}^{h} \frac{\partial h}{\partial x} \ln \left( \frac{z}{z_0} \right) = \int_{0}^{x} \frac{Bk}{C} \frac{\partial x}{\partial x} \iff \frac{x Bk}{z_0 C} = \left[ \frac{h}{z_0} \ln \left( \frac{h}{z_0} \right) - 1 \right] + 1 \]

\[ h \approx x^{0.8} \]

The integration is performed where the assumption, that the height of the internal boundary layer is equal to the roughness length for \( x=0 \) is used to establish the constant of integration.

Figure 9: Dashed blue control line illustrates that 1.16 predicts the internal boundary layer height \( h \) as a power law of fetch distance proportional to \( h \approx x^{0.8} \).
1.5.2 Similarity models

The similarity models, where both physical and dimensional arguments are used to derive expressions for the internal boundary layer growth, are conceptually placed in between the empirical and the more physical based slab models. Raynor (1975) suggested

\[ Z_{ibl} = \frac{u^* x}{U} \left( \frac{\Delta \theta}{\lambda} \right)^{1/2} \]

\[ \lambda = \text{lapse rate upwind} \]
\[ \Delta \theta = \text{land sea temperature difference} \]
\[ u^* = \text{friction velocity} \]
\[ U = \text{mean wind} \]

and Miyake (1965); Hunt and Simpson (1982); Melas (1990) suggested the more advanced model applicable for convective condition over mesoscale fetches.

\[ Z_{ibl}^{2/3} = B_m D \left( \frac{g}{T_0} C_i C_D \Delta \theta \right)^{1/3} u_h^{-2/3} x \]

\[ C_i = \text{empirical site dependant constant} \]
\[ C_d = \frac{u_*}{u_h} \]
\[ Q_0 = \text{kinematic heatflux} \]
\[ \Delta \theta = \text{land-sea temperature difference} \]
\[ B_m = 0.67 \]
\[ D = \text{constant} \approx 1 \]
\[ u_h = \text{wind speed above layer} \]
\[ U = \text{mean wind} \]

1.5.3 Physical slab models

The formulation of the physical slab models for boundary layer growth relies on the physical principle of heat and volume conservation for horizontal homogeneous conditions. Heat conservation is stated as

\[ \left( \frac{d \theta}{dt} \right)_{\text{at}} = \left( \frac{\theta^' w^'}{h} \right)_{s} - \left( \frac{\theta^' w^'}{h} \right)_{h} \]

\[ \left( \theta^' w^' \right)_{s} = \text{Kinematical vertical heat flux at the surface} \]
\[ \left( \theta^' w^' \right)_{h} = \text{Kinematical vertical heat flux at the inversion height} \]

The equation expresses that the potential temperature in the internal boundary layer is controlled by vertical heat flux divergence at the bottom and at the top. It is assumed that the air in the boundary layer is well mixed and therefore to the first order has constant density. The equation of continuity can therefore be written as
\[ \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = 0 \]

It is furthermore assumed Deardorff (1979); Gryning and Batchvarova (1990) that the boundary layer and the air above are separated by an infinitesimal thin layer with across which there is a temperature jump (zero order model) and that the kinematic heat flux inside the internal boundary layer has a linear vertical profile as outlined in Figure 10.

Figure 10: Assumption on temperature and heat flux profile from Gryning and Batchvarova (1990)

The slab model relies on the following physical considerations:

The volume conservation leads to a prognostic equation for the internal boundary layer height as the sum of 2 growth terms for the IBL, one related to internal processes and one related to external process Stull (1988).

Equation 1.18 expresses that the volume over the area is changed due to a combination of large-scale laminar subsiding /rising vertical motion of the air, denoted \( W \) and turbulent entrainment into the air above.

\[ A \frac{dh}{dt} = A W_{\text{entrainment}} + A W_{\text{large scale subsidence}} \]
Figure 11: Vertical change in height of the air column due to laminar and turbulent vertical entrainment

In order to derive a prognostic equation for $h$ in terms of available meteorological parameters, (1.17) and (1.18) are combined. The $W$ from large scale motion can be diagnosed from the divergence of the horizontal velocity field and can be obtained as outputs from NWP models or from a network of meteorological stations. The divergence of the wind field is influenced by the height and complexity of the topography and can therefore locally be quite large. $W$ from entrainment can be found following the physical argumentation of Gryning (2005); Gryning and Batchvarova (1990) by looking into the mechanism behind the entrainment process of $h$ and what consequences this process has on the heat budget. Their analysis with the assumption accounted for above, involves 4 physical processes leading to a formulation of a prognostic equation for $h$. The 4 processes are the following:

Process 1
A time-averaged heat flux into the internal boundary layer at the top is generated by the growth of the IBL though the entrainment zone with strength $\Delta$

$$-(\bar{\theta}'w')_h = \Delta \frac{dh}{dt}$$  \hspace{1cm} 1.19

Process 2
Temperature jump strength is changed in time because
- the internal boundary layer grows;
- the internal boundary layer is heated.

$$\frac{d\Delta}{dt} = \gamma \frac{dh}{dt} - \left( \frac{d\theta}{dt} \right)_{nl}$$  \hspace{1cm} 1.20

The solution can be approximated to
\[ \Delta = \frac{Ah - BkL}{(1 + 2A) - 2BkL} \gamma h \]

\[ L = -\frac{u_s^3}{\frac{Kg}{T} \left( \theta'w' \right)} \]

A, B = integration constants found in process 4

Process 3
Heat conservation

\[ \left( \frac{d\theta}{dt} \right)_{md} = \left( \frac{\theta'w'}{h} \right)_b - \left( \frac{\theta'w'}{h} \right)_h \]

The conservation equation expresses the balance between heated air entrained through the top and heat added at the surface.

Process 4
The energy that is used to entrain air at the top of the IBL is generated by the production of TKE inside the boundary layer and can therefore be evaluated from the prognostic equation for mean turbulent kinetic energy under horizontal homogeneous conditions Stull (1988). The right-hand side reads

\[ \int_0^h \frac{\partial}{\partial t} e'dz = \frac{\partial}{\partial t} e'dz + \varepsilon_T - \varepsilon \]

\[ \frac{\partial}{\partial z} \left( \frac{\theta'w'}{\rho} \right) \frac{\partial u}{\partial z} - \frac{\partial}{\partial z} \left( \frac{p'w'}{\rho} \right) \frac{\partial}{\partial z} \left( \frac{\theta'w'}{\rho} \right) - \varepsilon \]

where the temporal change in TKE is considered small following the argument of Mahrt and Lenschow (1976). Gryning and Batchvarova integrated the right-hand side of 1.22 from the surface to h, which provides an expression for the heat flux at the top, where only terms with a significant magnitude are considered. For a detailed description, see Gryning (2005) Gryning and Batchvarova (1990). The buoyancy term is split into a lower production term and an upper consumption term according to the assumption for the vertical profile of the heat flux, as outlined in Figure 1 Stage and Businger (1981).

Equation 1.22 reads
Parameterization terms and assuming dissipation rate is proportional to production rate

\[ C_{e_{h}} \frac{dh}{dt} \approx g \left( \theta'w' \right)_{h} + \frac{A}{T} \left( \theta'w' \right)_{h} + B u_{*}^{3} \]

A parameterization constant relating to convective turbulence
B parameterization constant relating to mechanical turbulence
C parameterization constant relating to exchange of turbulence over the inversion

Solving for \( \left( \theta'w' \right)_{h} \)

\[ -\left( \theta'w' \right)_{h} = A \left( \theta'w' \right)_{s} + \frac{Bu_{*}^{3}}{g \left( \frac{g}{T} \right)_{h}} dt \]

Inserting 1.23 and 1.21 into 1.20 leads to the prognostic expression for height of the convective boundary layer.

\[ \frac{h^{2}}{(1 + 2A)H - 2BkL} + \frac{Cu_{*}^{2}}{\gamma \left( \frac{g}{T} \right)_{h} (1 + A)h - BkL} \left[ \frac{dh}{dt} - w_{s} \right] = \left( \theta'w' \right)_{s} \]

1.24 can also be used to predict the height of the internal boundary layer when rewriting the material derivative in 1.10 using the definition

\[ \frac{dh}{dt} = \frac{\partial h}{\partial t} + u \frac{\partial h}{\partial x} + v \frac{\partial h}{\partial y} \]

\[ \frac{h^{2}}{(1 + 2A)H - 2BkL} + \frac{Cu_{*}^{2}}{\gamma \left( \frac{g}{T} \right)_{h} (1 + A)h - BkL} \left[ \frac{\partial h}{\partial t} + u \frac{\partial h}{\partial x} + v \frac{\partial h}{\partial y} - w_{s} \right] = \left( \theta'w' \right)_{s} \]

Equation 1.25 is the two-dimensional prognostic equation for the internal boundary layer height \( h \) and, \( u \) and \( v \) are the easterly and northerly wind components respectively.

The wind component can be obtained as output from a numerical model or from a dense measuring grid.

In the convective limit \( L \rightarrow 0 \), with no spin up of turbulence, \( C=0 \), no local temporal change in \( h \)

\[ \frac{\partial h}{\partial t} = 0 \]

and no subsidence \( w_{s} = 0 \) (1.11) a power-law dependence on fetch distance downwind is found again
\[ \int_0^h (h \partial h) = \int_0^1 (1 + 2.4) (\theta' w') \gamma \Rightarrow h = \sqrt{\frac{2(1 + 2.4)(\theta' w')}{u \gamma}} \]

Here \( u \) is assumed to be constant across the roughness change

The slab model described here has attained a large degree of success in validation studies (see Källstrand and Smedman (1997)) even though it has been simplified in a number of ways compared to a full numerical model. The areas of simplification are highlighted below and how these issues are dealt with in the numerical model will be addressed in the next section.

- No coupling between the physics in the model and the background wind obtained from a mesoscale model
- Constant lapse rate over the water
- Linear flux profiles
- Infinitesimally thin inversion
- Incompressible equation of continuity

### 1.5.4 NWP model

**Introduction and history**

The numerical approach to the study and prediction of weather phenomena has existed since Lewis Fry Richardson’s first attempt to calculate the weather for 20 May 1910 by using a numerical scheme for solving the governing equation for the atmospheric flow. The approach was, however, not practically feasible until the arrival of computers after the Second World War. The first numerical experiments was conducted in the early 1950s using Jule Charney’s simplified barotropic model for forecasting the 500 hPa flow over the northern U.S. with some success. An increase in the understanding of atmospheric processes and a huge increase in computational power have led to a generation of advanced computer models solving the full set of primitive equations, capable of resolving weather phenomena on an ever decreasing horizontal scale.

The introduction of non-hydrostatic models in the early 1960s Orura and Charney (1962) for the study of convective systems with considerable vertical accelerations, has made it possible to further decrease the horizontal resolution towards the atmospheric meso \( \beta \) scale and even lower (Orlanski scale classification Figure 12). The non-hydrostatic models are therefore, in terms of the ability to resolve multi-scale atmospheric processes with considerable vertical accelerations, a comprehensive physical alternative to the simpler models described in the previous sections.

Internal gravity waves are among the processes generated on a coastline by the abrupt change in surface properties. Numerous investigations have been performed in areas where upwelling cools the lower atmosphere and creates shallow boundary layer with strong capping inversions. Wavelengths of the order of 1 km to 10 km (Danish Galathea-3 expedition) Söderberg and Tjernström (2001) Ström et al. (2001), Brown et al. (2004) have been reported, dictating a horizontal resolution in numerical simulations of 100-2000 m in order to capture these flow phenomena Pielke and Kennedy (1980) , Young and Pielke (1983). The issues of horizontal resolution introduces problems relating to initializations of the simulations as an observational grid at the same resolution is needed to properly describe the initial meteorological situation but has been performed Doyle (1997) using a combination of aircraft, satellite, measurements and upper air soundings to create the initial field for the simulations and addresses the multi-process nature of coastal meteorology.
An alternative and widely-used approach in coastal meteorology is based on a numerical investigation using an idealized setup, addressing individual forcing processes Rotunno et al. (1992); Rogers (1995). But even idealized numerical modelling on the atmospheric meso γ scale introduces problems in terms of the PBL parameterizations. The focus in the following section will therefore be on the PBL parameterization in the used NWP model and how to interpret high-resolution results in the boundary layer.

**Model and its use**

The numerical model used in this study is the three-dimensional coupled Ocean/atmosphere meso scale prediction system (COAMPS) version 3.1.1 described by Hodur (1997) and continuously being developed at the Naval Research Laboratory. As part of this PhD study, the model was implemented on a Linux cluster at Risø National Laboratory. During the study, the model setup was used in a real data simulation configuration, consisting of daily data assimilation, initialization, and integration to produce weather forecasts for the atmosphere research groups along the route of Galathea-3.

The Naval global model NOGAPS was used to provide lateral boundaries for the triply nested setup centred over the noon position of the expedition. The model was also used in an idealized standalone setup, as a numerical laboratory where surface properties were controlled and specified through simple analytical functions. The model was started on a reference sounding and brought into equilibrium assuming horizontal homogeneous conditions in the vertical structure. In this way, a large degree of the forcing remained under easy control, simplifying the interpretation of the results. Validation against observations ensure that the simulated results behave realistically, but an exact correspondence cannot be expected in the highly idealized setup where the objective was to study the physical properties behind the observations more than aiming for the best fit to observations.
The scale of the meteorology of interest in this study is the build-up of an internal boundary layer taking place between the coastline and a mast 1500 m inland. The internal boundary layer is calculated at horizontal intervals of 50 m and according to the Orlanski scale classifications, relates to scales ranging from mesoscale $\gamma$ to microscale $\alpha$ scales and therefore pushes the limit of mesoscale modelling into the domain of microscale modelling. More details on the limitation and interpretation in this aspects will be addressed in the following sections.

**NWP concept**

The basic concept behind a numerical meteorological model is based on a set of conservation criteria that form a set of coupled nonlinear partial differential equations that must be fulfilled simultaneously. The coupled set of equations is known as the primitive equations, covering the conservation of mass, conservation of heat, conservation of momentum, conservation of water and any other scalar such as passive tracers. Analytical solutions to these equations exist for a limited number of highly idealized situations covered by the primitive set of equations, but no methods exist to solve the general set of equations and numerical methods have to be used. The basic principles and their implementation in the numerical model used in this study will here be explained.

The first step in a numerical approach is discretize the area of interest by setting up a number of grid points in x, y, and z direction. The distances between grid points denote the horizontal and vertical resolution, and are chosen in such a way that 6-10 grid points are used to resolve the smallest meteorological scale of interest.
Horizontal Resolution = \( \frac{\lambda_{\text{smallest}}}{10} \)

In COAMPS, the horizontal grid uses the Arakawa-Lamb (1977) scheme c staggering, which is commonly used in mesoscale models. The staggering of the dependent variables means that the \( u \) component of the wind is half a grid distance to the east and the \( v \) component is half a grid distance to the north while \( w \) are computed at the mass point. All scalars are computed at mass points, which means that the wind component is averaged to the grid point. The Coriolis term in the grid point \( i, j \) is found as follows

\[
f v_{i,j} = \frac{(f_{i+1,j} + f_{i,j})}{2} \left( v_{i,j} + v_{i+1,j} + v_{i,j-1} + v_{i,j+1} \right) \div 4
\]

The staggering of the dependent variables is known to increase the effective resolution, since derivatives are defined over a single grid distance instead of 2 with no staggering Pielke (2002)

The vertical coordinate is the terrain-following sigma Z system and the vertical velocity is calculated on integer sigma level while all other variables are calculated on half levels as seen in Figure 14, where \( H \) is depth of the Atmosphere, \( z \) is height, and \( z_s \) is height of the terrain in the grid point.

\[
\sigma = \frac{H(z - z_s)}{z - z_s}
\]

\[
z = z_i
\]

Figure 14: Coamps \( \sigma \) sigma coordinate system
Model equations

The Atmospheric portion of the Naval Research Laboratory Coupled Ocean Atmospheric Mesoscale System COAMPS is a finite-difference approximation to the non-hydrostatic fully compressible equations of motion following Klemp and Wilhemson (1977) with a suite of physical parameterizations of surface fluxes, boundary layer physics and moist processes described in Hodur (1997) and Hodur and Doyle (1998). The physical processes behind the parameterizations can be switched on and off in order to meet the complexity of the area of interest in the numerical experiment.

