RESEARCH ARTICLE

One-way mesoscale–microscale coupling for the simulation of atmospheric flows over complex terrain

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ABSTRACT

The microscale model WINDIE, initially developed for the simulation of neutral atmospheric flows over complex topography, is here extended to the study of stratified atmospheric flows with Coriolis effects, with particular focus on its application to wind farm development projects. The code now uses $E-I$ and $E-\varepsilon$ turbulence models, which have shown to be more adequate than the standard $E-\varepsilon$ model for the simulation of atmospheric flows. The validation tasks include 1D atmospheric boundary layers from the first two cases produced by the Global Energy and Water Cycle Experiment Atmospheric Boundary Layer Study: a stably stratified boundary layer and a diurnal-cycle over land, respectively. To test the applicability of the new code to real situations, a series of simulations were performed of the time-varying atmospheric flow (a 3-month period between February and May 2012) over a moderately complex topography in the Portuguese mainland, using the Weather Research and Forecasting with Advanced Research WRF (WRF-ARW) mesoscale code on a 3 km mesh to produce time-varying boundary conditions for the microscale code, in a dynamic coupling fashion. Comparisons with sonic anemometer measurements at the hill top and with WRF-ARW results from a finer horizontal resolution mesh ($\Delta x, \Delta y = 1/3$ km) showed that the code can adequately simulate real atmospheric flows over complex topography.

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KEYWORDS
atmospheric flow over complex terrain; mesoscale to microscale coupling; Reynolds-averaged Navier–Stokes (RANS); GABLS experiment; resource assessment; site assessment

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1. INTRODUCTION

In the context of the wind power industry, the crucial parameter needed to decide whether to install a wind farm is the long-term wind power resource at a given site. To be able to obtain such data, wind engineers will usually install meteorological masts at strategic locations, typically operating cup anemometers and wind direction vanes for a sufficiently long period of time to obtain accurate, representative wind data. Due to the spatial sparsity of this information, numerical simulations are then performed to complement the observations and produce a detailed wind flow distribution over the whole wind farm area. Traditionally, linear models, e.g. Wind Atlas Analysis and Application Program (WAsP¹), are then used to predict the wind speed distribution throughout the site, taking the representative wind data collected at the meteorological masts as input. This approach has been successfully employed in the installation of the vast majority of the world’s wind turbines.

However, the use of non-linear models has grown exponentially in the wind industry over the past years, complementing or replacing the more traditional linear analyses. Two main reasons have combined to promote this shift in technology. First of all, more and more wind farms have been installed in regions of large orographic complexity, seeking to maximize wind resource by moving up onto hill crests and mountain tops. At these locations, the wind industry has been confronted with more complex flow phenomena in which linear models are unable to predict accurately. As a result, the use of more sophisticated flow models has now been accepted as a more accurate way to identify not only areas detrimental to wind turbine operation due to high turbulence, wind shear or flow inclination but also the wind resource itself, as shown in the
recent editions of EWEA’s Wind Resource Assessment Workshops (EWEA\textsuperscript{2}). Additionally, as the industry has matured over the past decades and the level of refinement of wind studies has increased, there has been an increasing realization that thermal stratification can play an important role in wind farm production and that its effects should not be necessarily glossed over by the usual assumption that the annual average flow at a given site is neutral. As a result, the assessment of the different thermal (stratification) regimes that may occur at a wind farm location is becoming standard (see, e.g. Wharton and Lundquist\textsuperscript{3} and Rareshide et al.\textsuperscript{4}).

The recent years have seen an attempt to tackle these issues by applying non-linear computational fluid dynamics (CFD) models to wind power applications due to their ability to capture some important flow aspects not in reach of linear models, such as flow separation (see, e.g. Palma et al.\textsuperscript{5}). These codes are typically control volume methods that solve the Reynolds-averaged Navier–Stokes (RANS) equations, predicting separated mean and turbulent fields. They require that the flow be driven by realistic boundary conditions at the edges of the computational domain, and these are often difficult to specify in complex topography. The traditional procedure to cope with this problem is to align the numerical mesh with the inlet flow direction, to impose some standard logarithmic velocity profile (together with compatible profiles for turbulent quantities) and to use symmetry conditions at the lateral boundaries and a free slip boundary condition at the top of the domain (see, e.g. Maurizi et al.\textsuperscript{6} and Castro et al.\textsuperscript{7}). This approach is normally adopted when steady-state simulations are being performed for direction-filtered wind conditions based on the information obtained at the meteorological masts. A more flexible approach can be implemented if, e.g. precursor numerical weather prediction (NWP) simulations are carried out using a mesoscale model on a larger domain to produce the necessary boundary conditions for the CFD simulations.

This latter coupled approach proved to be more accurate than non-coupled procedures, as shown, e.g., by Beaucage et al.\textsuperscript{8} It was used successfully by Eidsvik et al.\textsuperscript{9} who used a three-level model system coupling to predict the microscale flow around Værnes airport, Norway. Other examples are those by Perivolaris et al.\textsuperscript{10} and Beaucage et al.\textsuperscript{8} who extracted time-averaged flow conditions from mesoscale simulations to drive a series of neutral regime CFD simulations for the flow over a wind farm. A similar approach is that of Zająckowski et al.\textsuperscript{11} where inflow boundary conditions and vertical profiles for assimilation are provided by a single time frame of a WRF simulation to conduct a neutral regime CFD code. The success of these coupled procedures is mainly because the resulting boundary conditions are more representative of local terrain conditions than the usual idealized profiles. Furthermore, it opens up the possibility of studying a wider range of stratification regimes and accounting for their individual effects.

It is also worth mentioning that the strategy of relying only on NWP codes to produce the wind speed distribution over the wind farm area is gaining momentum but is still impractical due to the computational and turbulence modelling requirements that arise from the high spatial resolutions that would be typically needed. On the other end of the ‘spectrum’, good results are also being obtained using hybrid RANS/large eddy simulations (LES) techniques applied to the simulation of atmospheric flows over complex terrain (see, e.g. Bechmann and Sørensen\textsuperscript{12,13}), although their high computational requirements still impair their use on a daily basis in wind power projects. Additionally, mesoscale models are also working on improving the modelling of flow over complex topographies by using alternative gridding techniques, such as the immersed boundary method of Lundquist and Chow\textsuperscript{14} to reduce the limitations of terrain-following grids on complex terrain.

In this work, the authors extend a neutral stratification $E-\varepsilon$\textsuperscript{1} microscale RANS code developed for the simulation of atmospheric flows over topography Castro\textsuperscript{15} with the inclusion of a dry potential temperature equation and Coriolis effects. Turbulence closure was achieved using a one-equation $E-l$ turbulence model (Delage\textsuperscript{16} and Weng and Taylor\textsuperscript{17,18}) and a two-equation $E-\varepsilon-l$ model (Xu and Taylor\textsuperscript{19,20}) that have both been proven to outperform the standard $E-\varepsilon$ model in atmospheric flows, due to the difficulty of the latter to predict the adequate reduction of the turbulent length scales with altitude (see, e.g. Weng and Taylor\textsuperscript{17} and Aspley and Castro\textsuperscript{21}). The simulations obtained using this coupled approach essentially yield virtual (i.e. calculated) time series obtained on high-resolution meshes, which can be used in several applications within the wind industry field. They may be used, e.g. to investigate particular flow phenomena that have occurred during wind farm operation or to aid in the short-term forecasting of wind farm production combined with appropriate wind turbine models or integrated within neural network or statistical strategies (see, e.g. Landberg et al.\textsuperscript{22} and Foley et al.\textsuperscript{23}). Finally, and perhaps more relevantly, they may be used to complement or replace linear models in estimating local wind resource, turbulence, wind shear and flow inclination.

In the following section, the microscale model is described along with the configuration used in the mesoscale NWP simulations. In Section 3, the microscale code is validated using well-documented flow cases, which includes the two first Global Energy and Water Cycle Experiment (GEWEX) Atmospheric Boundary Layer Study (GABLS) experiments, where the correct modelling of Coriolis and stratification effects is evaluated. In Section 4, the coupling technique is applied to almost 3 months of real flow over complex topography and compared to sonic anemometer data and mesoscale results to determine the performance of the proposed approach. The main conclusions and future work are presented in Section 5.

\textsuperscript{1}Here, $E$ is the turbulent kinetic energy, $\varepsilon$ is its dissipation and $l$ is a length scale, with more detailed descriptions given in Section 2.
2. NUMERICAL TECHNIQUES

In this section, the numerical techniques used in the simulations of the WRF-ARW mesoscale and the WINDIE microscale code are presented.

2.1. Mesoscale WRF code

In this work, version 3.3 of the Weather Research and Forecasting model with Advanced Research WRF (WRF-ARW) solver was used. The model was developed by the National Centre for Atmospheric Research, and a detailed description is available in the paper of Skamarock et al. Information regarding the topography, land–water masks, land use/land cover classification, albedo, etc., was supplied to the WRF model using the GTOPO30 data sets, made available by the US Geological Survey with a horizontal grid resolution of 0.0083° latitude/longitude (≈900 m).