The model solves the governing equations on an Arakawa- Lamb (1977) scheme c staggering horizontal grid and the vertical grid is a terrain following sigma coordinate system following Glen and Somerville (1975).

Closure schemes and other issues related to the model planetary boundary layer parameterization will be addressed in detail in next section

Equation of state

\[ p = \rho_{\text{ai}} R_d T_v \]

\[ R_d = \text{gas constant for dry air} \]

\[ T_v = \text{virtual temperature} \]

Conservation of heat

\[ \frac{d\theta}{dt} = \frac{Q_0}{\rho} + D_\theta + K_h \nabla^4 (\theta - \bar{\theta}) \]

\[ \frac{Q_0}{\rho} = \text{source term} \]

\[ D_\theta = \text{subgrid heat flux divergence horizontal diffusion} \]

Conservation of mass

\[ \frac{\partial \rho}{\partial t} + \frac{\rho u_j}{\partial x_j} = 0 \]

Pressure

COAMPS uses a scale form for pressure (Exner function) defined as

\[ \pi = C_p \left( \frac{P}{P_0} \right)^{\frac{R_g}{C_p}} \]

\[ R_g = \text{gas constant for dry air} \]

\[ C_p = \text{specific heat of moist air} \]

\[ P = \text{pressure} \]

\[ P_0 = \text{reference pressure at ground level} \]
which is decomposed into a mean pressure and a dynamic pressure part

\[ \pi = \overline{\pi} + \pi' \]

where the mean part is assumed to be in hydrostatic balance and can be found as

\[ \frac{\partial \overline{\pi}}{\partial z} = -\frac{g}{C_p \theta} \]

A prognostic equation for the dynamic part is derived, taking the material derivative of the definition of scaled pressure

\[ \frac{d(\pi)}{dt} = \frac{\partial (\pi)}{\partial t} + u_i \frac{\partial (\pi)}{\partial x_i} \]

and using the equation of state 1.26 to express pressure instead of density in the compressible continuity equation 1.28 and the heat equation leading to a prognostic equation for pressure.

\[ \frac{\partial \pi'}{\partial t} + \frac{c_p^2}{\rho_0} D_3 = u \left( \frac{\partial \pi}{\partial x} \right)_\sigma + v \left( \frac{\partial \pi}{\partial y} \right)_\sigma + \sigma \left( \frac{\partial \pi}{\partial \sigma} \right)_\sigma - \frac{R_d \pi}{C_v} \left( \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial \sigma}{\partial \sigma} \right) + \frac{c_p^2}{\rho_0} \frac{d \theta}{c} \]

\[ \text{Conservation of momentum} \]

Seen in a rotated earth system Newton second law of motion in the x direction \( m a_x = \sum f_x \) can be written

\[ \left[ \frac{\partial u}{\partial t} + c_p \theta \left( \frac{\partial \pi'}{\partial x} + G_x \frac{\partial \pi'}{\partial \sigma} \right) - K_m \frac{\partial \Delta^4 u}{\partial x} \right] = \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial \sigma}{\partial \sigma} + f v + D_u + K_a \Delta^4 u \]

\[ G_x = \frac{\partial \sigma}{\partial x} \left( \frac{\sigma - z_{topo}}{z_{topo} - z_{surface}} \right) \frac{\partial z_{skin}}{\partial x} \]

\[ D_u = \frac{\partial}{\partial x} + G_x \frac{\partial}{\partial \sigma} \rho \theta u + \left( \frac{\partial}{\partial y} + G \frac{\partial}{\partial \sigma} \right) \rho \theta v + G_z \frac{\partial \theta w}{z} \]

\[ D_u = \frac{\partial}{\partial \sigma} K_m \frac{\partial U}{\partial \sigma} = \text{subgrid Vertical momentum flux divergence (Mellor and Yamada 1982)} \]

and \( K_m \) is the eddy diffusivity for momentum. Integrated using an implicit scheme

\( K_a \Delta^4 u = \text{fourth order accurate horizontal diffusion used to control nonlinear instability as wave-wave interaction to a still smaller wavelength that conflict with the grid resolution} \)

\[ \text{Numerical solutions to governing equations} \]

As discussed in the previous section no analytical solution exists to the general set of primitive equation and therefore the equations are solved at each grid point by approximating the derivatives by a finite-difference method. This means setting up
an algebraic equation for each grid point, resulting in a set of algebraic equations with as many equations as there are grid points, each grid point being an unknown. Using numerical schemes that involve only the closest neighbour grid points dramatically reduce the computational time that is needed to solve the resulting sparse algebraic set of equations.

The scheme used in COAMPS is a smoothed centred-in-time or leapfrog scheme where the dependant variable $\phi$ is integrated forward in time as

$$\phi^{*+\Delta t} = \phi^{*-\Delta t} + 2\Delta t \phi^*$$

$$\phi^* = \phi^* + \alpha \left( \phi^{*+\Delta t} - 2\phi^* + \phi^{*-\Delta t} \right)$$

where the first equation corresponds to a time step forward in time for the variable $\phi$ which is forced by

$$\frac{\partial \phi}{\partial t} = \text{Forcing}$$

The asterisk denotes that the variable has not yet been smoothed through application of the second equation. The net effect is to damp any computational model in the integration while physical modes are unaffected Asselin (1972). The governing equations of motion as described are compressible and thus permit for both propagation of sound and gravity waves. The possibility of fast-moving sound waves within the solutions severely restricts the model time step in order to maintain numerical stability in the integrations. The issue is solved in COAMPS Following Klemp and Wilhelmson (1977) by having a larger time step for slow modes and a separate small time step for fast moving sound waves, a procedure known as time splitting.

**Model PBL**

This study is focuses on the development of the internal boundary layer in the coastal region and the associated profiles of turbulence temperature and wind speed. Special attention is put on the way that the numerical model allows for the surface to perturb the model atmosphere within a vertical distance defined as the boundary layer height and how to interpret the results.

The surface perturbs the atmosphere above it, mainly by the process of turbulence seeking to level out gradients in momentum and scalars. The surface always has zero momentum relative to the atmosphere and a gradient in momentum exist above the surface, where momentum fluxes tries to level out the gradient by mixing high momentum values to the surface and low momentum values upwards by a variety of stochastic movements. The net result of the process is that the momentum fluxes in the boundary layer act like a drag on the free flow in the atmosphere and are therefore known as the friction forces. The turbulence, being a separate atmospheric state which is poorly understood on a basic level, that drives the flux phenomenon, is a multi-scale phenomenon with length scales from the height of the boundary layer (1-2 km) down to Kolmogorov micro scale (1 mm) and is therefore in most simulations below what can be resolved by the computational grid.

The effects on the free atmosphere from processes in the boundary layer are therefore in all mesoscale models accounted for by different degrees of sub-grid parameterizations. The parameterization schemes for a grid volume for turbulent vertical fluxes are based on experimental data where an ensemble of measurements of mean and turbulent variables are correlated, to determine the constant used in the closure scheme.
The sub-grid values of turbulent fluxes as predicted by the parameterization scheme, is therefore an ensemble mean rather than a grid volume average and is to be interpreted as being the most likely value for the given set of mean prognostic variables. When comparing the computed effect from the turbulent flux, like internal boundary layer wind profiles and temperature profiles, with observations from a mast it is therefore only physically meaningful when a larger number of observations are present that can be averaged to generate an ensemble mean as is computed with a PBL scheme.

In the setup of the numerical model used in this, several vertical computational layers are placed inside the PBL. The model therefore resolves both the surface layer and the transition layer, which is defined as being the layer between the surface layer and the free atmosphere and is accounted for by the model closure scheme. The lower boundary values of turbulent fluxes are the fluxes generated between the surface and the lowest computational level are calculated following Louis (1979).

In the model domain, the turbulent moment and heat fluxes are diagnosed following a 1.5 order, level 2.5 closure scheme Mellor and Yamada (1982) as follows:

\[
\overline{u'w'} = -K_m \frac{\partial U}{\partial Z}
\]

\[
D_u = \frac{\partial}{\partial \sigma} K_m \frac{\partial U}{\partial \sigma}
\]

\[
\overline{\theta'w'} = -K_\theta \frac{\partial U}{\partial Z}
\]

\[l = kz / (1 + kz / \lambda)\]

Mixing length formulation from Blackadar (1962)

\[
\frac{De}{Dt} = \frac{\partial}{\partial z} \left[ K_m \left( \frac{\partial e}{\partial z} \right) \right] = K_m \left( \frac{\partial U}{\partial z} \right)^2 + K_m \left( \frac{\partial V}{\partial z} \right)^2 - \beta g K_m \frac{\partial \theta}{\partial z} - \frac{2e^{1/2}}{\lambda_1} + U \frac{\partial e}{\partial x} + V \frac{\partial e}{\partial y}
\]

\[\lambda = \alpha \int \frac{z e dz}{edz} \quad \alpha = 0.1 \text{ (stable stratification) } \alpha = 0.1 \cdot \frac{0.2}{z} \frac{z}{L} \text{ (unstable stratification)}
\]

\[S_m, S_h = \text{Polynomials of the Flux Richardson number}\]

\[Ri_f = \frac{g w'\theta'}{\theta u'w' \frac{\partial U}{\partial z} + v'w' \frac{\partial V}{\partial z}}\]

The mixing-length formulation used in this study Blackadar (1962) is to some extent similar to a recent mixing-length formulation derived from observations from masts extending above the surface layer Gryning et al. (2007). Inside the surface layer (constant flux layer), the linear increase with height part forms the shape of the wind profile into the well known logarithmic wind profile. Above the constant flux layer, the mixing length approaches a constant, which reacts like an over-speeding of the wind profiles into a linear relation with height. Corrections for stability are included.
in the polynomial expression of the local flux Richardson number in the expressions for $S_m, S_h$ and ensure that the wind profiles matches the stability corrected Monin-Obukhov surface layer scaling profiles inside the surface layer (Panofsky 1984).

**PBL modelling of the internal boundary layer**
The area of internal boundary layer development in the littoral region is characterized by both a change in surface roughness and surface temperature altering the profiles of meteorological variables inside the internal boundary layer, while the layer above resembles upstream conditions. The interface between the upstream generated layer and the local layer is characterised by a change in stability and turbulence affecting the way the internal boundary layer grows and how the wind profiles develops.

A mesoscale model with a high vertical resolution in the lowest part of the boundary layer do to a higher degree than the slab models, resolve the complex mutual coupling of stability and turbulence changes to the change in wind and temperature profiles in the littoral area.

The prognostic TKE equation in the 1.5 order level 2.5 closure accounts for the temporal and spatial variations in the flow conditions across the coastline and through the eddy viscosity formulations 1.32, adjust the momentum transport from the surface and upwards. A conflict does however exist on the lower boundary of the domain as the surface flux input to the domain is calculated with the Monin-Obukhov surface layer scaling based Louis (1979) scheme, that relies on assumptions on horizontal homogeneous surface properties and stationary conditions which is violated in a short the distance from a coastline. This is a known problem for all PBL mesoscale modelling with a surface scheme relying on Monin-Obukhov similarity and one has to assume that the effect of this inconsistency is not critical to the overall results following argumentation from Brooks et al. (2002).

**2 Measurements and site**

**2.1 Høvsøre**
The observational basis for this thesis consists of measurements from the Høvsøre National Test Station for Large Wind Turbines situated in the northwestern part of Denmark, close to the North Sea, as seen on Figure 14. The test site was established in order to have a facility where large wind turbines can be tested before they are put into production. As part of the turbine testing procedure, an extensive meteorological data record is being continually gathered from an array of observing towers situated as outlined in Figure 17.

The landscape around the site is characterized by a steep dyke along the coastline with an approximate elevation of 10-20 m and a width of 50-100 m. The landscape between the dyke and the meteorology towers at the test sites is flat grassland, with no major obstacles.
2.1.1 Data extraction

The meteorology observations at Høvsøre test station started on 27 February 2004 and the data record extends forward to the present day. The 30-minute averaged meteorological data used in this report are extracted from this time series and span a period from 27 February 2004 until 20 May 2008. Only observations falling within a narrow wind direction sector (260°-280° on the wind vane at 100 m height) are retained, in order to have a near-constant fetch distance from the coastline to the masts as well as to avoid contamination of measurements by the wakes of the turbines at the site.

Moreover, only observations with wind speeds between 5 and 25 m/s, as measured at 40 m height, are retained. The low wind speed criterion of 5 m/s is imposed in order to filter out odd flow situations with decoupled flow and low level stagnant layers. Wind speed measurements are taken by cup anemometers and momentum and heat flux measurements by sonic anemometers. The sonic anemometer measurements are quality-controlled in the following way.

If one of the towers sonic anemometer was malfunctioning during a given 30-minute period, then all the towers sonic anemometers for that period were excluded in the analysis.

An alternative method was considered where linear interpolation is carried out over the height with a malfunctioning sonic anemometer. However, this option was rejected, because the profiles of turbulent kinetic energy (TKE) were used to estimate the height of the internal boundary layer, and the use of an interpolation around heights with bad readings might have blurred this TKE signal. The chosen procedure implies that there exist periods where profiles of TKE are absent while profiles of the other meteorological variables are available, but ensures that all the measurement are physical.

Malfunctions of the sonic anemometers often occur in continuous, prolonged time periods, there are entire seasons during which there is a significant reduction in the number of turbulent kinetic profiles available and therefore the resulting averages for these seasons have a lower statistical significance. Care must therefore be taken in the interpretation of the data.
Figure 17: Test site outline including turbine and meteorology mast positions

The turbine stands are placed in the north-south direction at intervals of 300 m, with turbine stand 5 the southernmost one. A meteorological mast is located at a distance of 240 m due west of each turbine stand, with a height equal to the corresponding turbine’s hub height. In addition, a 116 m meteorological tower is placed south of the turbine array, and two light towers of 160 m are placed between stands 1 and 2 and stands 4 and 5, respectively. The measurements from the light towers at 160 m and the meteorology mast at 116 m were used in the observational analysis for this study. The meteorological variables available from the mast at heights are outlined in table 1.

Table 1 Measurements lay out

<table>
<thead>
<tr>
<th>Height</th>
<th>Wind speed</th>
<th>Wind dir.</th>
<th>Temperature</th>
<th>Fluxes</th>
</tr>
</thead>
<tbody>
<tr>
<td>160 m</td>
<td>+</td>
<td></td>
<td>+</td>
<td>+</td>
</tr>
<tr>
<td>116 m</td>
<td>+</td>
<td></td>
<td>-</td>
<td></td>
</tr>
<tr>
<td>100 m</td>
<td>+</td>
<td>+</td>
<td>+</td>
<td>+</td>
</tr>
<tr>
<td>80 m</td>
<td>+</td>
<td></td>
<td>+</td>
<td>+</td>
</tr>
<tr>
<td>60 m</td>
<td>+</td>
<td>+</td>
<td>+</td>
<td>+</td>
</tr>
<tr>
<td>40 m</td>
<td>+</td>
<td></td>
<td>+</td>
<td>+</td>
</tr>
<tr>
<td>20 m</td>
<td>-</td>
<td></td>
<td>-</td>
<td>+</td>
</tr>
<tr>
<td>10 m</td>
<td>+</td>
<td>+</td>
<td>+</td>
<td>+</td>
</tr>
<tr>
<td>2 m</td>
<td>-</td>
<td></td>
<td>+</td>
<td></td>
</tr>
<tr>
<td>Ground</td>
<td>-</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

2.2 Horns Rev
Upstream observations of the stability distribution in the marine boundary layer were found to be of importance in describing the wind climate observations at upper levels at Høvsøre. Observation of the sea-surface temperature and of the air temperature at 13 m height were taken from the Horns Rev offshore wind farm. A bulk Richardson number was calculated using 2.4 in order to get the stability distribution upstream.
Horn Rev, situated 14-18 km west of Blåvands Huk, is one of the world’s largest offshore wind farms. The objective of the development of Horns Rev wind farm was to explore the potential for offshore wind energy in Denmark, in order to meet the objective of the Danish energy plan of an installed offshore wind farm capacity of 4000 MW in Danish waters before 2030.