The WRF-ARW configuration used consisted of the Mellor–Yamada–Janjic planetary boundary layer (PBL) parametrization [turbulent kinetic energy (TKE) closure] with the eta similarity surface layer scheme (see Janjic), the full diffusion option \( \text{diff}_{\text{opt}}=2 \) with the TKE 1.5 turbulence model (see Skamarock et al.) and the WRF Single-Moment 6-class Microphysics scheme (Hong et al.). The WRF-ARW boundary conditions were extracted from the NCEP FNL Operational Model Global Tropospheric Analyses. All simulations were performed using two-way nesting. The Mellor–Yamada–Janjic PBL parametrization was selected due to its capability for predicting the turbulent kinetic energy (TKE) and also because of its common use in the study of diurnal cycles (see, e.g. Svensson et al. and Shin and Hong).

2.2. Microscale code

The core of the present unsteady RANS (URANS) code was developed by Castro for the simulation of neutral atmospheric flows over complex terrain, which gave rise to the WINDIE code used in wind engineering applications. In the present work, the code is extended to include stratification and rotational effects (fictitious Coriolis force). The code uses a dry potential temperature transport equation and, in the present study, a one-equation \( E-L \) turbulence model (Delage and Weng and Taylor) and a two-equation \( E-S-L \) (see Xu and Taylor) turbulence model were used. To reduce execution time, the code was parallelized, using techniques identical to those developed by Castro et al.

The same core was already validated for the simulation of neutral atmospheric flows over topography in the paper of Castro et al.

2.2.1. Mathematical model.

WINDIE uses the anelastic, Boussinesq and RANS approximations of the fundamental equations for fluid dynamics. Written in tensor notation for a generic coordinate system, the continuity, momentum and energy (dry potential temperature) equations are

\[
\frac{\partial}{\partial \xi_j} \left( \rho U_j \beta_k^j \right) = 0 \tag{1}
\]

\[
J \frac{\partial U_i}{\partial t} + \rho \frac{\partial}{\partial \xi_j} \left( U_k U_i \beta_k^j \right) = -\frac{\partial}{\partial \xi_j} \left( \rho \beta_i^j \right) + \frac{\partial}{\partial \xi_j} \left[ \tau_{ij} \beta_k^j \right] + Jf \epsilon_{ij3} U_j + J \frac{\rho}{\theta_0} \partial \delta_{ij3} \tag{2}
\]

\[
J \frac{\partial \Theta}{\partial t} + \rho \frac{\partial}{\partial \xi_j} \left( U_k \Theta \beta_k^j \right) = \frac{\partial}{\partial \xi_j} \left[ \frac{1}{J \text{Pr}} \frac{\partial \Theta}{\partial \xi} \beta_k^m \beta_k^j \right] \tag{3}
\]

respectively, with

\[
\tau_{ij} = -\frac{2}{3} \rho E \delta_{ij} + \frac{\mu_t}{f} \left( \frac{\partial U_j}{\partial \xi^m} \beta^m_i + \frac{\partial U_i}{\partial \xi^m} \beta^m_j \right) \tag{4}
\]

where \( \rho \) is the fluid density of a background hydrostatic reference state; \( U_i = (U, V, W) \) the mean velocity vector; \( \tilde{p} \) is the mean pressure deviation to the background reference state; \( \Theta = \theta_0 + \Theta \) the mean dry potential temperature of its background reference state and deviation, respectively; \( g \) is the magnitude of gravity acceleration; \( f \) is the Coriolis parameter; \( E \) is the turbulent kinetic energy; \( \mu_t \) is the eddy viscosity; and \( \text{Pr} \) is the turbulent Prandtl number. The molecular viscous effects were neglected.
The terrain-following coordinate system is defined by transforming the physical Cartesian coordinates into a computational system \((x_i \mapsto x_i')\), where \(x_i' = J \partial x_i / \partial x^k\) and \(J\) are the co-factors and the determinant of the Jacobian matrix of the coordinate transformation (cf. Knupp and Steinberg\(^{30}\)). This transformation simplifies the treatment of boundary conditions and the use of a structured mesh, whereby the physical domain boundaries are coordinate surfaces following the topography.

For the simulations presented in Section 3, a constant Prandtl formulation was used, which is described therein. In Section 4, the turbulent Prandtl number for the stable stratified regimes \((Ri > 0)\) was calculated using the formulation proposed by Anderson,\(^{31}\)

\[
Pr^{-1} = \frac{G}{1 + (Ri/Ri_0)^{\nu}}
\]

with \(G = 1\), \(Ri_0 = 0.94\), \(N = 1\), and the local gradient Richardson number calculated using

\[
Ri = \frac{g}{T} \left( \frac{\partial T}{\partial z} \right)^2 + \left( \frac{\partial V}{\partial z} \right)^2
\]

where \(T\) is the absolute temperature. For unstable regimes \((Ri \leq 0)\), a value of \(Pr = 0.74\) was used in Section 4.

The turbulence models used here shared the same transport equation for the turbulent kinetic energy \(E\)

\[
\frac{\partial E}{\partial t} + \rho \frac{\partial}{\partial x_j} \left( U_j E / \rho \right) = \frac{\partial}{\partial x_j} \left[ \frac{1}{\sigma_E} \frac{\partial E}{\partial x_j} \right] + \nu_T \frac{\partial^2 E}{\partial z^2} + \nu_T \frac{\partial^2 E}{\partial y^2} - \left( \frac{\partial E}{\partial z} \right) \left( \frac{\partial V}{\partial z} \right)
\]

where \(\nu_T\) represents the dissipation rate of \(E\), \(\sigma_E\) is a model constant and

\[
P_E = \nu_T J \sigma_E \frac{\partial E}{\partial x_j} \quad \text{and} \quad B_E = -\frac{g \mu_T}{\theta_0 \Pr} \frac{\partial \theta}{\partial z}
\]

are the mechanical and thermal production/destruction source terms, respectively.

In the \(E-l\) model (see Delage\(^{16}\) and Weng and Taylor\(^{17,18}\)), the eddy viscosity \(\mu_T\) was modelled using

\[
\mu_T = \rho C_l m \sqrt{E}
\]

with \(C_l = a^{1/2} (\alpha = 0.3\) and \(a = 0.182\) in Section 3 and \(\alpha = 0.182\) in Section 4) and

\[
l_m = \begin{cases} \left( \frac{1}{\alpha \pm \gamma} + \frac{1}{\nu_T} \right)^{-1}, & \frac{\partial \theta}{\partial z} < 0 \\ \left( \frac{1}{\alpha \pm \gamma} + \frac{1}{\nu_T} \right)^{-1}, & \frac{\partial \theta}{\partial z} \geq 0 \end{cases}
\]

where \(k\) is the von Kármán constant \((k = 0.4)\), \(L_O\) is the local Obukhov length,

\[
L_O = \frac{(u' w'^2 + v' w'^2)^{3/4}}{g k \beta w' \theta'^3}
\]

and

\[
l_0 = C_l_0 \frac{|U|}{f}
\]

with \(C_l_0\) a constant and \(|U|\) the magnitude of the geostrophic wind vector, \(\beta\) the coefficient of thermal expansion,

\[
u_T \left( \frac{\partial U}{\partial z} + \frac{\partial W}{\partial x} \right)
\]

\[
u_T \left( \frac{\partial V}{\partial z} + \frac{\partial W}{\partial y} \right)
\]
the kinematic shear stress components with the kinematic eddy viscosity \( v_i = \mu_i/\rho \), and

\[
\frac{u_i}{u_j'} = -\frac{v_i}{\rho} \frac{\partial \Theta}{\partial z} \tag{15}
\]

the vertical kinematic turbulent heat flux. The Monin–Obukhov similarity function for the velocity gradient was

\[
\phi_m = \begin{cases} 
(1 - \gamma \frac{z}{L_0})^{-1/4} & , \frac{z}{L_0} < 0 \\
1 + \beta_c \frac{z}{L_0} & , \frac{z}{L_0} \geq 0
\end{cases}
\tag{16}
\]

where \( \gamma = 19.3 \) and \( \beta_c = 4.8 \) are accepted constants (see, e.g. Foken32).

The length scale \( l_0 \) was specified in Section 3 (25, 40 and 270 m for the neutral ABL, stable ABL and diurnal cycle, respectively), and \( C_{l_0} = 0.003 \) was used for the coupled simulations of Section 4.

The turbulent dissipation rate \( \varepsilon \) was modelled using

\[
\varepsilon = \frac{(C_k E)^{3/2}}{l_d} \tag{17}
\]

where

\[
l_d = \begin{cases} 
\left( \frac{\phi_m}{E (L_0)} + 1 \right)^{-1} & , \frac{\partial \Theta}{\partial z} < 0 \\
\left( \frac{1}{E (L_0)} + 1 + \frac{(\beta_c - 1)}{\rho \mu_i} \right)^{-1} & , \frac{\partial \Theta}{\partial z} \geq 0
\end{cases}
\tag{18}
\]

See Weng and Taylor17,18 for a detailed description of the \( E-I \) model.

For the \( E-\varepsilon-I \) turbulence model (Weng and Taylor17 and Xu and Taylor19,20), a turbulent kinetic energy dissipation rate (\( \varepsilon \)) transport equation of the form

\[
J \rho \frac{\partial \varepsilon}{\partial t} + \rho \frac{\partial}{\partial z} \left( U_k \beta_k^l \right) = \frac{\partial}{\partial z} \left[ \frac{1}{\rho \mu_i} \frac{\partial \varepsilon}{\partial z} \right] + J \rho \left( C_{e1} \frac{\alpha E^2}{l_d^2} - C_{e2} \frac{E^2}{l_d^2} \right) \tag{19}
\]

was used, enabling, in conjunction with \( E \), the estimation of a eddy viscosity,

\[
\mu_i = \rho C_{\mu} \frac{E^2}{\varepsilon} \tag{20}
\]

with \( C_{\mu} \) a constant (\( C_{\mu} = 0.09 \) in all cases of Section 3; \( C_{\mu} = 0.033 \) is also used in Section 3.3 and exclusively in Section 4). The \( l_e \) length scale was modelled using

\[
l_e = \begin{cases} 
\left( \frac{\phi_m}{E (L_0)} + 1 \right)^{-1} & , \frac{\partial \Theta}{\partial z} < 0 \\
C_s E^{1/2} / N & , \frac{\partial \Theta}{\partial z} \geq 0
\end{cases}
\tag{21}
\]

with \( C_s \) a constant (\( C_s = 0.36 \) and 0.18 in Section 3 and \( C_s = 0.36 \) in Section 4) and

\[
N = \sqrt{g \frac{\partial \Theta}{\partial z}} \tag{22}
\]

the Brunt–Väisälä frequency.

The two constant sets that were used in the \( E-\varepsilon-I \) turbulence model are presented in Table I. The first column shows the standard set proposed by Launder and Spalding33 (STD), and the second column shows the set proposed by Duynkerke34 (ATMOS). Both sets were used in Section 3 for the code validation, and the ATMOS set was used in the WRF-ARW coupled case of Section 4.

As a final remark on the tuning of the turbulence models, it should be referred that the values of \( C_k \) were obtained from the standard and atmospheric values for \( C_{\mu} \) through the relation \( C_k = C_{\mu}^{1/4} \), whilst \( C_{\varepsilon} = 0.18 \) was found by numerical experimentation with the objective of obtaining a more dissipative behaviour for the \( E-\varepsilon-I \) turbulence model than that given by the \( C_{\varepsilon} = 0.36 \) value proposed in Weng and Taylor.17
2.2.2. Boundary conditions and coupling techniques.

In case of the WRF-ARW mesoscale–microscale coupled simulations, presented in Section 4, the time-varying boundary conditions for the microscale simulations were generated by interpolation of the mesoscale results (WRF-ARW) to the microscale grid, using tri-linear interpolation in space and linear interpolation in time-linear interpolation between two consecutive WRF-ARW history files [three-dimensional (3D) data] that are \( \Delta t_{\text{WRF}} \) minutes apart. The interpolated quantities were the velocity components \( u_i \), the potential temperature \( \theta \), the turbulent kinetic energy \( E \), the reference state density \( \rho \) and the sensible heat flux at the surface. The turbulent dissipation \( \varepsilon \) was not subjected to time-varying Dirichlet boundary conditions, because typical mesoscale simulations do not produce such fields. The adopted procedure for \( \varepsilon \) was linear extrapolation from the interior field. Because the topography description used in the coupled microscale and mesoscale simulations need not necessarily be the same, a procedure was adopted to force the WRF-ARW topography at the lateral boundaries of the microscale mesh and promote a smooth transition to the microscale topography inside a layer of configurable thickness. This procedure was based on the work of Eidsvik et al.\cite{Eidsvik2010} The microscale topography was produced from contour maps of terrain height, with 25 m resolution. The microscale density and potential temperature reference state vertical profiles were obtained by averaging the WRF-ARW results over horizontal planes.

For the coupled simulations of Section 4, relaxation regions adjacent to the lateral and top boundaries were used to improve the adherence of the interior fields of the microscale simulations to the boundary conditions. These zones follow the sponge layer formulation used by Durran and Klemp\cite{Durran1982} and were applied to the velocity, potential temperature and turbulent kinetic energy fields.

Boundary conditions at the ground were implemented using Monin–Obukhov similarity functions for the velocity and temperature gradients that used the re-formulated universal constants based on the KANSAS experiment (see, e.g. Foken\cite{Foken2006}). For all simulations, with the exception of the one-dimensional (1D) barotropic atmospheric boundary layer (ABL) validation cases of Section 3, the pressure field on all boundaries was obtained from linear extrapolation from the interior fields. For 1D cases, the pressure field was specified using the geostrophic wind.

3. PRELIMINARY CASES—CODE VALIDATION

The core of the present URANS code was developed by Castro\cite{Castro1997} and validated for the simulation of neutral atmospheric flows over topography by Castro et al.\cite{Castro2012}. During the development of the present one-way coupling procedure, additional validations tasks were performed, covering stable stratified inviscid non-rotating flows over topography and neutral (with Coriolis), nocturnal and diurnal-cycle flows over the 1D ABL. The first type of flows contains the dynamics that dominate the atmospheric flows above the boundary layer, whilst the other cases concern the behaviour inside the turbulent boundary layer, where turbulence modelling plays a major role in the quality of the results. Despite the importance of all the validation tasks, we only present here a summary of the 1D ABL cases, which cover phenomena more directly linked to the one-way coupling procedure analysis presented here, i.e. cases for the neutral, nocturnal and diurnal cycles of the ABL. The stable and diurnal-cycle cases follow the benchmark cases proposed by the GABLS project (see, e.g. Holtslag\cite{Holtslag2004} for a description of the project and its objectives). All cases were run in three dimensions with \( 3 \times 3 \) control volumes horizontally, and periodic boundary conditions. Results were extracted from the central control volume column.

3.1. Neutral ABL with Coriolis effects

The case presented here is a neutrally stratified and stationary 1D barotropic ABL. The main objective was to observe if the code can adequately reproduce the effects of the Coriolis acceleration on a neutral flow, taking as reference the results of
Results for both present models follow very closely the marked Ekman spiral behaviour. At upper levels, the flow is driven by an imposed barotropic geostrophic wind with a specified surface cooling rate. This case corresponds to one of the available GABLS cases and is based on LES of an Arctic SBL by Kosovic and Curry, where the flow is simulated here. This case is one of the available GABLS cases and is based on LES of an Arctic SBL by Kosovic and Curry,38 where the flow is driven by an imposed barotropic geostrophic wind with a specified surface cooling rate. The simulations were made using a time step of \( \Delta t = 1 \) s.

The present \( E-\epsilon-\lambda \) and \( E-\lambda-\epsilon-\lambda \) models both used a constant turbulent Prandtl number, \( Pr = 0.85 \) and \( l_\sigma = 40 \) m, as in the \( E-\lambda-\epsilon-\lambda \) model of Weng and Taylor.18 In case of the present \( E-\epsilon-\lambda \) model, two values of \( C_g \) (0.36 and 0.18) were tested in the typical characteristic length scale that affects the production term of the \( \epsilon \) transport equation (equation 19). The domain height was \( 4000 \) m, using a mesh of 281 vertical grid nodes (a resolution similar to that of Weng and Taylor18), more concentrated near the lower boundary where the first node was 1 m above the ground. The simulations were made using a time step \( \Delta t = 1 \) s.

Figure 2 shows the vertical profiles for horizontal wind velocity \( (U, V) \), turbulent kinetic energy \( E \) and shear stresses \( \overline{u'w'} \) for 1D barotropic and neutral ABL with Coriolis effects, Weng and Taylor,17 where results from a variety of ABL models are presented for 1D neutral, nocturnal and diurnal-cycle situations.

The flow has a geostrophic wind of \( (U_g, V_g) = (10, 0) \) m s\(^{-1}\), a Coriolis frequency of \( f = 10^{-1} \) s\(^{-1}\) and a surface roughness of \( z_0 = 0.1 \) m. The vertical domain was set to 6000 m, and a mesh of \( nk = 150 \) grid nodes was used. Mesh nodes were more concentrated near the bottom boundary, with the first node 1 m above the ground. The turbulent quantities were initialized using a friction velocity of \( u^* = 0.38 \) m s\(^{-1}\), \( \epsilon = u^*/\alpha \), \( \alpha = 0.3 \), with the friction velocity defined by

\[
u = \sqrt{\frac{\tau_w}{\rho}}
\]  

(23)

where \( \tau_w \) is the shear stress at the surface. The simulations were conducted over a period of 70 h, where the flow already attained a steady state. See Weng and Taylor17 for further details about the numerical experiment.