The observations from Horns Rev available for this study span the period from December 2003 to November 2004. For sea-surface temperature measurements, the sensor at mast M2 situated 4 m below the surface was used, while the air temperature was measured by the temperature sensor in the tower at 13 m above sea level.

### 2.3 Stability

The stability distributions in the lower boundary layer, shown in Figure 20 and Figure 21, indicate the relative frequency and the degree of stable and unstable atmospheric conditions for a given month.

To classify the stability of the weather situations, the Monin-Obukhov length, $L$, defined in 2.1, is evaluated from the sonic anemometer at 20 m, and the atmospheric stability conditions are subsequently sorted into 7 stability bins following Gryning et al. (2007). The stability bins are defined as:

<table>
<thead>
<tr>
<th>Stab class</th>
<th>Stab 1 Very unstable</th>
<th>Stab 2 Slightly unstable</th>
<th>Stab 3 Slightly stable</th>
<th>Stab 4 Neutral</th>
<th>Stab 5 Slightly unstable</th>
<th>Stab 6 Stable</th>
<th>Stab 7 Very stable</th>
</tr>
</thead>
<tbody>
<tr>
<td>Value range of $L$</td>
<td>-100 $&lt;L&lt; -50$</td>
<td>-200 $&lt;L&lt; -100$</td>
<td>-500 $&lt;L&lt; -200$</td>
<td>-500 $&lt;L&lt; 0$</td>
<td>200 $&lt;L&lt; 100$</td>
<td>500 $&lt;L&lt; 500$</td>
<td>200 $&lt;L&lt; 500$</td>
</tr>
</tbody>
</table>

From the definition of $L$, it is seen that convective classes are characterized by having positive heat flux, meaning that heat is moved from the surface and upwards and relating to situations where the atmosphere is colder than the surface; and vice-versa for the stable bins.

The distribution of stability changes over the year due to the yearly variation of surface heat flux and therefore relates to a change in the number of weather
situations, where the atmosphere is significantly warmer, warmer, slightly warmer, same temperature, or slightly colder and so forth, relative to the surface. Over land, the surface heat flux is seen (Figure 20) to be seasonal, having a maximum in unstable atmospheric conditions close to midsummer, while the maximum of stable atmospheric conditions is found in early spring. Above water (Figure 21), a time lag between the midsummer maximum in incoming solar radiation to the sea surface and the maximum in unstable weather situations can be seen. The maximum in unstable atmospheric conditions is seen to occur in August and correlates with the months where the SST in the eastern part of the North Sea is at its maximum. The maximum of stable atmospheric conditions over the water is found in early spring (March and April) when the SST in the eastern North Sea is at its minimum.

From this analysis, it is seen that stability distribution over the water and over the land exhibit similar features in terms of months with a maximum in stable atmospheric conditions, in that this occurs for both Høvsøre and Horns Rev in the early spring, while the maximum in unstable atmospheric conditions is phase-shifted by 2 months between Høvsøre and Horns Rev. The percentage of the most stable atmospheric conditions is seen to be larger at Horns Rev than at Høvsøre for all months.

The larger percentage of stable situations, together with the delay in the convective stability distributions over water compared to the convective stability distribution over land, are proposed to be the main meteorological phenomena behind the seasonal behaviour seen in the normalized wind profiles from Høvsøre. The observations are further presented and discussed in the next section. In Chapter 3, we investigate, using numerical mesoscale model simulations, to what degree the time lag in stability distribution discussed above can explain the seasonal behaviour in the wind speed. For the Høvsøre test site, L is calculated from the sonic measurements at 20 m according to:

\[
L = \frac{-u_{20m}^3}{k \left( g / T_{20m} \right) \left( w' T' \right)_{20m}}
\]

\( \kappa = \) von Karman constant
\( g = \) acceleration of gravity
\( T = \) temperature

and

\[
u_*^2 = \left[ u'^2 w'_{20m}^2 + v'^2 w'_{20m}^2 \right]^{0.5}
\]

The sonic anemometers at Horns Rev were found to be of low quality and were therefore taken out of this study. Instead, L was calculated from a bulk Richardson number, \( Ri_b \), using the SST (taken from the sensor 4 m below the sea surface), the thermometer at 13 m, and the wind speed measured at 15 m height. The bulk Richardson number is defined as

\[
Ri_b = \frac{g z (\Delta \theta)_{13m-(-4m)}}{\theta_{13m} u_{15m}^2}
\]
L can then be calculated from the bulk Richardson number 2.3 following Grachev and Fairall (1996), using

\[
\frac{z}{L} = C_1 Ri_b \quad \text{Unstable}
\]
\[
\frac{z}{L} = \frac{C_2 Ri_b}{1 - C_3 Ri_b} \quad \text{Stable}
\]

where \( C_1 = C_2 = 10, C_3 = 5 \).
Figure 20: Høvsøre test station: monthly stability distribution (bar graph, upper) and yearly distribution of the stability distribution (pie chart, lower). Left hand integers denote number of observation in stability class (legend, lower).

Figure 21: Horns Rev wind farm: monthly stability distribution (bar graph, upper) and yearly distribution of the stability distribution (pie chart, lower). Left hand integers denote number of observation in stability class (legend, lower). The data is from the period December 2003 to November 2004, for the 225°-315° wind direction sector. L is calculated from the bulk Richardson number following Grachev and Fairall (1996)
2.4 Phi function at Høvsøre

The objective here is to study how the seasonal behaviour of wind profiles at Høvsøre differs from the traditional descriptions of the wind profile. As stability plays an important role in this discussion, this section will start with an investigation of how the observed normalized wind profiles at Høvsøre differ from the traditional stability-corrected profiles.

Traditional stability correction in the homogeneous surface layer can be done in terms of a correction to the non-dimensional wind shear $\phi$ defined as:

$$\phi(z) = \frac{\kappa}{L}$$

where $L$ and the friction velocity are defined as in 2.1 and 2.2. In order to evaluate $\phi$ from the observations, a procedure suggested by Högström (1988) is used. It is based on a fit of the wind speed observations to a second order polynomial in $\ln(z)$:

$$u(z) = u_0 + A\ln(z) + B[\ln(z)]^2$$

Differentiating with respect to $z$ gives

$$\frac{\partial u}{\partial z} = \frac{A + 2B \ln(Z)}{Z}$$

and accordingly

$$\phi_m(z) = \frac{k}{u_*} (A + 2B \ln(z))$$

In order to solve 2.6 for the 3 unknowns, $u_0, A, B$

measurements of the wind speed at three or more heights are required. Equation 2.5 is valid for the homogeneous surface layer and the lowest measuring heights from the mast (10 m, 40 m, and 60 m) are therefore chosen to ensure that the surface layer assumption is violated as little as possible. The algebraic system of equations is, in accordance with 2.6,

$$\begin{pmatrix}
1 & \ln(10m) & (\ln(10m))^2 \\
1 & \ln(40m) & (\ln(40m))^2 \\
1 & \ln(60m) & (\ln(60m))^2
\end{pmatrix} \begin{pmatrix}
U_0 \\
A \\
B
\end{pmatrix} = \begin{pmatrix}
u(10m) \\
u(40m) \\
u(60m)
\end{pmatrix}$$
The fitted profiles evaluated from 2.6 are plotted against the measured profiles and good agreement is found by inspection. \( L \) is corrected for crosswind contamination following Kaimal and Gaynor (1991):

\[
\overline{w'T'} = \overline{w'T'}_{uncorrected} + 2u_z \frac{u'w'}{406}
\]

Numerous expressions for the \( \phi \) function exist in the literature and a choice has to be made. In this study, the functions from Panofsky (1984) are used:

\[
\text{for } L < 0: \quad \phi_m = \left(1 - \frac{16z}{L}\right)^{-1/4}
\]
\[
\text{for } L > 0: \quad \phi_{mstable} = 1 + \frac{5z}{L}
\]

2.4.1 Discussion on \( \phi_{west} \) versus \( \phi_{east} \)

The stability corrections to the non-dimensional wind shear are valid for the homogeneous surface layer, where the surface layer can be estimated to be the lowest 10% of the boundary layer and is defined as the vertical region of the boundary layer where the Reynolds stress is approximately constant.

First, a test is undertaken to validate that the Högstrom (1988) procedure is able to reproduce the traditional \( \phi \) functions. Data are extracted as described in Section 2.1.1, but for the 30°-90° wind direction sector only. Atmospheric conditions coming from this direction sweep over a landscape with a high degree of homogeneity and meet fairly well the criterion for the prediction of the surface layer wind profiles using traditional surface layer formulas. The evaluated \( \phi \) functions are plotted together with the Panofsky expression for \( \phi \) functions (Equation 2.8) in Figure 21 and reasonable good agreement is found by inspection.

This study focuses on the wind coming from the sea, defined as the 260°-280° wind direction sector. The horizontal homogeneity criterion is therefore violated by the presence of an upstream coast and a number of interesting features in terms of differences to traditional \( \phi \) functions are detectable in Figure 23. The observed features will be described in terms of over-speeding or under-speeding atmospheric conditions. Over-speeding occurs when the wind shear is larger than expected from traditional stability corrections, denoted by the green line; over-speeding is represented by black dots situated above the green line. Conversely, under-speeding occurs when the wind shear is smaller than expected from traditional stability corrections, and is shown as black dots situated below the green line.

Significant under-speeding in the westerly sector is found in the convective region \((z/L < 0.1)\), while the neutral part of the plot is characterised by over-speeding. In the stable regime, the \( \phi \) function splits up in two branches. The lower branch is strongly under-speeded compared to traditional wind profiles, while the upper branch is following or is slightly over-speeding compared to traditional wind profiles. It is suggested that this behaviour is seasonally dependent and relates to the observed time lag in stability distributions between a site representative for the upstream MABL and the coastal Høvsøre site as depicted in Figure 20 and Figure 21. The data are therefore divided into seasons and the seasonal dependency of the signal is investigated and discussed in the following sections.
Figure 22: Phi functions of non-dimensional wind shear evaluated from wind profiles from the north-easterly wind sector (30°-90°)

Figure 23: Phi functions of non-dimensional wind shear evaluated from wind profiles from the westerly sector (260°-280°)
2.5 Seasonality in over- and under-speeding of the wind

The average difference between the observed $\phi$ functions and the Panofsky $\phi$ functions are evaluated for each season and stability bin in order to obtain the seasonal dependence of the over-/under-speeding in the lowest 10-60 m surface layer and are denoted mean difference, defined in 2.9

$$\text{Mean difference} = \frac{1}{N} \sum_{N} \left[ \phi_{\text{observed}} - \phi_{\text{Panofsky}} \right]_{\text{Season Stability bin}}$$

Figure 24: shows the seasonal change in average difference between observed and traditional $\phi_m \left( \frac{z}{L} \right)$ evaluated on measurements from heights 10, 40, and 60 m

Figure 24 shows a clear pattern. The spring and winter seasons are always more windy than summer and autumn. Stability bin 2 shows the winter and spring to be over-speeded, and summer and autumn to be in accordance with the traditional $\phi$ functions. In stability bins 3 and 4, all seasons are over-speeded while stability bin 5 shows over-speeding for winter and spring and under-speeding for summer and autumn. Stability bin 6 shows pronounced under-speeding for all seasons.

It is surprising that stability bin 1 show under-speeding for all seasons, given that, according to the discussion of neutral internal boundary layers in chapter 1, a linear matching of wind profiles between the upstream, windier profile and the local and less windy profile would add an over-speeding component to the wind profile in all stability bins as long as the vertical region where linear matching is undertaken is between 10 m and 60 m where the $\phi$ functions are evaluated.

According to the discussion in chapter 1 and as stated in 1.25, the internal boundary layer height is a function of surface heat flux in such a way that the more heat is added, the higher the internal boundary layer gets and therefore also the area of linear matching of the profiles. This argument calls for the linear matching to take place at higher levels in stability bin 1; exactly at what levels, will for now be unaddressed, but even the unrealistic case of an internal boundary layer of more than 600 m would imply no linear matching and accordingly the evaluated $\phi$ functions
should agree with the traditional ones from Panosky (1984). Here it is less and that is surprising and unaccounted for. An explanation could be that it is windier over land at low levels than over the water, as reported by Ogawa and Ohara (1985) and Bergström et al. (1987). The possible physical principle behind the speed-up over land is not obvious, but will be further address in Chapter 3.

The increase in wind speed as the air moves inland appeared in the numerical simulations for a number of the inland convective cases. An example can be seen on Figure 25, which shows how the upper curvature shifts from being positive between 20 and 50 m in the offshore profile (dashed green) to being negative in the inshore profile (solid red line); this relates to a change of sign in B in Equation 2.6 from positive to negative, thereby reducing the $\phi$ functions. Further investigations should be undertaken to fully account for the under-speeding behaviour in bin 1; however, this falls beyond the scope of the present report. Indication are however given that the under-speeding phenomena is related to inland speed-up.

Another possible explanation for the observed under-speeding in stability bin 1 relates to the development of sea breezes. It is well known that the sea breeze structure includes a pressure gradient force at the surface oriented from the sea towards land, decreasing with height, and eventually reversing to be oriented from land to sea. These considerations follow from the integration of the hydrostatic approximation over a heated land.

The shallow sea breeze circulations are therefore a possible player behind the observed reduction of the wind speed and therefore the lowering of the shear seen in the $\phi$ functions at levels between 10 and 60 m. At the other end of the stability range, under-speedings are also found for all seasons but a more physically comprehensible explanation can be given. The situation here relates to a neutral MABL sweeping across a relatively cold land surface, creating strong stability at low levels, while the upper levels remains less stable, adding to a negative curvature and reducing the $\phi$ functions. The observed and Panofsky $\phi$ functions are plotted for each season and can be found in the appendix.

Figure 25: Inland speed from numerical simulations case $T_{\text{air}}=277$ K, $T_{\text{Sea}}=277$ K, $T_{\text{land}}=283$ K, Free atmosphere wind speed =10 m/s
2.6 Seasonal change of the normalized wind profiles at Høvsøre

The seasonality in wind shear discussed above can be seen when looking at the normalized wind profiles averaged for each stability bin and for each season as shown in Figure 26. The profiles have been normalized by the measured momentum flux at 20 m according to Equation 2.2.

The normalized bin-averaged wind profiles for the entire mast are clearly seen to have the same seasonality as shown above with significantly more wind in the spring and winter months than in the summer and autumn. Each stability bin has 20% to 30% difference between the normalized wind speed at 100 m in the spring and winter as compared to summer and autumn. The difference increases when going to higher altitudes and in addition there is a trend when going to the more stable bins being most extreme in bin 7 (will be shown in section 3.5) with a difference of 50%. Bin 7 has only 12 members for the winter and 6 for the summer and suffers therefore of a low statically significance. The number of winter members in the convective bin was to low and are not include in the analysis. Referring to the internal boundary layer discussion in Chapter 1, the equilibrium layer is seen to extend from the ground to 40 m. The slope for each season for each bin is identical from 10 m to 40 m, supporting the observations. The offset between seasons in the heights between 10 and 40 m is interpreted as seasonality in roughness length. The grassland upstream of the mast has higher vegetation in the summer than in the winter, giving rise to a rougher surface in the summer than in the winter. Profiles of wind speed, TKE, and potential temperature for all members in the calculated means for each season are plotted and can be found in Appendix.

![Normalized mean windspeed for each season and stability bin](image)

**Figure 26: Stability-bin-averaged and normalized wind profile for each season. The winter season is poorly represented in the convective bins and absent in bin 1.**

The objective of the numerical experiment is to reconstruct the seasonal pattern in the normalized wind profiles seen in figure 24 and identify the main physical driver behind this seasonal signal.
2.7 Seasonal change in the upper variance

The heights above 40 m were shown in the preceding discussion to have more wind in the winter and spring than in the summer and autumn for all stability bins. Independently of these observations, spring and winter are also the seasons with the highest amount of variance calculated at the 100 m level according to Equation 2.10. The variance can be seen in appendix, where all the contributing profiles are plotted together with the mean. The spread of profiles around the mean is related to the variance. Especially the summer shows profiles clustered closely around the mean. In the winter and spring, numerous low-level jets are seen. A low-level jet adds to the variability at 100 m in two ways. If the jet is situated above 100 m, positive differences relative to the mean wind speed are attained, and if the jet is located below, negative contributions are attained. Whether the contributions from the jets are on the positive or negative side of the mean, the jets add positively to the variance as the differences are squared. The jets and the seasonal variability in the variance are the subject of Section 3.4, where the influence of an upstream land on the downwind profile is investigated for typical summer and winter conditions and clear differences are found supporting the findings in the observations depicted in Figure 27.

\[ \sigma^2 = \frac{1}{N} \sum_{i=1}^{i=N} (u(100m) - \bar{u}(100m))^2 \quad 2.10 \]

Figure 27: Seasonality in variance of the wind speed at 100 m. Winter and spring have the largest degree of variance for all bins but the bin 5.
2.8 Logarithmic shape of the mean profiles

The normalized stability-bin-averaged profiles, where no distinction between seasons is made, are shown in Figure 28 below and a very simple behavior appears. The profiles up to 160 m for stability bin 1, 2, 4, 5 and 6 seem to be well predicted by a logarithmic function fitted through the lowest measuring heights. Depending on the area of interest, this simple approach may be applicable to get an average annual wind profile at higher levels based on measurements at 10 m and 40 and extrapolated to 160 m, but large changes in wind energy productions would be experienced over the year due to the seasonality in the upper profiles, as discussed in section 2.6.

Figure 28: The normalized stability-bin-averaged wind profile
3 Numerical model experiment

In the following, results will be shown from numerical simulations, where the model has been used as a numerical laboratory. The objective is to isolate the physical drivers behind the observed seasonal structure that has been described in the previous chapter. Both the physical driver behind the observed 20-50% difference in normalized wind speed between seasons for a given stability class and the seasonal change in variance at 100 m height will be addressed. The atmospheric conditions in each stability class span a period of 4 years and therefore represent an enormous amount of different atmospheric conditions, but having one thing in common - the wind direction is from the westerly sector. The westerly wind sector at Høvsøre hosts extremes by cold outbreaks of air masses originating from arctic areas moving through the Norwegian Sea and the North Sea to air masses where a high pressure system over central Europe drives air masses, originating from subtropical areas, over central Europe and into the North Sea as outlined in Figure 29. When the wind direction is straight westerly all the way from the North Atlantic, warm air, relative to the SST, is advected into the North Sea region in the winter and cold air, relative to the SST, is advected into the North Sea region in the summer. This distinction gives rise to different atmospheric conditions in the MABL, where low and strong capping inversions occur when the atmosphere is cooled by the surface and weak (if any) inversions occur when the atmosphere is heated by the surface. As described in chapter 1, the situations where the air masses are cooled by the sea surface can host non-stationary conditions such as initial oscillation Andreas (2000), Burk and Thomson (1995) B. Källstrand (1998), that cause changes in the upper wind profile, that cannot be diagnosed from local surface properties. These non-stationary aspects of the MABL complicate the description of the wind profiles and disqualify the use of Monin-Obukhov scaling in the surface layer, calling for an extended version of this theory for the boundary layer above the surface layer, such as those suggested by Högström et al. (2006) and Gryning et al. (2007), relevant for the description of the wind profiles up to 160 m.

A mesoscale model approach, used in forecast mode with a sufficient amount of vertical computational layer inside the atmospheric boundary layer and a daily data assimilation cycle, hosts the possibility to simulate all atmospheric conditions spanned by the 4 years of data that form the basis of the analysis. These profiles could then be added and averaged within stability classes to possibly obtain the seasonal behaviour of the wind profiles. This approach would be rather expensive in computational time and the amount of data generated would require unfeasibly large amounts of disk storage and if successful would answer if the seasonal behaviour could be reconstructed but not elucidate the physical driver behind. The approach adopted in this study is a simplification to the above in order to identify what physics is at play and set up numerical model simulations in such a way that these physical mechanisms can be isolated.
Figure 29: Examples of warm and cold air masses trajectories to Høvsøre

The coastal meteorology aspect at Høvsøre is studied using the Numerical model COAMPS. It is applied in the simulation of 3 different aspects of the boundary layer:

- Boundary layer build-up over a homogeneous sea surface, discussed in 3.2, addressing the structure of the simulated MABL, which is used as initial conditions for the CIBL simulations.
- Boundary layer build-up over a homogeneous sea surface, downstream from a remote coastline, discussed in 3.4, and addressing the variability in the MABL over the eastern part of the North Sea, depicted in Figure 27.
- CIBL simulations relating to boundary layer build-up over a land surface in the vicinity of an upstream coastline, discussed in 3.5, addressing the seasonality in the normalized wind profiles at Høvsøre, as seen in Figure 26, using results from 3.2 as initial conditions.

The boundary layer over open water often has a long fetch time with nearly homogenous surface forcing and can therefore adjust continually over many hours, while the internal boundary layer over land is the result of an interaction between the incoming marine air and the coastline taking place over a period of only a few minutes. This difference in fetch time causes the two boundary layers to adjust differently to the surface and introduces differences in the wind and temperature profiles.

A distinction between the MIBL and the CIBL is therefore made and these two types of boundary layers are addressed in separate numerical experiments, where the physical processes that act on the wind and temperature profiles in the boundary layers are controlled through the initial conditions and the description of the surface properties. Finally, results from the different simulations are brought together with the objective of reconstructing the seasonal pattern as it is observed and discussed in the previous chapter.

This chapter is, following this argumentation, divided into 4 sections, starting with a description of the parts in the model set up that are shared among the 3 different numerical experiments, and then a separate description of the results from the 3 experiment.
3.1 Model details shared for all simulations

Table 1 Setup details of the numerical model common for all simulations in this study

<table>
<thead>
<tr>
<th>Description</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Number of vertical levels</td>
<td>62</td>
</tr>
<tr>
<td>Model Top</td>
<td>3825 m</td>
</tr>
<tr>
<td>Upper boundary condition</td>
<td>W=0</td>
</tr>
<tr>
<td>Model setup relating to physics and PBL parameterization</td>
<td></td>
</tr>
<tr>
<td>Closure</td>
<td>Mellor Yamada (1982)</td>
</tr>
<tr>
<td></td>
<td>1.5 order level 2.5</td>
</tr>
<tr>
<td></td>
<td>Mixing length: Blackadar (1962)</td>
</tr>
<tr>
<td></td>
<td>Surface fluxes: Louis (1979)</td>
</tr>
<tr>
<td></td>
<td>No moisture</td>
</tr>
<tr>
<td></td>
<td>No radiation</td>
</tr>
<tr>
<td></td>
<td>Surface values kept constant and not allowed to be influenced by surface fluxes</td>
</tr>
<tr>
<td>Heights of computational levels below 1000 m</td>
<td>940</td>
</tr>
<tr>
<td></td>
<td>785</td>
</tr>
<tr>
<td></td>
<td>660</td>
</tr>
<tr>
<td></td>
<td>565</td>
</tr>
<tr>
<td></td>
<td>500</td>
</tr>
<tr>
<td></td>
<td>460</td>
</tr>
<tr>
<td></td>
<td>435</td>
</tr>
<tr>
<td></td>
<td>415</td>
</tr>
<tr>
<td></td>
<td>395</td>
</tr>
<tr>
<td></td>
<td>375</td>
</tr>
<tr>
<td></td>
<td>355</td>
</tr>
<tr>
<td></td>
<td>340</td>
</tr>
<tr>
<td></td>
<td>330</td>
</tr>
<tr>
<td></td>
<td>320</td>
</tr>
<tr>
<td></td>
<td>310</td>
</tr>
<tr>
<td></td>
<td>300</td>
</tr>
<tr>
<td></td>
<td>290</td>
</tr>
<tr>
<td></td>
<td>280</td>
</tr>
<tr>
<td></td>
<td>270</td>
</tr>
<tr>
<td></td>
<td>250</td>
</tr>
<tr>
<td></td>
<td>240</td>
</tr>
<tr>
<td></td>
<td>230</td>
</tr>
<tr>
<td></td>
<td>220</td>
</tr>
<tr>
<td></td>
<td>210</td>
</tr>
<tr>
<td></td>
<td>200</td>
</tr>
<tr>
<td></td>
<td>190</td>
</tr>
<tr>
<td></td>
<td>180</td>
</tr>
<tr>
<td></td>
<td>170</td>
</tr>
<tr>
<td></td>
<td>160</td>
</tr>
<tr>
<td></td>
<td>150</td>
</tr>
<tr>
<td></td>
<td>140</td>
</tr>
<tr>
<td></td>
<td>130</td>
</tr>
<tr>
<td></td>
<td>120</td>
</tr>
<tr>
<td></td>
<td>110</td>
</tr>
<tr>
<td></td>
<td>100</td>
</tr>
<tr>
<td></td>
<td>90</td>
</tr>
<tr>
<td></td>
<td>80</td>
</tr>
<tr>
<td></td>
<td>70</td>
</tr>
<tr>
<td></td>
<td>60</td>
</tr>
<tr>
<td></td>
<td>50</td>
</tr>
<tr>
<td></td>
<td>40</td>
</tr>
<tr>
<td></td>
<td>30</td>
</tr>
<tr>
<td></td>
<td>20</td>
</tr>
<tr>
<td></td>
<td>10</td>
</tr>
<tr>
<td></td>
<td>4</td>
</tr>
<tr>
<td></td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>1</td>
</tr>
<tr>
<td>Initialization</td>
<td></td>
</tr>
<tr>
<td>Sounding of wind speed and potential temperature are used</td>
<td></td>
</tr>
<tr>
<td>for the vertical structure of the mean initial state of the</td>
<td></td>
</tr>
<tr>
<td>atmosphere. The mean initial state is stationary and the</td>
<td></td>
</tr>
<tr>
<td>vertical structure is horizontal homogeneous. Pressure and</td>
<td></td>
</tr>
<tr>
<td>wind speed are in geostrophic equilibrium above the model</td>
<td></td>
</tr>
<tr>
<td>boundary layer. Subsequent perturbations to the mean state are</td>
<td></td>
</tr>
<tr>
<td>generated from a change in the surface fluxes</td>
<td></td>
</tr>
</tbody>
</table>

3.2 Marine atmospheric boundary layer over a homogeneous sea surface

The MIBL approaching the coast at Høvsøre is formed over the North Sea and is the result of multiple processes, as discussed in Section 1.3. In principle, all atmospheric processes are accounted for in the governing equations that are solved by the model. In the following, however, only processes forced by the initial temperature differences between the air and the sea surface are considered. This is controlled by our specification of the surface and initial conditions.

The simulations relates to the situations where an air column characterized by a weak stability is advected over a sea surface and is either cooled or warmed from below. The heat exchange at the surface results in a new physical state of the boundary layer in terms of wind profiles and stability. The situation where the air is colder than the sea surface is representative of a typical summer or autumn day over the North Sea, during which cold air from the North Atlantic passes over the North Sea, resulting in a convective internal boundary layer. The situation is reversed in the winter and spring months, as the North Atlantic air is relatively warm, resulting in the building-up of a stable internal boundary layer over the North Sea.
The stable boundary layer approaches, over time, an asymptotic equilibrium defined by a lower, well-mixed layer capped by an inversion at some height, supporting the ideas of Csanady (1974) and Smedman et al. (1997). Hereafter, the temperature profile in MIBL changes only slowly and the state is therefore referred to as the asymptotical equilibrium. The inversion gets slightly stronger but remains at the same height, while the lower part develops slowly towards a neutral state. The temperature profiles in the MIBL are, as observed by numerous investigators and discussed in 1.5, important for the subsequent CIBL build-up and therefore also important for the wind profiles as observed in the mast at Høvsøre.

### 3.2.1 Model setup

To isolate the effects on the boundary layer structure over the sea surface that are generated by the surface heat exchange, the following model set-up is adopted.

**Table 2: Model setup for MABL simulations**

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Specification</th>
</tr>
</thead>
<tbody>
<tr>
<td>Horizontal domain</td>
<td>16 × 16 grid points</td>
</tr>
<tr>
<td>Lateral boundary conditions</td>
<td>Periodic</td>
</tr>
<tr>
<td>Initial profile</td>
<td>Constant temperature lapse rate = 0.0002 K/m</td>
</tr>
<tr>
<td></td>
<td>Constant wind speed and wind direction</td>
</tr>
<tr>
<td>Surface conditions</td>
<td>Temperature = 277 K</td>
</tr>
<tr>
<td></td>
<td>Constant roughness = 0.0001 m</td>
</tr>
<tr>
<td>Integration time</td>
<td>72 h</td>
</tr>
</tbody>
</table>

### 3.2.2 Choice of initial conditions

To cover the surface forcing that takes place over the North Sea on a yearly basis, a range of initial conditions are selected. The temperature difference between the lowest computational level and the sea surface, and the momentum flux at the lowest level, dictate the cooling or warming rate over a sea surface with a constant temperature and roughness through the Louis (1979) parameterization scheme. In other words, for a given wind speed, the exchange of surface heat increases with the temperature differences between the lowest computational level and the surface, and for a given temperature at the lowest computational level, the heat exchange rate increases with wind speed. The combination of a range in wind speeds and a range in temperature differences between the sea and the atmosphere is assumed to be the main physical driver behind the generation of the MABL over the North Sea and a series of 18 simulations with a limited wind range and a limited temperature range are performed as outlined in Figure 30. The 18 simulated MABL are assumed to span the variety of conditions that occur during the four seasons. This assumption implies that a season can be constructed by a unique weighting of the 18 MABL types.
Figure 30: Shows the air masses that are advected over the sea. Each air mass has a different potential temperature and the simulations are performed with 3 different wind speeds. The temperature of the sea surface is kept constant at 277 K, the potential temperature of the air mass denotes the temperature at the lowest computational level. Temperatures on levels above are calculated using a constant lapse rate of 0.2 K / 1000 m for the potential temperature.