Vertical profiles of the velocity components \( (U, V) \), turbulent kinetic energy \( E \) and shear stresses \( \overline{u'w'} \) are presented in Figure 1. These models predict a finite depth ABL, the presence of a supergeostrophic wind around \( z = 500 \) m and a marked Ekman spiral behaviour. At upper levels, the \( (U, V) \) velocity components approach the geostrophic wind \( (U_g, 0) \). Results for both present models follow very closely the \( (U, V) \) results of Weng and Taylor,17 with an almost exact match between the two \( E-\epsilon-\lambda \) models.

The turbulent kinetic energy and shear stresses are also presented in Figure 1. These quantities have their maximum magnitudes at the ground and show an intense decrease with increasing altitude inside the ABL. All three quantities show good agreement with the reference results, with the \( E-\lambda \) model predicting a slight reduction in the vertical profile of \( E \), e.g. \( -3\% \) of the reference results at the ground.

Overall, it can be seen that there is good agreement between the present and reference results.

### 3.2. Nocturnal boundary layer

The 1D nocturnal moderately stratified stable boundary layer (SBL) studied by Cuxart et al.37 and Weng and Taylor18 was simulated here. This case corresponds to one of the available GABLS cases and is based on LES of an Arctic SBL by Kosovic and Curry,38 where the flow is driven by an imposed barotropic geostrophic wind with a specified surface cooling rate. This case was also studied by Beare et al.39 where simulations from 11 different LES models were taken to produce a set of reference results used by Cuxart et al.37 The flow under study was simulated during a 9 h period at the end of which a steady-state behaviour was nearly achieved. For further details on the experimental setup, see the paper of Cuxart et al.37

The present \( E-\lambda \) and \( E-\epsilon-\lambda \) models both used a constant turbulent Prandtl number, \( Pr = 0.85 \) and \( l_\sigma = 40 \) m, as in the \( E-\lambda-\epsilon-\lambda \) model of Weng and Taylor.18 In case of the present \( E-\epsilon-\lambda \) model, two values of \( C_g \) (0.36 and 0.18) were tested in the \( I_\epsilon \) characteristic length scale that affects the production term of the \( \epsilon \) transport equation (equation 19). The domain height was \( 4000 \) m, using a mesh of 281 vertical grid nodes (a resolution similar to that of Weng and Taylor18), more concentrated near the lower boundary where the first node was 1 m above the ground. The simulations were made using a time step \( \Delta t = 1 \) s.

Figure 2 shows the vertical profiles for horizontal wind velocity \( (V_h = \sqrt{U^2 + V^2}) \) and potential temperature at the end of the simulation period. The flow evolves to produce a wind profile with a wind maximum near the SBL top and an elevated
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Figure 2. Vertical profiles of horizontal wind velocity (left) and potential temperature (right) at the end of the simulations period. The reference LES results are represented by a shaded area. See Cuxart et al. 37 for details.

Table II. Sensible heat flux and friction velocity.

<table>
<thead>
<tr>
<th>Models</th>
<th>$u_*$ (m s$^{-1}$)</th>
<th>$\overline{w'p'_{s}}$ (K m s$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>LES</td>
<td>0.29; 0.02</td>
<td>-0.012; 0.002</td>
</tr>
<tr>
<td>Research</td>
<td>0.29; 0.03</td>
<td>-0.013; 0.004</td>
</tr>
<tr>
<td>High-order</td>
<td>0.29; 0.03</td>
<td>-0.013; 0.004</td>
</tr>
<tr>
<td>$E-I$</td>
<td>0.270</td>
<td>-0.012</td>
</tr>
<tr>
<td>$E-\epsilon-I$ ($C_s = 0.18$)</td>
<td>0.290</td>
<td>-0.013</td>
</tr>
<tr>
<td>$E-\epsilon-I$ ($C_s = 0.36$)</td>
<td>0.289</td>
<td>-0.016</td>
</tr>
</tbody>
</table>

Average and standard deviation values for categories of models of Cuxart et al. 37 are presented. The last three lines show the present results.

temperature inversion. From the vertical profiles of the horizontal wind velocity, it can be seen that the present $E-I$ model and the $E-\epsilon-I$ model of Weng and Taylor 18 are the ones that show better agreement with LES results band and are almost identical to each other. It can be also seen that the two $E-\epsilon-I$ results are quite different, with $C_s = 0.36$ producing a wind maximum further away from the ground. The reduction in $C_s$ to 0.18 increased the production of $\epsilon$ inside the boundary layer and led to a better agreement with LES. Nevertheless, the $C_s = 0.36$ result falls comfortably inside the group of wind speed profiles presented by Cuxart et al. 37 by the higher-order and research schemes, a range of results that is represented by the Manned Spacecraft Center (MSC) and National Aeronautics and Space Administration (NASA) results.

The LES vertical profiles of potential temperature at the end of the simulation period, shown in Figure 2, reveal the presence of an upper temperature inversion between 150 and 200 m, topped by the constant potential temperature gradient of the outer flow. The present $E-I$ model showed better agreement with the LES results, in particular in the upper part of the boundary layer where it falls in the middle of the LES results band. Below $\approx 140$ m, the present $E-I$ model follows the cooler portion of the LES results band. The present $E-\epsilon-I$ model with $C_s = 0.18$ and the $E-I$ model of Weng and Taylor 18 show quite similar profiles and are also good approximations of the LES results, displaying, however, a cooler boundary layer than LES all the way up to $\approx 160$ m. The higher levels of turbulent mixing predicted by the present $E-\epsilon-I$ model with $C_s = 0.36$ inside the SBL led to a higher altitude upper temperature inversion, in agreement with the position of the wind speed maximum. This upward shift of the SBL height in relation to LES is predicted by the majority of the models presented in Cuxart et al., 37 from which the MSC and NASA models represent a maximum and a minimum, respectively.

Some statistics for friction velocity ($u_*$) and turbulent sensible heat flux at the ground ($\overline{w'p'_{s}}$) are shown in Table II. It can be seen that the present results (last three lines) are within the range of the models presented by Cuxart et al. 37

Overall, it can be concluded that the main features of the nocturnal stable stratified boundary layer were well captured. The $E-I$ and $E-\epsilon-I$ models with $C_s=0.18$ showed good agreement with the reference LES results. The higher turbulent

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mixing of the $C_s = 0.36$ model led to a thicker SBL, when compared to the LES results, a behaviour also present in some of the higher-order and research models shown by Cuxart et al.\textsuperscript{37}

### 3.3. Diurnal cycle

The next validation case corresponds to the second GABLS experiment (Svensson et al.\textsuperscript{27}), an intercomparison study for the diurnal cycle, which is based on two consecutive clear days from the Cooperative Atmosphere–Surface Exchange Study-1999 (CASES-99 Poulos et al.\textsuperscript{40}). The selected CASES-99 period showed a strong diurnal cycle over relatively dry land and without clouds. This GABLS study evaluated results from 30 different single-column models (SCMs), ranging from operational models with first-order closures and coarse grids to research high-order closure models with fine vertical resolutions. The GABLS setup consisted of a constant geostrophic wind forcing, with $U_g = 3 \text{ m s}^{-1}$, $V_g = -9 \text{ m s}^{-1}$ and $f = 0.887 \times 10^{-4} \text{ s}^{-1}$, a prescribed skin temperature and large-scale divergence. The simulation period begins at 16:00 local time (LT) 22 October 1999 and continues for 59 h. Results were analysed for the 34 h period between 20:00 LT 22 October and 06:00 LT 24 October. For a more detailed description of the experiment setup, see the paper of Svensson et al.\textsuperscript{27}

Simulations were performed using the $E$–$l$ model ($C_k = 0.548$ and $C_{\mu} = 0.09$) and $E$–$e$–$l$ model ($C_s = 0.36$ and $C_{\mu} = 0.033$; see Table I for complete listing of model constants) for a mesh of 75 vertical grid nodes on a vertical domain of $H = 4000 \text{ m}$, with the first grid node above the ground placed at $z = 0.5 \text{ m}$, resulting in a mesh with 57 grid nodes below the first 1000 m. A constant Prandtl number formulation was used with $Pr = 0.74$. The time integration was performed using a time step of $\Delta t = 1 \text{ s}$.

The use of different $C_k$ and $C_{\mu}$ values produced appreciable differences for the turbulent kinetic energy fields but only small, almost irrelevant, differences in the other quantities under analysis. As a result, all results presented were obtained using $C_k = 0.548$ ($E$–$l$ model) and $C_{\mu} = 0.09$ ($E$–$e$–$l$ model), except for Figure 4 where the turbulent kinetic energy is analysed for all $C_k$ and $C_{\mu}$ variants. The use of $C_s = 0.18$ in the $E$–$e$–$l$ model produced results very similar to those obtained with the $E$–$l$ model, and, because of that, they were not included in the analysis.