3.2.3 Model results
In the following, the simulated marine boundary layer structure will be discussed in terms of

- Wind speed
- Temperature difference between air and sea
- Fetch/integration time

3.3 Boundary layer structure: Wind speed
To investigate how the structure of the MABL layer depends on wind speed or rather momentum flux, profiles are extracted from the simulations after 36 h of integration. The initial potential temperature of the air mass is the same for the three simulations and in this example is 286 K. Three initial constant wind profile is varied from 10 m/s, to 20 m/s.
Figure 31: Boundary layer stability and height after 36 h of integration time, for various wind speeds

A clear inversion is seen in Figure 31 as a local maximum in lapse rate for the three MABL simulations. Both the height at which this inversion occurs, and also the stability below the inversion, are correlated with wind speed, as expected. The height of the inversion increases with increasing wind speed, from approximately 150 m for an initial wind speed of 10 m/s to 400 m for the initial wind speed of 20 m/s. The proportionality between wind speed and momentum flux is in accordance with the Ekman formulation for the neutral boundary layer depth where the constant of proportionality reads:

\[ h_c = 2ck\pi \left( \frac{u_*}{f_c} \right) \]

\( c \) is constant of proportionality, 0.1

\( f_c \) is the Coriolis parameter, at 57°N found to 1.225*10^{-4} s^{-1} \( \text{Stull (1988)} \)

The constant of proportionality between boundary layer height and momentum flux has been investigated by numerous investigators using various methods and a comprehensive list can be found in the COST 1998 report where the relation between boundary layer height and momentum flux reads:

\[ h = c_{\text{COST}} \frac{u_*}{f_c} \]

\( c_{\text{COST}} \)
To test if the simulations follow the same degree of proportionality, the boundary layer heights from the simulations and the boundary layer heights diagnosed from Equations 3.1, 3.2 and based on $u_*$ from the simulations, are compared in Figure 32. The boundary layer heights plotted in Figure 32 are based on an initial temperature profile with the air temperature at the lowest computational level being identical to the sea surface, 277 K and wind speeds ranging from 10 m/s to 20 m/s. Little agreement is found in terms of absolute values as the Ekman formula predicts a boundary layer height more than a factor of 2 greater than the simulated neutral boundary layer, but a proportionality between momentum flux and boundary layer height is nevertheless found. Better agreement is found when comparing to the boundary layer heights diagnosed from the Coast 1998 constant (Table 3).

Table 3 Constant of proportionality for diagnosing the height of the neutral boundary layer, COST 710 report

<table>
<thead>
<tr>
<th>Author</th>
<th>$C_{COST}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Clarke 1970</td>
<td>0.2</td>
</tr>
<tr>
<td>Benkley and Schulman 1979</td>
<td>0.185</td>
</tr>
<tr>
<td>Mahrt et al. 1982</td>
<td>0.06</td>
</tr>
<tr>
<td>Arya 1981</td>
<td>0.14</td>
</tr>
<tr>
<td>Koracin and Berkowicz 1988</td>
<td>0.07</td>
</tr>
<tr>
<td>Delange 1974</td>
<td>0.04</td>
</tr>
</tbody>
</table>
3.3.1 Boundary layer structure: Air and sea temperature difference

We now examine the influence of the air-sea temperature difference on the structure of the marine internal layer after 36 hours of simulation time. The initial wind speed and the sea temperature is the same in the simulations shown but the temperature of the air is changed to cover the seasonal variability.

Figure 33: Lapse rates for various air-sea temperature differences and an initial wind speed of 15 m/s.

Figure 33 confirms what can be expected intuitively, namely that the stability in the vertical region between the sea surface and the capping inversion is proportional to the difference in air-sea temperature. The height of the inversion is also proportional to the difference in air-sea temperature, ranging from 250 m to 500 m. No capping inversion is obtained in the cases where the air is colder than the sea surface.

3.3.2 Boundary layer structure: Evolution in time

Considered here is how the MIBL evolves in time in terms of the vertical temperature distribution. Figure 34 shows a situation where warmer air (283 K) is blown over a colder surface (277 K) for 36 h. Large gradients in temperature exist in the beginning of the simulations, relating to warm offshore flow in the coastal region. Turbulent heat transport is constantly trying to level out the temperature differences in the MIBL by transporting cold air upward. It is seen in both Figure 35 and Figure 34, how the simulation converges towards an asymptotic equilibrium characterized by low lapse rates in the lower part of the MIBL and strong capping inversions at the upper part of the MIBL, supporting the ideas of Csanady (1974) and Smedman et al. (1997)
3.3.3 **Summary: MABL simulation over a homogeneous sea surface**

It has been shown that wind speed, air-sea temperature difference, as well as fetch time are important factors in the simulation of the marine boundary layer structure in terms of stratification and height of the capping inversion, the latter being interpreted as the height of the internal boundary layer. The simulations showed that for seasons where large air-sea temperature differences prevail, as expected during winter and spring months over the North Sea for atmospheric conditions coming from the westerly sector, low inversions heights are common. It was seen in Figure 31 that for an initial wind speed of 10 m/s and an air-sea temperature difference of 9 K, inversion heights comparable to the height of the met mast, namely 130 m to 150 m, were simulated.

The time of integration, discussed above as fetch time, relate to the time the marine boundary layer spends over the North Sea before it interacts with the coastline. This was seen to partly determine the stability and the structure of the MABL. For the simulation with an initial temperature of 286 K, the lapse rate at 50 m decreased by a factor of 4 between 3 and 15 hours of integration time. Referring to the discussion in chapter 1, and as found by numerous investigators, Melas (1992) and Gryning and Batchvarova (1990), it should be expected that the fetch time spent in strongly stable conditions has a profound impact on the IBL build-up when the MABL passes over a coastline. This argumentation is not relevant to measurement campaigns, as fetch time is a variable that is impossible to control, but when it comes to setting up a numerical experiment, careful consideration should be given to the integration time allowed for the development of the MABL. This issue is further addressed in Section 3.5.

All the simulations in which the initial air temperature was larger than the sea-surface temperature are summarized in Figure 36 in terms of time to reach the asymptotical equilibrium state following the plotting procedure from Smedman et al. (1997). Smedman et al. (1997), using the MIUU hydrostatic model with a second-order closure scheme (Enger 1990), found a linear relationship between the square
root of fetch time and the fractional air-sea temperature difference with a proportionality constant of $S = 75$ (see Equation 1.12). In contrast, a linear fit (green line in Figure 36) is found to be very poorly representative of the data points obtained from the simulations performed in this study, which are much better fitted to a power law with coefficients and resulting sum of errors as outlined in Table 4. Only small errors were made to the fitting procedure when adding the physical argument that the line should pass through the origin. The fundamental difference in correlation between fetch time and air sea temperature differences, outlined in Figure 36, from the simulation done here with COAMPS and the Swedish MIUU model, is likely due to the difference in level of turbulence closure and possibly, to a minor degree, due to the rather subjective methods of determining the time where the simulation has reached its asymptotical equilibrium state.

**Table 4**

<table>
<thead>
<tr>
<th>Fitting method</th>
<th>Smedman et al. (1997)</th>
<th>Power law $ax^b + c$</th>
<th>Power law passing through the origin $ax^b + 0$</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Coefficients</strong></td>
<td>a=91.47</td>
<td>a=11.98</td>
<td>a=7.1474</td>
</tr>
<tr>
<td></td>
<td>b=0.07</td>
<td>c=-6.4</td>
<td>b=0.2585</td>
</tr>
<tr>
<td><strong>Sum of square errors</strong></td>
<td>13.86</td>
<td>0.379</td>
<td>0.39</td>
</tr>
</tbody>
</table>

**Figure 35:** The development of the lapse rate for different initial air temperatures (277 K, 280 K, 283 K, and 286 K) at 10 m (top) and 50 m (bottom). The lapse rates are seen to be larger than 0.05 K m$^{-1}$ at 10 m in the first 3 hours of integration for initial temperature of 286 K while the lapse rate at 50 m, for the same runs remains under 0.05 Kelvin m$^{-1}$ for the first 3 h. The slopes of the lapse rates are seen to be larger at 10 m than at 50 m showing the cooling to be more effective at 10 m than at 50 m but eventually at both heights the slopes approach a constant, which is the asymptotic equilibrium.
3.4 Boundary layer structure and an upstream coastline

In the previous section, the marine boundary layer was developed over an idealized sea with an infinite fetch.

It is speculated that in some situations, this may be too crude an approximation for the North Sea because in spite of being 970 km from north to south and 560 km from east to west, it is bounded by land on all sides except for to the north, which opens up to the Norwegian Sea.

Air masses approaching Denmark from directions other than the north-westerly will therefore have passed over an upstream landmass. The changes that the upstream landmass causes on the structure of the marine boundary layer are investigated at a position a few km west of Høvsøre test station. Simulations are carried out for a series of idealized scenarios with different amounts of reality in the land properties and for a range of land surface temperatures.

The investigation undertaken addresses the development of the MABL downstream of a landmass using a numerical laboratory approach. An alternative and more traditional approach would be to apply the concept of internal boundary layers whereby the model suggested by Sempreviva (1990) and Jensen and Peterson (1977) predicts a lower internal boundary layer reflecting North Sea properties and an upper level reflecting the heterogeneity of the upstream landmass. However, partly due to the long fetch distance, this approach is not feasible, as the internal boundary layer will be fully developed after 10-20 km downstream fetch in the convective case and 100-400 km in the stable case, thereby filling the whole boundary layer. This could lead to the commonly used assumption that the MABL at Høvsøre is solely reflecting the properties of the North Sea. During this study, it is found that the structure of the MABL, in terms of lapse rates, TKE profile and normalized wind profiles, is influenced by upstream conditions. The dependency is found to be enhanced by the surface cooling over the North Sea and is therefore reenforced in

Figure 36: Time for the simulations to reach the asymptotic equilibrium state as a function of the fractional temperature difference between air and sea surface, for various initial wind speeds.
seasons where the surface cooling is a prevailing boundary layer process as is the case in winter and spring (Figure 20). The use of a mesoscale numerical model as numerical laboratory, enables to investigate how much the upstream landmass influences the down stream wind profile for a range of land and sea surface temperature. It is compared to a case where only a sea surface is present, representing the traditional approach, and conditions where the North Sea is relatively cold (spring and winter) or relatively warm, summer and autumn.

The objective of this study is to investigate the seasonality found in the variance at 100 m as described and shown in Figure 27. The hypothesis is that the degree to which the upstream landmass influences the downwind normalized wind profiles varies according to whether the sea surface is colder or warmer than the air. The physical process motivating this hypothesis is explained below.

3.4.1 Model set up

A series of marine boundary layer simulations is performed with a more realistically model setup as compared to the infinite fetch runs presented in the previous section. Each runs host an increasing amount of surface complexity described in the following but all holds an upstream coast as seen in Figure 37. The simulations investigate the perturbation to the initial profile generated by 3 different surfaces for a range of temperatures.

The domain is chosen to cover Denmark at it is easternmost boundary and Scotland at it is westerly boundary with the North Sea in between. Details relevant for this simulation can be found in Table 5. The simulations are integrated for 36 hours in order to ensure that initial artificial transients have disappeared and that the disturbances generated on the upstream coast have had sufficient amount of time to propagate eastwards where they are investigated at position marked by a red circle in Figure 38.

*Table 5 Model details for simulations relating to the influence of an upstream coast*

<table>
<thead>
<tr>
<th>Horizontal domain</th>
<th>81×81 grid points</th>
</tr>
</thead>
<tbody>
<tr>
<td>Boundary condition</td>
<td>Western: Rayleigh damping on the inflow Eastern southern and northern: Radiation condition. Proposed by Orlanski (1976) with the exception that the Doppler-shifted phase speed ((u ± c)) is specified and temporally invariant at the boundary Durran et al. (1993)</td>
</tr>
<tr>
<td>Initial profile</td>
<td>Sounding from Valencia sounding station 03-03-2007 and 14-09-2005 respectively</td>
</tr>
<tr>
<td>Surface</td>
<td>Land sea mask and terrain as outlined in Figure 37 Land surface temperature range as outline in Figure 42 and Figure 52 Land roughness =0.01 m Sea roughness =constant =0.0001 m</td>
</tr>
<tr>
<td>Integration time</td>
<td>36 h</td>
</tr>
</tbody>
</table>
3.4.2 Choice of initial conditions

All runs in one season are initiated on the same sounding from Valencia, situated on the westernmost coast of Ireland close to the western boundary of the domain. Two soundings are used to provide the initial vertical structure of temperature, wind speed and wind directions and are assumed to be representative for a spring and the other for an autumn atmosphere respectively. The wind directions in the chosen soundings are close to west, in the lowest 1000 m of the atmosphere.
To study perturbations in the marine boundary layer generated as a consequence of
1. Upstream curvature in the coastlines,
2. Change of roughness,
3. Change in surface heating/cooling
4. Terrain.
3 parallel runs are done and enables to both study the perturbations generated in each run but also to mutually compare them, in order to point out the most important source to perturbations.

3.4.3 Spring runs
The objective of this series of simulations is to investigate the spring upstream influence on the normalized wind profile in the MABL just before it passes over the coastline to Jutland at the last sea point, (Figure 38). The sea surface temperature of the Atlantic Ocean and the North Sea respectively are kept at 280 K and 277 K and land surface temperatures are change between 277 K, 283 K, and 290 K. The simulation are compared to a run with no land and a SST configuration relating to a spring situation with a Atlantic SST kept at 280 K and North Sea SST kept at 277 K as outlined in Figure 39.

Figure 38: Profiles are extracted from the simulations after 36 h at the position indicated by the red circle at position 56 degree north and 8 degree east. Cross sections follow the direction indicated by dotted lines.
Pure sea run for spring season

Figure 39: Pure sea spring SST configuration. Light blue colour at the right part of domain relates to SST for the Atlantic Ocean and kept at 280 K while dark blue relates to SST for the North Sea kept at 277 K

Figure 40: Shows the surface temperatures configurations for the spring simulations with all land points kept at 277 K while SST as outlined in Figure 39
**Tland=283**

*Figure 41: Shows the surface temperatures configurations for the spring simulations with all land points kept at 283 K, SST as outline in Figure 39*

**Tland=290**

*Figure 42: Shows the surface temperatures configurations for the spring simulations with all land points kept at 290 K, SST as outline in Figure 39*
3.4.4 Boundary layer processes over a cold North Sea

Here follows first a qualitative discussion of the physical processes in the winter and spring. Figure 43 depicts in a schematically way the physical principles behind the frictional decoupling taking place when a well mixed boundary layer makes a transition from a warm rough land surface to a cold and smooth sea surface. When air is moving over a landmass and across a coastline to the smoother sea, the momentum flux at the surface is decreased due to the lower roughness of the sea surface and the increase in stability reduces the turbulence and the associated momentum flux. A speed up will take place close to the top of the internal boundary layer as the negative momentum flux into the box (Figure 43) is smaller than the negative momentum flux out of the box, due to the new level of turbulence above the sea surface. The speed up is influenced by the difference in the atmospheric stability over land and over the sea through to the stability induced change of turbulence levels. When the stability over the water is strongly stable, the down stream flow is known to frictional decouple from the surface. The frictional decoupling of the surface drag creates an imbalance in the equilibrium of forces acting in the boundary layer above. The air in this region is accelerated towards geostrophic equilibrium. It will subsequently start an oscillation around the geostrophic equilibrium as discusses in Blackadar 1957 and Edgar et al. 2000 with a magnitude in wind speed equivalent to the departure from geostrofical wind normally between 2-5 ms⁻¹. The period of the oscillation on latitude 56 N, can be found to be

\[ P_{\text{initial}} = \frac{2\pi}{f_{\text{Coriolis}}} = 14 \text{ hours} \]

This is the physical process studied in the spring simulation below and what is expected to be a player in carrying information of the upstream surface heterogeneity downstream, and in this way add to the structure of the boundary layer in addition to the structure dictated by the sea surface in the area. This process is considered to be an important driver behind the observed seasonality in variance at Høvsøre as discussed in section 2.7 and depicted in Figure 27

Figure 43: Principle in the boundary layer physics in the spring simulations
3.4.5 Boundary layer processes over a cold North Sea simulated

To numerical study how this takes place and how the state of the marine boundary layer evolves an example is given from surface run 2-3 with configuration seen at Figure 42: Shows the surface temperatures configurations for the spring simulations with all land points kept at 290 K, SST as outline in Figure 39. The simulated atmospheric condition relates to a spring day where Atlantic air is sweeping across an Atlantic Ocean with a SST kept at 280 K, before it encounters a heated Scotland with surface temperature kept at 290 K and finally sweeping across a cold North Sea with SST kept at 277 K.

**Figure 44: Wind speed contours ms⁻¹ in a cross section as outlined in Figure 38**

Figure 44 shows an example of wind speed changes when the air passes over Scotland in the west and Denmark to the east. The area of interest is focused on the boundary layer over the North Sea where a large part of the boundary layer, the region below 300 m, is seen to oscillate. The period of this oscillation can be estimated to be close to 14 h indicating, that the variations in the wind speed are caused by the change of momentum flux as the air moves from above the heated ground to above the cold water. Right on the eastern coastline of Scotland, it is seen that the air slows down at the levels below 100 m and this switch off, is compensated by a speed up of the layer above. The upper speed up in off shore flow close to a coastline has been reported by numerous investigators (Højstrup 1999, Lange et al. 2003) who found larger wind speed with height in the surface layer as predicted by stability corrected M-O surface layer scaling for the Rødsand site in Baltic sea for the neutral an stable cases in offshore flow with a 10 km fetch. The authors referred to the speed up as due to a surface discontinuity violating the assumption behind M-O scaling. The numerical model accounts for the land-sea discontinuity and the upper speed up effect referred to, is nicely reproduced. The upper speed up can be explained accordingly as “a frictional cut off” of momentum and is seen to be dominated by the negative buoyancy that acts to reduce the momentum flux at the lowest levels and overcompensating for the smaller loss of friction over the smooth sea surface.
3.4.6 Cross section comparison for spring runs
To show that the landmass upstream cause the flow to oscillate as a function the
excess temperature of the upstream landmass, three examples are shown below. The
wind speed in the tree runs is subtracted from the run with a pure sea surface.

![Figure 45: Surface run 2-1. Color bar (also valid for Figure 46 and Figure 47) relates to wind speed difference between surface run 2-1 and the pure sea run](image)

The flow is seen to be largely unaffected by the landmasses and is similar to the pure sea simulation

![Figure 46: Surface run 2-2](image)

An increase in wind speeds below 80 m is seen over Scotland as the flow encountered a new surface with an increase in surface roughness, slight increase in temperature and whish enhances the turbulence. The wind speed below 80 m is seen to increase somewhat when the air moves inland downstream from the coast line.

The low level speed up is here explained by an imbalance between the frictional loss of momentum in the lower levels and momentum gain, added to the lower levels by an increase in buoyant TKE production, overcompensating the frictional loss by transporting higher momentum values down. This process can explain the upper
level decrease in wind speed. These effects are seen to be confined to region over the landmass and decrease when the air sweeps across the North Sea. For land points kept at 290 K as seen on Figure 47 this process is seen sharper.

Inland speed up over both Scotland and Denmark at levels below 300 m are clearly seen and is likely caused by the frictional loss of momentum that is over compensated by the buoyant TKE production, transporting higher momentum values down.

When the land surface is very warm relative to the incoming air, a kind of initial oscillation is seen to take place over the North Sea. The wind speed in the region over the North Sea, between ground and 340 m height, is seen to oscillate around the values of the pure sea case.
3.4.7 Profile comparison of TKE, wind, and lapse rate, spring case
The profiles from the runs with the coldest and the warmest land temperature from the 4 runs are compared at the position marked by the red circle in figure 36. The objective is to study how the boundary layer is affected by changes to upstream landmass properties.

![Normalized wind profiles extracted for at upstream land kept at 290 and 277 respectively](image)

*Figure 48: Normalized wind profiles extracted for at upstream land kept at 290 and 277 respectively*

The temperature on the upstream landmass is seen to influence the normalized wind profile offshore Høvsøre. The signal is most clearly seen in surface run 1. The normalized wind profile for surface run 1 is nearly identical with the pure sea runs when the upstream land is coldest (277 K) while the profile are significantly slowed down when the upstream landmass is at its warmest (290 K). Similar pattern can be seen in the behaviour of surface run 2. Surface run 3 run(with topography) is seen to have the opposite behaviour.

![Turbulent kinetic energy](image)

*Figure 49: Turbulent kinetic energy*

The features from the normalized wind profile are also visible in the TKE profile. The results from the cold upstream landmass, surface run 1, and the pure sea case are
seen to be identical while conspicuous differences are seen when the upstream land is heated. Surface run 3 are again seen to have an opposite behavior with larger differences with a cold upstream land and nearly identical TKE profiles when the upstream land is heated.

The thermal structure of the boundary layer follows the previous findings for the turbulence and normalized wind profiles. Surface run 2 has a fundamental different structure when the upstream land is heated. For the cold upstream land both surface run 1 and 2 are close to identical with the pure sea case.

### 3.4.8 Summery Spring Runs

Upstream surface test showed that the development of the internal marine boundary layer down stream depended to some extent on the upstream landmass and to a larger degree what temperature the landmass was given. The difference from the pure sea case was found to be between the ground and up to some height depending on land temperature as seen in Table 6. The findings here conflicts with the traditional picture with a low level internal boundary layer reflecting North Sea properties and possibly a residual layer above, reflecting the upstream conditions. It is clearly shown that the changes introduced by the upstream coast, effects the whole boundary layer. The influence from the upstream coast is not clearly visible when horizontally comparing the surface wind speeds over the North Sea, but the structure of the boundary layer is changed, due the upstream heated coast as seen in all the profiles having some sensitivity to the description of the land.

---

**Figure 50: Spring Lapse rate**

The thermal structure of the boundary layer follows the previous findings for the turbulence and normalized wind profiles. Surface run 2 has a fundamental different structure when the upstream land is heated. For the cold upstream land both surface run 1 and 2 are close to identical with the pure sea case.

---
Table 6  Downstream comparison for spring run

<table>
<thead>
<tr>
<th>Run</th>
<th>Height of surface influence on Lapse rate meters</th>
<th>Height of TKE deviation from pure sea run</th>
</tr>
</thead>
<tbody>
<tr>
<td>Surface run 1-1</td>
<td>0 m</td>
<td>0 m</td>
</tr>
<tr>
<td>Surface run 1-2</td>
<td>175 m</td>
<td>400 m</td>
</tr>
<tr>
<td>Surface run 1-3</td>
<td>250 m</td>
<td>2000 m</td>
</tr>
<tr>
<td>Surface run 2-1</td>
<td>200 m</td>
<td>175 m</td>
</tr>
<tr>
<td>Surface run 2-2</td>
<td>185 m</td>
<td>220 m</td>
</tr>
<tr>
<td>Surface run 2-3</td>
<td>350 m</td>
<td>2000 m</td>
</tr>
<tr>
<td>Surface run 3-1</td>
<td>200 m</td>
<td>400 m</td>
</tr>
<tr>
<td>Surface run 3-2</td>
<td>175 m</td>
<td>200 m</td>
</tr>
<tr>
<td>Surface run 3-3</td>
<td>160 m</td>
<td>75 m</td>
</tr>
</tbody>
</table>

The heights are found as the first level where the lapse rate and TKE profile deviates from the pure sea run

3.4.9 Autumn runs

The range of surface heat flux added to the upstream atmosphere when the air is passing over Scotland is the same as in the spring simulation. It is controlled by the temperature difference between the surface and lowest sounding level which is 30 m. Therefore is the amount of heat added to boundary layer over the upstream landmass the same for autumn and spring runs but as shall be shown, the effect of the boundary layer process taken place over the North Sea is different in the autumn compared to the spring situation.

The sea surface temperatures of the Atlantic ocean and the North Sea respectively is kept at 288 K and 290 K and land surface are change between 284 K, 290 K, and 297 K.

Pure sea run for Autumn season

Figure 51: Autumn SST configuration in pure sea run yellow colour at right part of domain relates to SST for Atlantic and kept at 287 K while orange relates to SST for the North Sea kept at 290 K
Tland = 297

Surface run 1-3

Surface run 2-3

Surface run 3-3

Tland = 290

Surface run 1-2

Surface run 2-2

Surface run 3-2
$T_{land} = 284$

Surface run 1-1

Surface run 2-1

Surface run 3-1

Figure 52: Surface temperatures in autumn simulations

3.4.10 Cross section comparison for autumns runs
To see that the flow respond differently for a typical autumn surface configuration cross plots of wind speed are shown below for surface run 2-1 2-2, and 2-3. The wind speed in the three runs is subtracted from the run with a pure sea surface. Colour scheme is constant in all plots.

Figure 53: Surface run 2-1: The flow are seen to be unaffected by the presence of land
A decrease in wind speeds below 160 m is seen over Scotland as the flow encounters a new surface with an increase in surface roughness and therefore enhanced turbulence and frictional loss of momentum. The low level slow down in wind speed is expected according to the build up of internal boundary layer over a rough surface. The decrease in wind speed is seen to be confined to the landmass and decreases when the air sweeps over the North Sea.

A complex picture is seen when the land is heated. Both upstream and downstream from Scotland can changes compared to the pure sea case clearly be identified, but the changes fade out over the north sea and return to a situation somewhat close the pure sea run after the 1350 km mark.

3.4.11 Profile comparison of TKE, wind, and lapse rate, Autumn case

The profiles from the runs with the coldest and the warmest land temperature from the 4 runs are compared it the position marked by the red circle in figure 36. The objective is to study how the boundary layer is affected by changes to landmass properties for the autumn situation.
Figure 56: Normalized wind profiles for spring run

Profiles of normalized wind speed are close to identical for all surfaces when land temperature are set to 284 K but departures are found when the land is heated to 297 K. Surface 1 and 2 are seen to over speeds compared to the pure sea case.

Figure 57: Profiles of Turbulent kinetic energy for autumn runs

The profiles for turbulent kinetic energy is seen to almost insensitive to the land surface properties.
Figure 58: Profiles of lapse rates for autumn runs

The lapse rates are seen to be rather insensitive to the surface properties

Table 7 Down stream comparison for autumn runs

<table>
<thead>
<tr>
<th>Surface run 1-1</th>
<th>Height of surface influence on Lapse rate meters</th>
<th>High of TKE deviation from pure sea run</th>
</tr>
</thead>
<tbody>
<tr>
<td>Surface run 1-2</td>
<td>250 m</td>
<td>130 m</td>
</tr>
<tr>
<td>Surface run 1-3</td>
<td>350 m</td>
<td>220 m</td>
</tr>
<tr>
<td>Surface run 2-1</td>
<td>0 m</td>
<td>0 m</td>
</tr>
<tr>
<td>Surface run 2-2</td>
<td>250 m</td>
<td>130 m</td>
</tr>
<tr>
<td>Surface run 2-3</td>
<td>300 m</td>
<td>220 m</td>
</tr>
<tr>
<td>Surface run 3-1</td>
<td>260 m</td>
<td>240 m</td>
</tr>
<tr>
<td>Surface run 3-2</td>
<td>260 m</td>
<td>230 m</td>
</tr>
<tr>
<td>Surface run 3-3</td>
<td>300 m</td>
<td>180 m</td>
</tr>
</tbody>
</table>

Summary autumn runs
A similar pattern as the spring runs is seen in the autumn runs

Summary boundary layer structure and upstream coast line
The objective was to numerically study, if a seasonality in the variability in atmospheric conditions, outlined for the normalized wind speed in Figure 27, could be reconstructed with a numerical experiment.

Taking the mean situation as the situation where the MABL over the North Sea reflects only the sea surface and studying the departures from this state, generated by the presence of land with different temperatures, a clear seasonality is seen. It is seen that the presence of heated land generates larger relative and absolute difference in the profiles for wind, turbulence and lapse rate, from the runs with no land when the SST configuration in the model relates to a spring situation with warmer Atlantic and colder North Sea than the case, relating to the autumn SST configuration with colder Atlantic and warmer North Sea.

In the latter case the MABL developed was after some time seen to end in a slightly convective state with little sensitivity to the land properties and in the former case,
significant higher sensitivity to land properties conditions was found. These patterns are outlined in Figure 59 and do to some extent support the findings in the observations indicating that the increase in atmospheric variability in winter and spring is relating to upstream heterogeneity, in surface properties, that seems to be better carried over a relatively cold sea surface than over a relatively warm sea surface. It can be argued that the amount of simulations done in this study, 9 for each season is insufficient to show a seasonal trend in variance. The 9 runs for each season are based on the same sounding and it might appears that the perturbations from the 9 surfaces to some extents depends on the structure of the initial profile. These issues do however fall beyond the scope of the report but an increase in the confidence in the results could be achieved increasing the amount of initializations and thereby heavily increasing the amount of simulations. These results discussed here were obtain from the first picked profile from the Valencia sounding station with westerly wind profile below 1000 m for both spring and autumn. However one must bear in mind, that the simulated boundary layers are ensemble means and the structure describes here is therefore ensemble means as function of surface properties and represents the most likely boundary layer structure for the given set of mean variables. Meaning that if the initial sounding is representative for the structure of the atmosphere coming to Ireland in the westerly flow, the simulations performed here will relate to how a seasonal representative atmosphere is perturbated from a landmass, that is either colder isothermal or warmer than the air above and how this perturbation is carried downstream.

**Effects from upstream heated landmass**
This effect can be studied looking at surface runs 1 and 2 which show similar behaviour when the temperature on the land is changed as seen in Figure 59. The signal is clear, the higher the temperature over, the larger departures from the pure sea case. This signal is enhanced when the water configuration relates to spring and is weaker when it relates to autumn, indicating that the MABL over a surface in a SST configuration relating to a spring situation is a better carrier of information of upstream heterogeneities than the MABL over a SST in autumn configuration.

**Effects on upstream heated landmass with topography**
The signal is somewhat complicated when adding topography to the terrain. It is seen here that increasing the temperature, reduced the departures from the pure sea case, indicating that processes other than the frictional decoupling are at play. It however found that departures from the spring related SST are larger than the autumn related sst configuration, supporting the seasonality in variance observation is due to the upstream effect being efficiently transported over a relative cold surface than over a relative warm surface.
3.5 Simulating costal Internal boundary layer builds up at Høvsøre.

3.5.1 Introduction

It was demonstrated in the previous section that an air mass with a history of sweeping across a warm upstream landmass, introduced changes to the lower boundary layer turbulence, wind and temperature profile as far as 1000 km downwind. It was also found that the upstream influence had seasonality with a significant stronger sensitivity to an upstream landmass in the winter and spring compared to summer and autumn where the water is warmer than the air. In terms of the interpretation of observations from Høvsøre, the previous findings implies that the MABL approaching Høvsøre does not solely reflects the properties of the North Sea but to some extent also reflects whether the temperature of the upstream land is warmer or colder relative to the North Sea. More accurate simulations of the MABL would possibly be attained including these physical aspects to the simulations of the MABL approaching the Høvsøre site. These findings underlines the complexity and variety of boundary layer processes that influencing the wind profile at a coastal site and encouragement towards an increase in complexity of the MABL simulations is motivated.

It is however the objective of the present study to seek out the main physical driver behind the observed seasonal signal in the normalized bin averaged wind profiles showing an increase above 40 m for winter and spring seasons compared to summer and autumn as depicted in Figure 26.

An increase in complexity including the amount of excess temperature of upstream land temperature relative to the North Sea, as a additional degree of freedom, would increase the number of time consuming high resolution simulations.

The increase in complexity in the simulations might also end up blurring the signal of the physical process behind the seasonality in normalized wind speed, as the cold water induced variability is adding to the signal in the winter and spring. A high degree of idealization and simplification is therefore maintain in order to isolate the
role of the observed time lag in the surface stability distributions in terms of a surface representative $L$, between the Høvsøre site and the upstream Horns Rev site representative for the state of the MABL.

### 3.5.2 Simplification and method

The objective for the study is to construct a normalized bin averaged wind profile for each season reassembling the seasonal pattern discussed in 2.6 and as depicted in Figure 28. This is done by exposing the set of representative MABL to a coastline for a range of temperature differences between land and sea. The mean meteorological fields in the incoming MABL will be perturbed by the presence of the coast in terms of changes to the momentum fluxes generated by the increase in the surface roughness and heat flux due to a shift in surface temperature at the coastline. This procedure results in a number of simulations that are used to construct an average wind profile for each stability bin at a position in the domain relating to the position of the mast at Høvsøre, in terms of distance from the coast. The averaged wind profile for each stability bin is constructed using a weighing of each simulation based on the probability of the upstream MABL to occur in that season. The simulation of representative MABL and how the weighting is performed is described in the following.