Figure 3 shows time series for the 10 m wind speed and vertical profiles of potential temperature at 14:00 LT. The wind speed observations (left panel) show a large variability, and the numerical results are almost always inside the observational range. During the first night, the present numerical results follow observations taken at the central tower closely. In this period, the LES results predicted higher wind speeds, which was found in Svensson et al.\textsuperscript{27} to be in agreement with the predictions made by SCMs using the first grid node above 5 m (first grid node of LES around 6 m). In the night-to-day transition, around 08:00 LT, there is a rapid increase in wind speed, which was predicted by all models. Furthermore, in the day-to-night transition, around 19:00 LT, the experimental results show a reduction of wind speed that was captured in the simulations. During the day, the wind speed reaches its maximum values, with the peak around 15:00 LT, with the present simulations and LES showing similar results in agreement with the observations obtained at the central tower. The observations for the second night showed lower wind speeds than in the first, a behaviour that was not predicted by the numerical models, as also found in Svensson et al.\textsuperscript{27} The reason for these discrepancies is likely to be the excessive forcing from geostrophic winds: whereas a mean constant forcing was used by the models, soundings and COAMPS 3D simulations showed a decreasing geostrophic wind during that campaign period; see Svensson et al.\textsuperscript{27} for details.

![Figure 3. Time series of observed and modelled (left) 10 m horizontal wind velocity (m s$^{-1}$) and (right) potential temperature (K) at 14:00 LT. Figures show time evolutions of the observed experimental results and LES simulation results of Svensson et al.\textsuperscript{27} and the present numerical results.](image-url)
Considering the temperature profile, right panel of Figure 3, all results show the existence of an unstable surface layer, where the temperature decreases, followed by a well-mixed layer—near adiabatic temperature profiles—that prevails until nearly 800 m, followed by a temperature inversion. Between 600 m and the beginning of the capping inversion, the observations and LES results show the presence of a stable (positive) temperature gradient, consistent with non-local mixing by dry convection, which is not reproduced by the present codes. As suggested by Svensson et al.,27 this is a limitation of the local nature of the TKE-based models. Besides that model limitation, the present codes follow the observations and LES results well, with the $E-l$ model producing a warmer boundary layer.

Time series for turbulent kinetic energy is presented in Figure 4. All results, i.e. observations, present models using two variants of $C_k$ and $C_{\mu}$ (see Figure 4 legend and Table I for a complete list of the constants in use), and several numerical results presented by Svensson et al.27 were averaged over the height of the 55 m central tower (see Svensson et al.27 for details). From the observations in Figure 4, the existence of a marked diurnal cycle can be seen, with higher levels of $E$ during the day and residual levels during the night. The night–day transition occurs around 08:00 LT with a rapid increase in $E$. The turbulent kinetic energy continues to grow during the afternoon, presenting its highest values around 15:00 LT, showing later a rapid decrease, starting around 16:00 LT, which ends in the day–night transition at approximately 18:00 LT. In qualitative terms, the behaviour of the observed turbulent kinetic energy was well predicted by the models, with the exception of the small-scale event around 03:00 LT that none could predict and is probably due to local effects not taken into account by these 1D simulations. Nevertheless, all models overestimate the turbulence at night and underestimate it during the day.

The decrease of $C_k$ and $C_{\mu}$ (Figure 4) led to an expected increase of the turbulent kinetic levels (see, e.g. page 95 of Apsley and Castro21) and to an overall better agreement with the observations, mainly due to the higher $E$ values predicted for the unstable period of the day. The $E-l$ and $E-\varepsilon-l$ models predicted very similar results for the most intense convective period of the day, around 15:00 LT, whilst in the rest of the day, the $E-l$ model showed almost always lower levels of $E$, a better result for the stable (night) periods. The $E-\varepsilon-l$ model with $C_{\mu} = 0.033$ predicted well the rapid increase of $E$ that occurred during the night–day transition, which starts around 08:00 LT. From Figure 4, it can be seen that all variants of the present models fall well inside the range of results presented by Svensson et al.27

3.4. Partial conclusion

The validation cases presented earlier support the conclusion that the WINDIE code can adequately respond to Coriolis and stratification effects. The three 1D barotropic simple cases enabled us to conclude that both $E-l$ and $E-\varepsilon-l$ models performed adequately in the simulation of turbulent atmospheric neutral and stratified flows. Due to the better results obtained in the diurnal-cycle case by the code configurations that used the lower values of $C_k$ and $C_{\mu}$, those configurations will be used in the next case (Section 4), a real atmospheric 3D case over complex topography.

4. ONE-WAY COUPLED MESOSCALE–MICROSCALE SIMULATIONS OVER COMPLEX TERRAIN

The main objective of the study presented in this section was to determine if a one-way coupling procedure between the WRF-ARW mesoscale code and the WINDIE microscale model could provide an adequate framework for the simulation of atmospheric flows over complex topography, with special interest to wind power applications. To that purpose, nearly

![Figure 4. Time series of observed and modelled turbulent kinetic energy $E$ ($m^2 s^{-2}$). Results averaged over the tower height (55 m).](image-url)
3 months of experimental data—from February to May 2012—were used as reference data. These were collected at a meteorological tower equipped with ultrasonic anemometers and temperature/humidity sensors at two heights: ±80 m above ground level (agl). The microscale simulation results were additionally compared with the mesoscale results used to produce the microscale boundary conditions and also with results from mesoscale simulations performed at higher horizontal spatial resolution (9 km), comparable to that used in the WINDIE simulations themselves. It must be stated that the experimental data used in the present study were, at the time of writing, the only data available to the authors with the desired characteristics, i.e. sonic anemometer time series data collected within complex topography.

The site under study, represented in Figure 5, is located in continental Portugal. The site topography is moderately complex, with maximum terrain slopes of 21°, computed from a raster map of the area with 25 m resolution. The terrain altitude around the meteorological tower varies from 500 to 1000 m, approximately, with the mast itself located at an altitude of ±1050 m (cross in Figure 5). The measurement mast was equipped with Thies ultrasonic anemometers (Thies, Göttingen, Germany) model P6007H, measuring the three wind velocity components and acoustic temperature at 45 and 83 m agl, and KPC 1/6-ME/Galtec (Galtec, Esslingen, Baden-Württemberg, Germany) temperature/humidity sensors at 42 and 80 m agl measuring relative humidity (RH). The data were available as time series of 10 min samples (although the instrument measures at 20 Hz, the output frequency of the data logger was 1 Hz), with mean, maximum, minimum and standard deviation statistics for each 10-min averaging period. The sonic anemometers were calibrated following IEC standards. Tower-induced turbulence was not judged significant after data validation, and no filtering was applied.

Information about the wind conditions obtained with the instrumentation installed at the top of the meteorological tower (80–83 m agl) is presented Table III. The campaign period approximately covered the second half of winter and the first half of spring 2012 (see Table III for dates). By analysing Table III, it can be seen that the mean wind speed increased during the campaign period and that the middle period was the warmest. The mean RH increased as the campaign progressed, from an approximately mean dry weather in the first period to humid in the last. Whilst there were no rainfall events during the first period and only six in the second, precipitation occurred almost daily in the third period.

The WRF-ARW simulations were made using two-way nesting and the setup presented in Section 2.1. The WRF-ARW ‘coarse’ grid simulations, used to produce the WINDIE boundary conditions, used a parent and 2 two-way nesting levels, with mesh resolutions of 27 km–9 km–3 km, with the largest domain covering an area of almost 900 km by 900 km. The ‘finer’ grid mesoscale simulations were similar but with two additional nesting levels, of 1 and 1/3 km resolutions. All WRF-ARW simulations were made using a domain top placed at 5000 Pa and a vertical mesh of 33 grid nodes, with the

---

**Figure 5.** Site topography representation. Extension of 20 km × 20 km. Meteorological tower marked by a cross. On top and on the left side, ‘cut’ projections of the topography in the easting and northing directions, passing trough the tower location.
were avoided repetition of (sometimes very long) simulations. The chosen periodicity of the mesoscale and microscale history files techniques for both codes. This procedure will also make the process immune to bad setup options of the times series and so that those series will be made from post-processing the history files of both codes, enabling the use of the same numerical cal mesh. In this work, mesoscale simulations with dimension of the mesoscale grid cells, which determine the cut-off frequency of the low-pass filter behaviour of the numerical conditions, and, after some preliminary numerical experiments, it was observed that the gain in frequency response obtained due to the boundary condition dynamics that make the microscale fields evolve with an almost linear trend in time scales – 20 min. When comparing numerical and measured time series, all data were synced to the reference mesh used for the total campaign period, this is the s = mesh used for the complete campaign and the three sub-periods under study. Microscale simulations covering the complete campaign period were performed using a mesh of 51 × 51 × 41 grid nodes on a domain extension of 20 km × 20 km in the horizontal directions and a top boundary placed at an altitude of 4.5 km. The mesh was almost uniform in the horizontal directions, with a minimum grid spacing of Δx, y = 350 m. In the vertical direction, the mesh was concentrated near the ground, with a minimum control volume height of 10 m near the surface (i.e. control volume centre placed at 5 m agl) and expanding geometrically upwards. The simulations were performed using a time step of Δt = 1.5 s. Selected fields were written to file every 20 min for creation of time series and other post-processing quantities. The topography and roughness information were defined from a local survey of the site, combined with satellite imagery. To analyse the sensibility of the microscale results to the grid characteristics, some preliminary simulations with the E–I model were made for period 3. The characteristics of the meshes can be seen in Table IV, and the analysis of the results are presented in the next section. A detailed description of code configurations and constant sets is presented in Section 2.2.

In this coupling procedure, the microscale code boundary conditions are produced by linear interpolation in time between two consecutive WRF-ARW history files (3D data), which are ΔtWRF apart and whose optimum value is connected to the dimension of the mesoscale grid cells, which determine the cut-off frequency of the low-pass filter behaviour of the numerical mesh. In this work, mesoscale simulations with Δx, y = 3000 m grids were used for producing the microscale boundary conditions, and, after some preliminary numerical experiments, it was observed that the gain in frequency response obtained by using smaller values than ΔtWRF = 20 min was not enough to justify the increase in storage requisites.

For the construction of the mesoscale and microscale time series, some options were considered, and it was decided that those series will be made from post-processing the history files of both codes, enabling the use of the same numerical techniques for both codes. This procedure will also make the process immune to bad setup options of the times series and so avoid repetition of (sometimes very long) simulations. The chosen periodicity of the mesoscale and microscale history files were ΔtWRF = 20 min, with a phase shift of ΔtWRF/2 min between the two. Tests showed that increasing the frequency of WINDIE’s results, e.g. to ΔtWRF = 5 min, did not produce any noticeable differences in the time series, a behaviour due to the boundary condition dynamics that make the microscale fields evolve with an almost linear trend in time scales equal to or less than ΔtWRF = 20 min. When comparing numerical and measured time series, all data were synced to the same instants in time using linear interpolation to a common time vector with 20 min sampling period. This procedure will function as low-pass filter for the observations and compensate the lower frequency content of the numerical results.

The sampling and averaging used in the sonic anemometer data (i.e. 10 min averaging of 1 Hz samples) is assumed, in the field of wind energy, to adequately represent the turbulent scales of the wind speed spectrum, i.e. eddies with

**Table III.** Wind conditions for the complete campaign and the three sub-periods under study.

<table>
<thead>
<tr>
<th>Period</th>
<th>Data (days)</th>
<th>Vₚ (m s⁻¹)</th>
<th>T (°C)</th>
<th>RH (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Min</td>
<td>Max</td>
<td>Mean</td>
<td>Min</td>
</tr>
<tr>
<td>1</td>
<td>11 Feb to 1 March (20)</td>
<td>0.20</td>
<td>19.5</td>
<td>7.2</td>
</tr>
<tr>
<td>2</td>
<td>2 March to 31 March (30)</td>
<td>0.18</td>
<td>21.5</td>
<td>8.6</td>
</tr>
<tr>
<td>3</td>
<td>1 April to 8 May (38)</td>
<td>0.23</td>
<td>22.3</td>
<td>9.3</td>
</tr>
<tr>
<td>1 + 2 + 3</td>
<td>(88)</td>
<td>0.18</td>
<td>22.3</td>
<td>8.6</td>
</tr>
</tbody>
</table>

Wind speed and temperature obtained at 83 m agl and relative humidity obtained at 80 m agl.

**Table IV.** Meshes used in the grid sensibility tests.

<table>
<thead>
<tr>
<th>Mesh</th>
<th>Grid</th>
<th>Δ(x, y)min (m)</th>
<th>Δzmin (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>s</td>
<td>51 × 51 × 41</td>
<td>350</td>
<td>10</td>
</tr>
<tr>
<td>v−</td>
<td>51 × 51 × 29</td>
<td>350</td>
<td>20</td>
</tr>
<tr>
<td>v+</td>
<td>51 × 51 × 47</td>
<td>350</td>
<td>5</td>
</tr>
<tr>
<td>h+</td>
<td>71 × 71 × 41</td>
<td>250</td>
<td>10</td>
</tr>
</tbody>
</table>

s = mesh used for the total campaign period, this is the reference mesh; v− = less grid density in the vertical direction than the reference mesh; v+ = more grid density in the vertical direction than the reference mesh; h+ = more grid density in the horizontal directions than the reference mesh.

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durations between 10 s and 10 min. The turbulence models used in the WRF and WINDIE codes were developed and tuned to represent this same portion of the wind speed spectrum, although some significant differences can be identified in their formulations. In the case of the WINDIE code, the turbulence models aim to capture all the turbulent portion of the spectrum, averaging over an infinite number of independent realizations (Reynolds ensemble averaging), independently of time steps and grid spacing. In the case of the WRF code, the typical PBL models are based on a clear separation between resolved and sub-grid turbulent scales, and the proposed PBL closures assume that the grid size is large enough so that the whole turbulence spectrum has a sub-grid nature. It follows that a comparison between these three ways of describing the ABL turbulence will be somewhat subjective, but it is also the only practical way to assess modelled turbulence results in this context.

Some problems could arise with the finest mesoscale resolution of $1/3$ km used here, which could be said to be inside a region often called ‘terra incognita’, where the length scale of turbulent structures approaches that of the mesh and where Wyngaard identified that the tensor nature of the eddy diffusivity may be relevant. Nevertheless, the fact that at such spatial resolutions the use of the aforementioned PBL closures are not fully justified, a noticeable improvement in results was observed by reducing the mesh resolution from 3 to $1/3$ km, justifying the use of the mesoscale fine mesh in the appraisal of the coupling procedure results. This improvement should not be seen as an argument against the need of a better turbulence modelling at these particular length scales but only that other factors, like mesh resolution, are also very active on the establishment of the results.

Although this work focuses on the validation of the physical models, it is nonetheless important to highlight that the proposed methodology also brings important benefits in terms of computational efficiency. The coupling procedure used here significantly reduced processing time, with the combined wall clock time of a coarse WRF simulation and subsequent WINDIE simulation being always less than half the time necessary to run an equivalent WRF fine grid case, when using the same number of processors. Post-processing of WRF results for the WINDIE simulation runs in parallel with both WRF and WINDIE runs and therefore does not add to the wall clock time.

4.1. Results

The performance of the microscale code and coupling procedure was determined by analysis of horizontal and vertical velocities, $V_h = \sqrt{U^2 + V^2}$ and $W$, as well as the dry temperature, $T$, and the turbulent kinetic energy, $E$. The deviations between numerical results and measured data were characterized using linear regression techniques [intercept ($b$) and slope ($m$) of the regression line and coefficient of correlation $R^2$], the root mean square error ($rms$) and bias ($\epsilon$). To analyse the wind direction, wind roses were used.

Because the microscale code does not include humidity, it was necessary to use a dry temperature (potential or otherwise) to compare the different temperature results. It was decided to use dry temperature (not potential), and the data obtained at the two hygrometers were used to convert the sonic virtual temperature into temperature—the dry sonic temperature being almost identical to the temperature measured at the hygrometers. The conversion from potential temperature to temperature depends on the exact altitude of the monitoring points, and because of the use of different mesh resolutions, some small temperature offsets are expected. Because of this, performance indicators for temperature will be more focused on $m$ and $R^2$, quantities that are not distorted by this offset.

4.1.1. Effect of grid resolution.

Some preliminary tests were performed to analyse the sensitivity of the microscale code results to different grid configurations (Table IV). To the standard mesh ($s$), three other meshes were added: mesh $h+$ with a $\approx 70\%$ reduction in $\Delta(x,y)_{min}$ at the domain centre and increased number of nodes to approximately maintain uniform control volume dimensions; mesh $v+$ where the distance of the first grid node to the ground was halved to $\Delta z_{min} = 5$ m and mesh $v-$ where that same distance was doubled to $\Delta z_{min} = 20$ m. The number of vertical nodes of these last two meshes was adjusted in order to maintain the same vertical resolution of mesh $s$ at the top of the domain, thus preserving the characteristics of the sponge layer at the top domain boundary.

Performance indicators of these preliminary tests are presented in Table V for the $E-\epsilon-I$ turbulence model (same behaviour for the $E-I$ model). It can be seen that increasing the grid resolution ($h+$ and $v+$ meshes) resulted in globally better results, although systematic improvements are hard to clearly identify. The exception occurs in the modelling of the vertical velocity component, $W$, where a higher horizontal resolution ($h+$) has improved all statistics at both heights. This is likely to be due to a better description of the terrain and its slope, which allows for a more accurate calculation of the wind inclination.

Regarding the horizontal wind speed, $V_h$, an increase in vertical resolution ($v+$) has improved both the bias and the $rms$ errors and $R^2$, which could be explained by the model capturing the vertical wind profile more accurately. It should be noted that high correlations for horizontal wind speed are especially useful, since they allow the use of measure-correlate-predict techniques to extend the simulation period to longer periods of measured data, without the need to simulate the entire

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Table V. Statistics for several mesh resolutions in period 3 using the $E-\varepsilon-l$ turbulence model.