**Representative MABL:**

The incoming MABL at Høvsøre is the end result of a complex combination of numerous processes acting over an unknown amount of hours as discussed in section 3.2. In order to limit the degrees of freedoms in the attempt to reproduce the incoming MABL at Høvsøre, simplifications were made to the simulations allowing only processes forced by a set of initial conditions to act in the MABL simulation. The simplification used in section 3.2 is reused here and a set of simulated MABL is constructed, being the end result of processes solely forced by an initial range of air-sea temperature differences and a range of wind speeds. The implications this setup imposes in terms of MABL structure is discussed in section 3.2.

The constructed set of MABL is assumed to be representative for all atmospheric conditions in westerly wind directions occurring over the North Sea throughout the year. The assumption implies that a particular season of atmospheric conditions over the North Sea can be approximated by a weighted sum of the members in the representative set of MABL.

The representative set of MABL is here discussed in terms of the vertical lapse rate of potential temperature. It is well known, that the lapse rate is an important physical property of the incoming MABL for subsequent build up of an internal boundary layer downstream from a coastline as reported by Garratt (1990) Gryning and Batchvarova (1990), Källstrand and smedman (1997) as discussed in section 2.5. The lapse rates in the simulations, was in section 3.2 shown to be influenced by both wind speed, air-sea temperature differences and fetch time and are therefore all parameters, that need to be assessed in order to set up the generation of the set of representative MABL.

The initial air-sea temperature difference and the wind speed are varied through a range of values in order to vary the amount of heat exchange over the surface in the simulated MABL. The seasonal distribution of heat exchange over the sea surface determines the stability of the lower boundary and is assumed to be the main physical driver behind the observed seasonal signal in the normalized wind profiles (Figure 28). The simulated representative MABL are therefore classified according to the ratio between the surface heat flux and momentum flux as denoted by the M-O length and defined in 3.4.
\[ L = \frac{-u_*^3 \theta_{10m}}{k g u T} \]

\[ k = \text{von karman} = 0.4 \]
\[ g = \text{acceleration of gravity} \]
\[ u_* = \text{momentum flux at lowest computational level (Louis 1979)} \]
\[ T_* = \text{heat flux at lowest computational level (Louis 1979)} \]

The surface momentum flux and the surface heat flux are used in the computation of \( L \) for each of the simulated MABL and taken from the model surface parameterization scheme (Louis 1979) described in 1.5.4. The fetch time, being equal to the integration time, relates to how long the air is allowed to sweep across the sea surface. Referring to the previous discussion in section 3.2, it was found that the fetch time strongly influences the thermal structure of the boundary layer. The best approach would therefore be to use the time that the averaged MABL spend over the North Sea, but unfortunately this parameter is not observable and a somewhat artificial decision is made to use 6 h in the simulation of the representative set of MABL. The 6 hours fetch time is in Figure 34 seen to imply that the boundary layer for a positive temperature difference between the air and the sea is in a state with a nearly uniform lapse rate characterizing the thermal structure of the MABL. The short fetch time implies, that the simulated MABL is characterized by being in a more stable state for a given initial air-sea temperature difference, compared to a simulation with the same initial air-sea temperature difference, but for a longer fetch time, allowing a two layer structure to develop in the boundary layer as described in section 4.2. The growth of the downwind internal boundary layer being sensitive to the upstream lapse rate (Källstrand and Smedman 1997) and therefore assumed to be different in the two cases listed here, making the MABL fetch time an important parameter for simulations of the wind profiles, as they are observed at Høvsøre. The optimal solution might be to allow for a range of fetch times but again this implies a dramatic increase in the number simulations and only a fetch time of 6 hours is for that reason considered in this study.

### 3.5.3 Seasonal weighting

The seasonal weighting of the simulated MABL relates to the probability for each MABL to occur in a given season. It is assumed that a MABL is fully characterised by the computed surface \( L \). The probability/weight of the MABL in a given season is then the same as the probability of the computed \( L \) to occur in that season divided by the number of simulated MABL with the same \( L \), expressed in equation 3.5.

\[ \text{Seasonal weight} = \frac{P_{\text{season}}}{N} \]

\[ P_{\text{season}} = \text{Probability for } L \text{ to occur in a given season} \]
\[ N = \text{number of simulations in stability bin} \]

The seasonal distribution of \( L \) as observed at offshore wind farm, Horns Rev and depicted in Figure 61 is used to establish the seasonal probability/weight of the simulated MABL illustrated in Figure 60. As an example, a simulated MABL characterized by a surface \( L \) belonging to the stable bin \( 1 < L < 50 \), where the stability characterized by \( L \), can be seen to occur 43 % of the time in the winter and less than 5 % in the summer. This implies that for the
winter season, the 6 simulated MABL in bin \(1 < L < 50\) equally share the 43% winter probability and are each assigned a corresponding high winter weight, while the same 6 MABL are sharing less than 5% and therefore corresponding low summer weights.

The assumption behind the simulated MABL being fully characterised by the surface \(L\), implies that the properties of a simulated MABL with high momentum flux and associated high heat flux, is similar to a simulated MABL with low momentum flux and therefore low associated heat fluxes as both situation corresponds to the same \(L\). The assumption is physically reasonable in terms of lapse rate in the surface layer, in that the height of the surface layer is proportional to the momentum flux. In terms of lapse rate in the surface layer, this implies that the larger amount of heat added to a high layer has a similar effect on the lapse rates as the smaller amount of heat has in the corresponding lower layer. However when the objective is to predict the normalized wind at heights around 100 m large differences among the two cases are very likely to occur, making this an area where additional accuracy could be attained when a distinguishing between the low momentum flux cases and the high momentum flux cases are made. The probability/weight of the MABL as a function of both stability and wind speed, might be an area of improvement, where the upstream wind speed distribution is used in analogy with the stability distribution.

Figure 60: Seasonal weights for the generated MABL. The syntax for the MABL is listed in Table 8 Syntax for MABL used in figure
Figure 61: Seasonal L distribution at Horns Rev December 2003 to November 2004

Table 8 Syntax for MABL used in figure

<table>
<thead>
<tr>
<th>MABL nr</th>
<th>m01</th>
<th>m02</th>
<th>m03</th>
<th>m04</th>
<th>m05</th>
<th>m06</th>
<th>m07</th>
<th>m08</th>
<th>M09</th>
</tr>
</thead>
<tbody>
<tr>
<td>Wspd</td>
<td>10</td>
<td>10</td>
<td>10</td>
<td>10</td>
<td>10</td>
<td>10</td>
<td>15</td>
<td>15</td>
<td>15</td>
</tr>
<tr>
<td>Air tmp Kelvin</td>
<td>273</td>
<td>276</td>
<td>277</td>
<td>280</td>
<td>283</td>
<td>286</td>
<td>273</td>
<td>276</td>
<td>277</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>MABL nr</th>
<th>m10</th>
<th>m11</th>
<th>m12</th>
<th>m13</th>
<th>m14</th>
<th>m15</th>
<th>m16</th>
<th>m17</th>
<th>m18</th>
</tr>
</thead>
<tbody>
<tr>
<td>Wspd m/s</td>
<td>15</td>
<td>15</td>
<td>15</td>
<td>20</td>
<td>20</td>
<td>20</td>
<td>20</td>
<td>20</td>
<td>20</td>
</tr>
<tr>
<td>Air tmp Kelvin</td>
<td>280</td>
<td>283</td>
<td>286</td>
<td>273</td>
<td>276</td>
<td>277</td>
<td>280</td>
<td>283</td>
<td>286</td>
</tr>
</tbody>
</table>

3.5.4 CIBL at Høvsøre
Each MABL is hereafter exposed to a coastline characterized by a higher roughness, the same, a colder or 2 warmer land surface temperatures as outlined in Figure 62 and Figure 63. The procedure implies that a MABL have an equal probability of being exposed to each of the land surface characterized by 4 different temperatures. The perturbation of the mean variables in the MABL takes place gradually as the air moves inland and is resolved by 30 grid points, situated between the coastline and the position relating to the Høvsøre mast position. On the coastline itself a maximum in surface momentum fluxes is produced and decays slowly as the drag slows down the flow. However situations where the flow below 20 m was accelerated some distance inland before it reached at steady velocity are among the simulated cases. The inland speed up effect when passing from a smooth to rough surface, has been reported in numerous observational studies during strong inland thermal convections.(Ogawa and Ohara 1985) interpreted the inland speed up as a consequence of the land-sea temperature difference and the associated pressure.
gradient from a sea breeze. Bergström et al. (1987) argued that the frictional loss of momentum at the lower levels was compensated by transport of momentum from layers above, generated by the increase in buoyancy production of TKE. The simulations done in this study support the physical argumentation from Bergström et al. (1987) and this explanation is used to explain the behaviour of inland speed up in some of the runs. The inland speed up in some situations underlines the strength of a numerical modelling approach compared to the more simple and universal approaches discussed in 1.5 as some of the observed variety in the littoral region is captured, underlining the coupled nature of the flow response in this region where frictional and buoyancy processes interacts.

Figure 62: Roughness map in CIBL simulations

3.5.5 CIBL Simulation details

Table 9 Details of the CIBL simulation

<table>
<thead>
<tr>
<th>Horizontal domain</th>
<th>120×60</th>
</tr>
</thead>
<tbody>
<tr>
<td>Horizontal resolution</td>
<td>50 m by 50 m</td>
</tr>
<tr>
<td>Boundary condition</td>
<td>Western: Rayleigh damping on the inflow Eastern, southern and northern: Radiation condition. Proposed by Orlanski (1976) with the exception that the Doppler-shifted phase speed ( u \pm c ) is specified and temporally invariant at the boundary Durran et al. (1993)</td>
</tr>
<tr>
<td>Initial profile</td>
<td>Soundings extracted from the representative MABL simulations.</td>
</tr>
<tr>
<td>Surface</td>
<td>Land sea mask as outlined in Figure 62 Surface temperature range as outline in Figure 63 Land roughness ( =0.01 ) m Constant sea roughness ( =0.0001 ) m</td>
</tr>
</tbody>
</table>
Figure 63: CIBL experiment set up

Figure 64: Cross section turbulent kinetic energy in one of the neutral CIBL simulations to illustrate how the model generates an internal boundary layer
3.5.6 Model skills: Wind profile prediction

Figure 65: Observed profile selected averaged for seasons and stability bins

Figure 66: Simulated profiles sorted to stability bins and hereafter seasonal averaged
Figure 67: Shows the simulated profile in their respective stability bin

Model skills absolute comparison

The seasonal pattern, as discussed in section 2.6 and depicted in Figure 26 is to some extent reproduced in Figure 66. The relatively larger wind speeds observed in spring and winter is clearly detectable in all 5 stability bins. Also the trend towards higher wind speed at the top of the seasonal averaged profile with increasing stability is clearly seen and will be further discussed. The height of the equilibrium layer is also seen to be captured quite well to 40 m in all bins, expect for the most stable one where both observation and simulations agree on an equilibrium layer below 40m.

All profiles in the neutral bin, as seen in Figure 67, are found to be in local equilibrium up to 40 m, expect for one profile that is seen to have a significant lower equilibrium level. This profile is simulated with an upstream SST kept at 277 K and initial air temperature 286 K, sweeping across a coast with temperature 283 K. This profile has a high weighting in the spring and winter compared to the summer and autumn explaining the behaviour of the bin averaged seasonal profiles.

The experiment proved skill in predicting the seasonality in the normalized wind profiles for Høvsøre when incorporating information on the upstream stability distribution. Using the time lag in stability distribution in terms of L between the North Sea and Høvsøre made it possible to partial reproduce the observed seasonality indicating some of the important physical processes behind the observation are captured in the idealized setup presented here.

The observed normalized bin averaged wind profiles are, to a reasonable degree, seen to follow the conceptual model suggested by Sempreviva (1990); Jensen and Peterson (1977). The wind profiles from 100 m and up are characterized by a smaller slope and according Sempreviva (1990) and Jensen and Peterson (1977) can the smaller slope be interpreted as an equilibrium with the upstream smoother sea surface. This feature is not captured in the numerical setup and the upper part of the simulated bin averaged wind profile is seen to be over predicted for all seasons and bins. The difference between the simulations and the observations for wind speed above 100m clearly shows that the highly idealized simulations utilities here, does not capture correctly the boundary layer processes, responsible for the MABL structures over the North Sea as seen in the observations above 100m.
The most likely explanation for this mismatch between the observations and the simulations is a possible too low simulated height of the averaged upstream constant flux layer. The conceptual model suggested by Sempreviva (1990) and Jensen and Peterson (1977) assumes logarithmic upwind wind profile dictating a upwind average constant flux layer height of at least 160 m in order to diagnose the characteristically kink feature in the wind profile as seen in the observations from Høvsøre. Referring to the discussing in 3.3.1 and as seen in Figure 33, the simulated heights of the MABL are between 100-900m dictating constant flux layer heights at the order of 10 to 90 m. The simulated constant flux layer heights are therefore considerably lower than, what must be assumed to be the case for the averaged height of the upstream MABL, responsible for the kink in the upper wind profiles in the observations. The under prediction of the constant flux layer height is due to that the chosen range of initial wind speed apparently is to low, causing correspondingly to low a height of the constant flux layer.

Model skills relative comparison
An increase in agreement between the observations and the simulations are found when comparing the relative differences, denoted RD, calculated as outlined in 3.6, 3.7 and depicted in Figure 68. The 100 m height is chosen for a number of reasons. The most important one is that it relates to the height of the internal boundary layer at the position of the observations mast and therefore reflects both local and upstream properties.

Bin 1: \(-100 < L < 50\)
Good agreement is found between simulations and observations for the most convective bin where both spring and summer seasons are well predicted as seen in Figure 68. No observations were available from the winter season. Observed relative difference at 100m between spring and the low wind season, is 24 % while the simulation gave 26 % while the season with the lowest relative difference was observed to be autumn with 4.8 and simulation gave 7.8.
The ratio between simulated relative spring-summer and relative simulated summer-autumn values is also in good agreement with the observed ratio of relative spring-autumn and summer-autumn values as seen in Figure 69. This result is encouraging as both observation and simulation are well presented in this stability bin as can be seen from Table 10.

Bin 3: \(-500 < L < 200\)
The observed seasonal signal is under predicted by a factor of two in the slightly convective stability class. However is the factor two ratio between spring-summer and autumn-summer captured very well by the simulation as seen in Figure 69. Its worth noting that this stability class has 8 simulated profiles are therefore not as well represented in the simulations (Table 10) compared to bin 1, bin 4 and bin 7.
Bin 4: $500 > L > 500$

For the neutral stability class good agreement is again found where both the observations and simulation are well presented. The simulated relative spring departure can be seen to be 25% while the observed spring departure reads 20%.

The winter season is also captured well by the simulations with a simulated relative departure of 23% and the observed relative departure reads 19%. However is the summer season seen to be under predicted as the simulation reads only 2.9% while observations shows 12.5%.

Bin 50 < $L < 200$

Bin 6 is under predicted by the simulations with nearly a factor of two for all seasons – however is the ratio among the seasonal relative departures is capture well by the simulations as seen Figure 69. Bin 6 has like bin 3, 8 simulated profiles and therefore among the stability class with the poorest representations by simulated profiles.

Bin 7 $L < 50$

This stability class is rare at Høvsøre and only a few observations exist for the 4 year period spanned by the observations. Winter is the best represented season with 12 observations existing for this season while 6 observations can be found to represent the low wind summer season. Care must therefore be taken in the interpretation and evaluation of the model skill for this stability class as the observations are poorly represented. However do the simulations predict a relative departure for the winter season of 39% while the observations show 51%. The most stable stability class is the class with the most extreme difference in normalized wind speed at 100 and agreement on this feature is found between the simulations and the observations.