<table>
<thead>
<tr>
<th>Period 3</th>
<th>$V_h$ (m s$^{-1}$)</th>
<th>$W$ (m s$^{-1}$)</th>
<th>$E$ (m$^2$ s$^{-2}$)</th>
<th>$T$ ($^\circ$C)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>$b$</td>
<td>$m$</td>
<td>$R^2$</td>
<td>$\epsilon$ (%)</td>
</tr>
<tr>
<td>45 m agl</td>
<td>s</td>
<td>0.937</td>
<td>0.899</td>
<td>0.775</td>
</tr>
<tr>
<td></td>
<td>h+</td>
<td>0.962</td>
<td>0.911</td>
<td>0.772</td>
</tr>
<tr>
<td></td>
<td>v+</td>
<td>0.932</td>
<td>0.895</td>
<td>0.775</td>
</tr>
<tr>
<td></td>
<td>v−</td>
<td>0.920</td>
<td>0.908</td>
<td>0.774</td>
</tr>
<tr>
<td>83 m agl</td>
<td>s</td>
<td>0.928</td>
<td>0.917</td>
<td>0.773</td>
</tr>
<tr>
<td></td>
<td>h+</td>
<td>0.977</td>
<td>0.924</td>
<td>0.771</td>
</tr>
<tr>
<td></td>
<td>v+</td>
<td>0.935</td>
<td>0.912</td>
<td>0.774</td>
</tr>
<tr>
<td></td>
<td>v−</td>
<td>0.900</td>
<td>0.930</td>
<td>0.772</td>
</tr>
</tbody>
</table>

Coefficients $b$ and $m$ belong to the correlation line $y(x) = mx + b$. $V_h$ = horizontal velocity; $W$ = vertical component of the velocity vector; $E$ = turbulent kinetic energy; $T$ = the dry temperature; $R^2$ = coefficient of determination; $\epsilon$ (%) = percentage bias error; rms = root mean square error.
### Table VI. Statistics for the campaign period.

<table>
<thead>
<tr>
<th>Period 3</th>
<th>$V_h$ (m s$^{-1}$)</th>
<th>$W$ (m s$^{-1}$)</th>
<th>$E$ (m$^2$ s$^{-2}$)</th>
<th>$T$ ($^\circ$C)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>$b$</td>
<td>$m$</td>
<td>$R^2$</td>
<td>$\epsilon$ (%)</td>
</tr>
<tr>
<td>45 m agl</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>WRF c</td>
<td>1.400</td>
<td>0.623</td>
<td>0.681</td>
<td>-20.38</td>
</tr>
<tr>
<td>WRF f</td>
<td>0.928</td>
<td>0.845</td>
<td>0.672</td>
<td>-4.24</td>
</tr>
<tr>
<td>E–e–l</td>
<td>1.270</td>
<td>0.878</td>
<td>0.724</td>
<td>3.23</td>
</tr>
<tr>
<td>E–l</td>
<td>1.420</td>
<td>0.839</td>
<td>0.721</td>
<td>1.18</td>
</tr>
<tr>
<td>83 m agl</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>WRF c</td>
<td>1.536</td>
<td>0.718</td>
<td>0.693</td>
<td>-10.31</td>
</tr>
<tr>
<td>WRF f</td>
<td>0.962</td>
<td>0.925</td>
<td>0.688</td>
<td>3.72</td>
</tr>
<tr>
<td>E–e–l</td>
<td>1.368</td>
<td>0.869</td>
<td>0.727</td>
<td>2.90</td>
</tr>
<tr>
<td>E–l</td>
<td>1.469</td>
<td>0.853</td>
<td>0.724</td>
<td>2.42</td>
</tr>
</tbody>
</table>

Coefficients $b$ and $m$ belong to the correlation line $y(x) = mx + b$. In bold, the best result for each parameter.

$V_h$ = horizontal velocity; $W$ = vertical component of the velocity vector; $E$ = turbulent kinetic energy; $T$ = dry temperature; $R^2$ = coefficient of determination; $\epsilon$ (%) = percentual bias error; rms = root mean square error; "WRF c" and "WRF f" = mean WRF (c)ourse and (f)ine mesh simulations, respectively.
period. In what concerns turbulent kinetic energy, $E$, conclusions are harder to draw, with improvements in $m$ and $R^2$ brought about by mesh $h+$, but with the other parameters profiting from a reduction in vertical resolution, particularly at 43 m agl. Finally, results for temperature $T$ varied little between the various meshes, considering that $\epsilon$ and $rms$ are affected by the previously mentioned offset.

In summary, it can be said that these mild grid refinements ($h+$ and $v+$) produced an observable global improvement in the numerical results. A systematic exercise of grid refinement was not pursued because of the increased computational effort and also because of the impossibility to go beyond the errors that always occur in turbulent flow modelling, of which discretization errors are only one component. We believe that the mesh used in the remaining of this section (mesh $s$ of Table IV) can be considered a ‘useful accurate’ mesh, to borrow Eidsvik’s terminology. It would nonetheless be ideal to use even higher resolutions to capture additional microscale effects.

4.1.2. Comparison between mesoscale and microscale results.

Performance indicators for the simulations covering the complete campaign period are presented in Table VI. Results from the two WRF-ARW simulations described previously are also included (‘WRF c’ is the coarse 3 km mesh simulation, and ‘WRF f’ is the fine 1/3 km mesh simulation) and used as guidance for the performance of the microscale code. Because the $E-l$ and $E-\varepsilon-l$ models produced very similar results, the figures presented next show only results from the $E-\varepsilon-l$ model. Figure 6 show wind roses obtained at 83 m agl, and Figures 7–10 show times series of selected quantities.

By analysing the observations wind rose in Figure 6, the existence of two main sources of wind, north-northwest and eastern winds, can be seen. In this regard, the simulation results showed a more uniform spread of the northwest winds, whilst observations show a larger concentration on the 300–330° sectors. The converse occurred for the eastern winds, now with the observations showing a more uniformly wind distribution inside 60–120° sectors contrary to the more concentrated

![Figure 6. Wind roses at 83 m agl for the complete campaign period.](image)

![Figure 7. Time series of horizontal velocity ($V_h$) at 83 m agl for the complete campaign period.](image)
60–90° winds predicted by the simulations. Nevertheless, it can be said that the mesoscale and microscale wind roses are very similar, both following qualitatively well the observations pattern.

The performance indicators for the horizontal wind velocity are presented in Table VI. This quantity is perhaps the most important for wind energy applications, since its cube is directly related to the power of the flow available to the wind turbines. Times series of $V_h$ at 83 m agl are presented in Figure 7. By analysing the WRF results of Table VI for both heights, it can be seen that a global improvement resulted from the mesh refinement (9 × refinement)—the ‘WRF c’ results underestimated $V_h$ for both heights, as suggested by the time series in Figure 7. Nevertheless, not all quantities improved, with values of $R^2$ at both heights and the $rms$ error at 83 m agl showing the opposite behaviour. Both microscale code results ($E_{-e-l}, E_{-l}$) were produced using boundary conditions extracted from the ‘WRF c’ results, and they produced an obvious global improvement, as can be seen in Table VI for both heights. The microscale code results are more similar to those obtained using the ‘WRF f’ mesh, as can be also seen in Figure 7, due to the use of similar horizontal grid
resolutions—in this respect, the ‘WRF f’ results can be thought of as target results to attain using the microscale code coupled with the ‘WRF c’ results. In conclusion, it can be said that the microscale results compared well with those of the ‘WRF f’ simulation, producing performance indicators of equivalent magnitude.

The vertical component of the wind vector (W) is associated with non-desirable vertical forces over the wind turbine mechanical components, and its correct prediction is important in wind turbine siting tasks. The performance indicators for W are presented in Table VI; and time series, obtained at 45 m agl for some of the results, is presented in Figure 8. The observed trend was relatively well captured by ‘WRF f’ and the microscale results, whilst ‘WRF c’ filters off the oscillations of the vertical component. Nevertheless, this quantity presented very high $\epsilon$ errors (bias), due to the small mean values of W from observations ($W_{obs} \approx 0$). In this respect, ‘WRF c’ actually showed a lower $\epsilon$ error than ‘WRF f’, even whilst predicting an almost constant and very low vertical velocity, as can be seen in Figure 8—a direct consequence of reduced horizontal mesh resolution. The lower $\epsilon$ and rms errors were produced by the microscale code results, and the higher $m$ values were produced by the ‘WRF f’ results (Table VI). The $E_{\epsilon\ell}$ microscale code presented the best $R^2$ at 45 m agl, whilst ‘WRF c’ yielded the highest $R^2$ at 83 m agl. From Figure 8, it can be seen that the ‘WRF f’ results followed better the negative vertical velocities measured by the sonic anemometers, whilst the microscale code was more effective in the prediction of the positive vertical velocities.