Notes on selection on low wind season

The season with the lowest wind speed was found to be the summer for the 4 year data set from Høvsøre and therefore used as the low wind reference season for the observations. Autumn is seen (Figure 66) to be the season with the lowest wind speed for the simulations. The weighting function responsible for the relative magnitude of the seasonal bin averaged profiles, is based on observations from Horns Rev where autumn consist of September and October, both months with high percentage of convective situations where the missing November would have contributed towards a more stable distribution. Autumn is then, without the stable November, the season with the highest percentage of convective months and accordingly this is the season being the one with the lowest calculated wind speed at 100 m. If observations from Horns Rev were available for the same period as from Høvsøre, summer would probably appear to be the season with most convective atmospheric situation and the lowest wind speed accordingly.
Figure 68: Simulated and observed relative difference between seasons at 100 m found according to 3.6 and 3.7

\[
\text{RD-winter} = \left[ \frac{\text{Winter}_{\text{Wspd}_{100m}} - \text{Summer}_{\text{Wspd}_{100m}}}{\text{Summer}_{\text{Wspd}_{100m}}} \right] \times 100
\]

Simulated profiles 3.6

\[
\text{RD-winter} = \left[ \frac{\text{Winter}_{\text{Wspd}_{100m}} - \text{Autumn}_{\text{Wspd}_{100m}}}{\text{Autumn}_{\text{Wspd}_{100m}}} \right] \times 100
\]

Observed profiles 3.7

Figure 69: Simulated and observed ratios of RD from winter/spring to summer/autumn respectively.
Table 10 Number of observed and simulated wind profiles in each bin

<table>
<thead>
<tr>
<th>Stability class</th>
<th>Season</th>
<th>Bin 1</th>
<th>Bin 2</th>
<th>Bin 3</th>
<th>Bin 4</th>
<th>Bin 5</th>
<th>Bin 6</th>
<th>Bin 7</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Winter</td>
<td>-100</td>
<td>-200</td>
<td>-500</td>
<td>-500</td>
<td>200</td>
<td>50</td>
<td>1</td>
</tr>
<tr>
<td></td>
<td></td>
<td>&lt;L&lt; -50</td>
<td>&lt;L&lt; -100</td>
<td>&lt;L&lt; -200</td>
<td>&gt;L&gt; 500</td>
<td>&lt;L&lt; 500</td>
<td>&lt;L&lt; 200</td>
<td>&lt;L&lt; 50</td>
</tr>
<tr>
<td>Nr. Of Simult. profiles</td>
<td>Winter</td>
<td>0</td>
<td>1</td>
<td>3</td>
<td>414</td>
<td>158</td>
<td>100</td>
<td>12</td>
</tr>
<tr>
<td></td>
<td>Spring</td>
<td>23</td>
<td>74</td>
<td>67</td>
<td>213</td>
<td>57</td>
<td>71</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>Summer</td>
<td>100</td>
<td>112</td>
<td>165</td>
<td>323</td>
<td>109</td>
<td>110</td>
<td>6</td>
</tr>
<tr>
<td></td>
<td>Autumn</td>
<td>12</td>
<td>15</td>
<td>42</td>
<td>343</td>
<td>103</td>
<td>90</td>
<td>8</td>
</tr>
</tbody>
</table>

3.5.7 Observed and simulated height of the internal boundary layer

The profiles of turbulent kinetic energy are here used to give an estimate of the height of the internal boundary layer. According to the definition of an internal boundary, the profile of turbulent kinetic energy will show a kink, at the top of the internal boundary layer, where the turbulence produced on the local surface meets the turbulence produced on the upstream surface. The heights of the kink are identified on the observed and simulated profiles respectively, and used as an estimated of the height of the internal boundary layer for each stability bin.

Table 11 Simulated and observed heights of the internal boundary layer at the mast

<table>
<thead>
<tr>
<th>Internal boundary layer height</th>
<th>Bin 1</th>
<th>Bin 3</th>
<th>Bin 4</th>
<th>Bin 6</th>
<th>Bin 7</th>
</tr>
</thead>
<tbody>
<tr>
<td>Simulated</td>
<td>120 m</td>
<td>120 m</td>
<td>80 m</td>
<td>60 m</td>
<td>40 m</td>
</tr>
<tr>
<td>Observed</td>
<td>100 m</td>
<td>100 m</td>
<td>100 m</td>
<td>100 m</td>
<td>100 m</td>
</tr>
</tbody>
</table>

There was found no difference in the height of the internal boundary layer between seasons neither for the simulations nor in the observations. This indicates that height of the internal boundary layer, estimated from the TKE-profile, is insensitive to the upstream stability or rather the sensitivity to upstream stability is smaller than the vertical resolution in the experiment.

The observations were also insensitive to surface stability while the simulations showed quite strong dependence on the level of surface stability decreasing the height of the IBL as the stability increased. However the height of the internal boundary layer heights, estimated from the normalized wind profiles can in Figure 82 clearly be seen to decrease with increasing stability supporting results from numerous investigators such as Venkatram (1977), Højstrup (1981) and Gamo et al. (1983)
3.5.8 Model Skill: TKE profile prediction

The skills in the numerical setup for prediction the profile of turbulent kinetic energy is poor, compared to the skills found in prediction the normalized seasonal bin averaged wind profiles. In Figure 70 and Figure 71 are shown the simulated and the observed profiles of turbulent kinetic energy normalized by the value at 10 m. Expect for the summer profile in stability class 1 no agreement is found between the
simulations and the observations indication that the upstream stability distribution is not a dominant player behind the structure of bin averaged seasonal TKE profile. However does the setup capture the behaviour of the spring profile in 3-5 out of 5 cases being the profile with highest normalized TKE values.

Table 12 Shows number of TKE profiles cleaned for bad and missing values in each stability bin together with nr. of simulated and seasonal weighted TKE profiles

<table>
<thead>
<tr>
<th>Stability class</th>
<th>Bin 1</th>
<th>Bin 2</th>
<th>Bin 3</th>
<th>Bin 4</th>
<th>Bin 5</th>
<th>Bin 6</th>
<th>Bin 7</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>-100</td>
<td>-200</td>
<td>-500</td>
<td>-500</td>
<td>200</td>
<td>50</td>
<td>1</td>
</tr>
<tr>
<td></td>
<td>&lt;L&lt;</td>
<td>&lt;L&lt;</td>
<td>&lt;L&lt;</td>
<td>&gt;L&gt;</td>
<td>&lt;L&lt;</td>
<td>&lt;L&lt;</td>
<td>&lt;L&lt;</td>
</tr>
<tr>
<td></td>
<td>-100</td>
<td>-200</td>
<td>-200</td>
<td>500</td>
<td>500</td>
<td>200</td>
<td>50</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>50</td>
</tr>
<tr>
<td></td>
<td>v100</td>
<td>v50</td>
<td>v200</td>
<td>v500</td>
<td>v500</td>
<td>v500</td>
<td>200</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>200</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>50</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Season</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Winter</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>43</td>
<td>40</td>
<td>9</td>
<td>0</td>
</tr>
<tr>
<td>Spring</td>
<td>22</td>
<td>74</td>
<td>65</td>
<td>203</td>
<td>56</td>
<td>68</td>
<td>2</td>
</tr>
<tr>
<td>Summer</td>
<td>86</td>
<td>79</td>
<td>112</td>
<td>237</td>
<td>96</td>
<td>103</td>
<td>6</td>
</tr>
<tr>
<td>Autumn</td>
<td>7</td>
<td>11</td>
<td>27</td>
<td>210</td>
<td>47</td>
<td>72</td>
<td>7</td>
</tr>
<tr>
<td>Nr. of Simult. profiles</td>
<td>9</td>
<td>5</td>
<td>8</td>
<td>11</td>
<td>3</td>
<td>8</td>
<td>26</td>
</tr>
</tbody>
</table>
4 Conclusions, outlook and future improvements

The objective of this study was to use a combination of observations of the upstream stability distribution and a numerical model to reproduce the seasonal difference in the normalized wind speeds as observed at Høvsøre and identify the dominant physical process behind.

The model was run in a controlled setup where the time lag in the stability distribution between the upstream offshore mast at Horns Rev and the coastal Høvsøre mast was posed as being the main physical driver behind the signal and a simulation series was completed in order to test the hypotheses.

The experiment resulted in a number of bin averaged normalized wind profile for each season, that could be directly compared to observations. Good agreement was found in terms of reproducing the observed height of the equilibrium layer at the mast that was found to be 40 m in all stability classes except for the most stable class. The upper part of the bin averaged wind profile was however found to be over predicted in the simulations indicating of that the upstream representative MABL did not represent satisfactory the upstream conditions.

The relative differences in the seasonal averaged and normalized wind profiles between a reference low wind season and the 3 remaining seasons, at 100 m were found to be in reasonably good agreement with observations in 3 out of 5 stability bins. The ratio between the seasons in the stability bin with poor skills was found to be well predicted by the method and further confidence was gained in that, a tendency towards an increase in accuracy with an increase in number of simulations was found.

In can therefore be conclude that among the multiple processes taking place in the coastal boundary layer over the North Sea at winds from west, the dominant physical player behind the observed seasonal signal in the normalized wind speed, is likely to be the time lag in surface stability distribution between the coastal site and the offshore site.

The methods proved to be inaccurate in prediction absolute values of the upper normalized wind speed profile as a systematic over predictions was found. A possible explanation for the lack of skills in predicting the absolute shape and magnitude of the upper profiles was discussed as being caused by low heights of the constant flux layer in the MABL simulation. This problem could possibly be solved allowing for a larger wind speed range in the initial wind profiles for the MABL simulations. An improvement is also expected to be attained with an extension to the assignment of probability/weights to the simulated MABL, using a combination of the seasonal wind speed distribution and the seasonal stability distribution. The wind speed is considered to play an important role as it is well known that the wind speed / momentum flux is an important scaling parameter for diagnosing the height of the boundary layer and therefore also the height of the surface layer that again influence the shape of the profiles across an internal boundary layer.

The coastal meteorology research area as it is studied during this work, includes the prediction of the wind profiles above the surface layer. On the contrary to the well validate M-O surface layer scaling and associated proven abilities to predicts the wind profile in the surface layer, no such validated method exist for the region above the surface layer nor a universal scientific understanding of the physical processes at play it this region. Recent attempts to predict upper level wind profiles from surface properties over homogeneous terrain, suggested by Högström et al. (2006) and Gryning et al. (2007) have similarities with the Blackadar (1962) mixing length formulation used in the model. This work contributes in that the use of a full mesoscale model made it possible to account for selected upstream processes in the
coastal area of Høvsøre where the surface properties were relatively simple and to study the to what degree the upstream processes could reproduce observations.

It was shown that the normalized wind profile at and above 100 at a coastal site, in addition to traditional surface properties and the upstream stability distribution also is influenced by remote processes such as boundary layer oscillations triggered by an excess in surface temperature of U.K compared to the North Sea SST as discussed in section 3.4. These finding underlines that the prediction of the wind profiles above the surface layer in the littoral region is complex and involves a number of non local processes. This work presents a method for this task, where the incorporation of a numerical model played a central part in its ability to resolve and carry non-local processes.

The presented method proved itself to be reasonable successful in predicting upper level relative difference from season to season based on a seasonal distribution in the upwind surface stability. It might be used as a tool to estimating averaged wind energy production climate on a seasonal time scale.
5 Acknowledgement

Anne Grethe: I am forever grateful that you were my wise and loving companion on this journey in hilly terrain

Freja: Your arrival has meant the world to me and put everything in perspective.

Jim: Your patience, persistence and insight in the COAMPS code made this work possible - On top of all this, you were a, supportive colleague and friend during my stay at the NRL, Monterrey.

Lotte: Your optimism and encouragement has powered this study

Sven-Erik: You have contributed in so many ways - Maybe the most important contribution was to know exactly when to apply pressure and when not to.

Aksel: Thanks for supportive and educational supervision during quite some years now – It was always pleasant and rewarding to discuss ideas in your office.

Jean François: Valuable scientific comments and loads of improvements to the manuscript.

The scientific community- Thanks.
6 Some acronyms and definitions

AVHRR=Advanced Very High Resolution Radiometer

Capping inversion = Top of the planetary boundary layer, found as the vertical region with a significant larger vertical gradient in potential temperature and is often associated with sharp vertical gradient in relative humidity aerosols and other scalars also.

CIBL= Coastal Internal Boundary Layer

Ceilometer= A ceilometer is a device that uses a laser to determine the height of a cloud base or in case of no clouds it can be used to detect the height of the boundary Layer.

Drag=Known as the friction force and caused by the vertical divergence in momentum flux in a boundary layer.

Implicit Scheme= A integration scheme using information from the future time step as well as present value.

Internal gravity waves= Atmospheric waves propagating both horizontally and vertically, with buoyancy as the restoring force – Observed frequently in upper atmosphere and in capping inversion over upwelling coastal waters.

IBL=Internal Boundary layer.

MIBL=Marine internal boundary layer.

MABL=Marine atmospheric boundary layer.

SST=sea surface temperature.

NWP= Numerical weather prediction.

PBL=Planetary boundary layer.

Upwelling =The rising of cold, usually nutrient-rich waters from the ocean depths to the ocean surface.

Upper lid = Capping inversion denoting the top of the boundary layer. Characterized by a large positive gradient in potential temperature.

NOAA=National Oceanic & Atmosphere Administration.

KNMI=Koninklijk Nederlands Meteorologisch Institute.

Headland = A bit of coastal land that juts into the sea; cape.

The Danish Galathea-3 circum-global expedition was the third of the ship expeditions headed by Denmark. Galathea 3 took place from 11 August 2006 to 27 April 2007. The Ceilometer measurements (Figure 6) shown in this report are gathered in The “ Carbon cycle project “ whish was among the many research project onboard.
### 7 References


Melas D. (1990) “ Sodar estimates of the surface heat flux and mixed layer depth compared with direct measurements” Atmospheric Environ. 24 A pp 2847-2853

Melas, D., Kambezidis, H. D., (1992) “The depth of the internal boundary layer over an urban area under Sea breeze conditions” Boundary-Layer Meteorol. 61 (3) 247-364


Thorpe, A. P. and Guymry, T. h. (1997) ” The noctual jet” Quart, J. R. Meteoroł. Soc. 103 pp 633-653


Venkatram, A (1977) “A model for of internal boundary layer development” Boundary-Layer Meteorology 11 p 419-437

Yamamoto, G., Yokohama, O., and Yoshikado, H. “Structure of the free convective internal boundary layer during Sea breeze” J. Meteol. Soc. Japan 60 1284-1298

8 Appendix

8.1 Observations winter

Figure 72: $\phi$ functions for winter

Figure 73: Profiles of normalized wind, turbulent kinetic energy and differences in potential temperature for the winter season
Figure 74: Bin averaged profiles for winter
8.2 Observations spring

Figure 75: $\phi$ functions for spring

Figure 76: Profiles of normalized wind, turbulent kinetic energy and differences in potential temperature for the spring season
Figure 77: Profiles of normalized wind, turbulent kinetic energy and differences in potential temperature for the spring season

Figure 78: Bin averaged profiles for spring
8.3 Observations summer

979 Profiles for Wind sector 260-280

Figure 79: φ functions for summer

Figure 80: Profiles of normalized wind, turbulent kinetic energy and differences in potential temperature for the summer season
Figure 81: Profiles of normalized wind, turbulent kinetic energy and differences in potential temperature for the summer season

Figure 82: Bin averaged profiles for summer
8.4 Observations autumn

Figure 83: $\phi$ functions for autumn

Figure 84: Profiles of normalized wind, turbulent kinetic energy and differences in potential temperature for the autumn season
Figure 85: Profiles of normalized wind, turbulent kinetic energy and differences in potential temperature for the autumn season

Figure 86: Bin averaged profiles for autumn