Information on turbulent kinetic energy $E$ is presented in Table VI and in Figure 9, which show time series for 45 m agl. By analysing the time series, a typical diurnal-cycle structure emerges. However, a detailed observation of the time series also shows the presence of some episodes of large amounts of nocturnal turbulence (see, e.g. period between days 105 and 110). Furthermore, it can also be seen that the first half of the campaign period has, on average, lower turbulence levels than the second, probably a result of the higher stability of the flow during the colder months. By analysing the performance indicators presented in Table VI, it can be seen that the higher values of $m$ and $R^2$ (better predictions) were obtained for all models at 45 m agl. Nevertheless, the $\epsilon$ and rms errors did not follow this trend, with the ‘WRF f’ and $E_{\epsilon\ell}$ showing lower errors at 83 m agl. A detailed analysis of the time series at 45 m agl showed a tendency for the $E_{\epsilon\ell}$, $E_{\ell\ell}$ and the ‘WRF f’ simulations to over-predict the nocturnal high intensity turbulent episodes, which contributed to the over-estimation of $\epsilon$ found for those simulations (see days 105–100). The decay of turbulence intensity with height was over-predicted by all simulations, as can be inferred by the $\epsilon$ trend between 45 and 83 m agl for all the cases (Table VI). The low $R^2$ values obtained here are a manifestation of the difficulty in the prediction of turbulence, in particular the short peaks of turbulence observed. Globally, the best $E$ results were obtained using the $E_{\epsilon\ell}$ model.

Observing Figure 10, where time series of $T$ is presented, a marked diurnal cycle (short period variations) superimposed on longer (≈15–20 days) temperature variations is visible . It was during the first days of campaign and during the first hours of day 97 (7 April) that the lower temperatures were observed, around $-3^\circ$C. The higher measured temperatures, around 18$^\circ$C, occurred between days 68 and 75 (9–16 March). From the correlation perspective ($m$ and $R^2$ values), temperature was the better predicted quantity, with both indicators generally above 0.8 for all codes. As explained previously, the $\epsilon$ and rms errors are plagued with offset errors due to differences in the altitudes. A reflection of this problem was the production of a systematic lower temperature for the ‘WRF f’ results in relation to ‘WRF c’, and this comes from the existence of very similar potential temperature predictions (not shown) and a higher topography at the measuring point, due the use of higher horizontal spatial resolution (note that to first-order approximation, $T = \theta - gz/c_p$). The better results were obtained using the microscale simulations, where the higher $m$ and $R^2$ values were obtained.

Figure 11 shows vertical profiles of the horizontal velocity ($V_h$), potential temperature ($\theta$) and turbulent kinetic energy ($E$) at selected times—two 00:00 LT instances (upper panels) and two 12:00 LT instances (lower panels) on 9 April (on the left) and 13 April (on the right). The main objective here was to determine if the microscale results follow the basic structure of the vertical profiles predicted by the WRF mesoscale code (fine mesh), for either regime. Sonic results of $V_h$ and $E$ were also included in the figures to help the analysis. Due to the very good agreement between the potential temperature profiles from the two codes, the sonic results for this quantity were discarded.

In the two upper plots in Figure 11, both at midnight, when the atmosphere is typically stably stratified, both codes showed very small turbulence levels and positive potential temperature vertical gradients. In case of the upper left plot, the minimum $E = 0.1$ m$^2$ s$^{-2}$ used in the WRF simulations was obtained for all altitudes, indicating an almost non-turbulent atmospheric flow at this location. Both velocity profiles in the upper left panel can be thought of as the combination of an almost linear high profile (resembling a Couette-type flow driven by the geostrophic velocity) and a lower-level local velocity maxima, the latter due to a speed-up effect that is a typical feature at hill top locations (see, e.g. Jackson and Hunt$^{43}$). The trend seen in the observations suggested a lower-level velocity maximum at a height of $\leq$ 45 m agl (height of the lower sonic), a prediction more closely followed by the microscale code. The flow at midnight on 13 April (top right of Figure 11) presented a thicker layer, inferred from the $E$ vertical profiles, and the velocity profiles showed something similar to a low-level jet, with the velocity maximum occurring at approximately 80–100 m, followed by a slow and oscillating convergence to a barotropic geostrophic wind. Both codes overestimated the sonic observations of $V_h$.

In the bottom half of Figure 11, two 12:00 LT occasions are shown, when the atmosphere is typically unstably stratified, corresponding to a convective boundary layer. In agreement with the behaviour found in the nocturnal instances, the flow on the bottom left presented a shallower layer than the one on the right, as can be inferred from the vertical profiles...
of $E$ and $\theta$. In both occasions, the $\theta$ profiles presented an unstable layer adjacent to the ground, followed by a nearly uniform distribution with height, i.e. a well-mixed layer, and then capped by a temperature inversion. By analysing the $E$ profiles, it can be seen that the higher vertical resolution used in the microscale code enabled the prediction of a maximum just above the ground, something that is corroborated by the sonic results on the right panel. In terms of $E$ magnitude, the two codes predicted different profiles, with the WRF results better approaching the sonic results on 9 April, whilst WINDIE yielded more accurate results on 13 April. The velocity showed, for the two occasions and the two codes, the presence of a surface region where the velocity rapidly increases with height, followed by a layer of nearly constant speed, i.e. a well-mixed layer, which extends almost until the top of the boundary layer, followed then by a transition layer where the velocity adjusts to the geostrophic wind. The sonic results for $V_h$ were always higher than predicted by the two codes, which, nevertheless, captured well the trend registered in the observations, with the microscale model achieving better agreement.

Globally, it can be seen that the mesoscale and microscale codes showed a fairly good agreement in the structure of its vertical profiles of $V_h$, $E$ and $\theta$, despite some magnitude differences inherent to the different numerical formulations. Of particular relevance was the verification that the turbulent kinetic energy predicted by the microscale code showed the desired decrease in magnitude with height, a trend that the standard $\varepsilon$ model, used by this team in previous works, was not capable of reproducing.

4.2. Partial conclusion

From the results presented in the previous section, it can be concluded that the proposed coupled strategy produced good results in the simulation of a real atmospheric flow over complex terrain. This is illustrated by the positive comparisons between model results and sonic observations. Improvements could be observed in many performance indicators (for $V_h$, $W$, $E$ and $T$) brought about by the coupling procedure between the coarse mesoscale simulation and WINDIE, even if these were not wholly uniform across the full scope of quantities, periods simulated or heights above ground level. Comparisons between fine grid WRF-ARW and WINDIE simulations of similar resolution had comparable performance, and there was good agreement between the vertical profiles of both codes. However, simulation wall clock time for running a ‘WRF c’ + WINDIE case was less than 50% of that needed to run a ‘WRF f’ case, when using the same number of processors.
5. CONCLUSION AND FUTURE WORK

The microscale code WINDIE, initially developed for the simulation of neutrally stratified atmospheric flows over complex topography, was here extended and validated for the simulation of non-neutral atmospheric flows with Coriolis forces. Turbulence was modelled using the $E-l$ and $E-\epsilon-l$ models, which have proved to be more adequate than the standard $E-\epsilon$ model for the simulation of atmospheric flows, due to their improved estimation of the turbulent length scales.

The validation tasks included two cases from the GABLS project: a stably stratified boundary layer and a diurnal cycle, and the code performance was demonstrated to be equivalent to that presented in the benchmark studies by codes with similar characteristics, i.e. RANS codes with one-equation or two-equation turbulence models, sharing their advantages and also their shortcomings.

To test the applicability of the new code to real situations, a series of simulations were performed for a time-varying atmospheric flow (a nearly continuous 3-month period starting in February 2012) over a moderately complex topography in the Portuguese mainland. The proposed one-way coupling procedure used the WRF-ARW mesoscale code (three-level two-way nested simulation with finer horizontal grid of $\Delta(x, y) = 3 \text{ km}$) to produce time-varying boundary conditions for the WINDIE code and comparisons with sonic anemometer measurements and with WRF-ARW results that showed that the code can adequately simulate real atmospheric flows over complex topography. Evaluation of model performance was carried out using linear regression parameters (intersect $b$ and slope $m$ of the regression line and coefficient of correlation $R^2$), and common statistical indicators (the $rms$ and bias $\epsilon$).

These results showed the ability of the microscale code to produce similar and often better performance indicators than those obtained with the five-level nested WRF-ARW results, which used a horizontal grid resolution similar to that used in WINDIE. Of particular relevance was the verification that the turbulent kinetic energy predicted by the microscale code that showed the expected decrease in magnitude with height, a trend that the standard $E-\epsilon$ model for the simulation of atmospheric flows, due to their improved estimation of the turbulent length scales.

The coupled procedure developed here could be used, e.g. to produce high spatial resolution time series for assimilation by standard codes on the wind industry, e.g. WAsP, and can also be used in the short-term wind prediction for wind farms, specially on the earlier times of operation when the lack of historical data prevents a good calibration of neural network prediction tools. On the other hand, it should be noted that the majority of CFD codes do not model a variety of atmospheric physics (e.g. radiation, cloud and land surface) present in mesoscale models that, in some situations, may be more important than the benefits that CFD can bring.

The research team is currently integrating an updated version of the canopy model of Lopes da Costa and on the development of a modified version of the coupled approach, where the microscale boundary conditions are constructed from directionally filtered time-averaged (year long) WRF-ARW simulation results.

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