

Modeling the Flow Response to Surface Heterogeneity during a Semi-Idealized Diurnal Cycle

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ABSTRACT: To characterize the effects of subgrid surface heterogeneity, the blending-height concept has been developed as a coupling strategy for surface parameterization schemes used in numerical weather prediction models. Previous modeling studies have tested this concept using stationary conditions with one-dimensional strips of surface roughness. Here, large-eddy simulations are used to examine the response of the blending height and effective surface roughness to tiled land-cover heterogeneity, or a two-dimensional chessboard pattern of alternating high and low vegetation given a diurnal cycle of solar irradiance in subarctic conditions. In each experiment, the length scale of the roughness elements is increased while the total domain fraction of each vegetation type is kept constant. The effective surface roughness was found to decrease with increasing length scale of surface cover heterogeneity, which is shown to have a significant impact on estimated wind turbine power calculated from logarithmic wind profiles. In stable conditions, the blending height in cases with large heterogeneity length scales was found to exist well above the surface layer. Because the behavior of the blending height has implications for coupled models, a simple model for the blending height as a function of heterogeneity length scale is introduced.

KEYWORDS: Land surface; Boundary layer; Wind; Large-eddy simulations; Diurnal effects; Renewable energy

1. Introduction

To curb global carbon emissions and meet worldwide energy needs, wind power has grown to be an important alternative to fossil fuels. According to the International Energy Agency (IEA), onshore wind power is forecast to reach 8.4%, or 2235 TW h, of the global energy generation by 2026—up from 5.6% in 2020 (IEA 2021). As developers seek locations with good wind resources far from human habitation, rural forested regions and other sites with heterogeneous land cover have been considered for wind farm sites. In the United States, a total of 1961 or 3.97% of wind turbines were located in forested regions as of 2017 (Xiarchos and Sandborn 2017). In Sweden, as 70% of the land cover is forested, onshore wind farm development is assumed to continue to grow in forested areas (Mohr et al. 2018).

As forests and other roughness elements act to decrease wind speeds close to the ground and increase turbulence levels, the modeling of the extent of these effects has been crucial in the development of technologies related to wind energy. In assessing the location and construction of wind farms, accurate simulation of turbulence properties is key to determining fatigue loads, which over forests can be considerable, due to the higher turbulence level (Arnqvist et al. 2015). Accurate modeling of momentum, heat and moisture exchange remains beneficial long after the wind farm becomes operational. Mesoscale models used to forecast wind power and variables related to wind turbine ice accretion in the hours to days ahead forecast horizons have been shown to be sensitive to the subgrid representation of forests (Mohr et al. 2018; Janzon et al. 2020).

Therefore, the accurate parameterization of land-cover effects is important for all phases of wind farm operations.

Land cover often displays variations that influence its exchange properties, including those of momentum, radiative, latent, and sensible heat. These heterogeneous surface characteristics have been shown to affect the formation of convective clouds (Patton et al. 2005; Huang and Margulis 2013; Lee et al. 2019), momentum fluxes (Bou-Zeid et al. 2004; Verkaik and Holtslag 2007; Nakao and Hattori 2019), and scalar fluxes (Calaf et al. 2014). Much effort has been made in modeling the impact of the patchiness—or heterogeneity—of forests and other roughness elements over a region. Changes in surface roughness cause vertical wind perturbations that affect momentum transport. This influences the wind shear in the vicinity of the roughness changes (Claussen 1987). The response to the changes in the surface cover diffuses vertically and downstream through advection in a growing layer known as the internal boundary layer (IBL) (Kaimal and Finnigan 1994). The properties within the volume of the IBL are related to the response of the atmosphere to the surface cover and the vertical extent of the IBL is defined by the vertical and downwind diffusion of these properties. In a setting with many horizontal roughness changes, the highest elevation where there are still noticeable differences between the properties of the different IBLs is referred to as the blending height h_b (Claussen 1995). This is the height at which the vertical fluxes of momentum, heat, and scalars become horizontally homogeneous; that is, no longer influenced by the individual patches of differing land cover below.

Although the profiles of wind, temperature, and scalars upwind and downwind of the roughness change have been shown to be affected by the local land surface below h_b , above h_b the profiles are horizontally homogeneous. Provided that the blending height is lower than the surface-layer height, one can

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approximate regional turbulent fluxes above h_b using Monin–Obukhov similarity theory (MOST) (Monin and Obukhov 1954). Here, an a priori effective roughness length $z_{\text{eff},m}$ is determined from mean vertical wind profiles such that the roughness length homogenizes the region with otherwise heterogeneous roughness (Bou-Zeid et al. 2004, 2007). This is a powerful result for the coupling of surface-layer parameterizations within models often used in wind energy applications: if one can estimate an effective roughness length and an appropriate value of h_b over the grid cell, then existing theory can be used to estimate the height at which the conditions within the model grid cell are horizontally homogeneous. If the blending height is higher than the first model level, there is a need to have a surface-layer scheme that computes fluxes over individual types of land cover within the grid cell and aggregate the effective fluxes appropriately (Bou-Zeid et al. 2020).

Numerous modeling studies have been conducted (Claussen 1987; Bou-Zeid et al. 2007; Miller and Stoll 2013) in order to determine formulations of h_b due to roughness changes and develop parameterization schemes around it. Many of these studies use idealized conditions: Bou-Zeid et al. (2004, 2007) used large-eddy simulation (LES) in neutral conditions to develop a model for determining the effective roughness and blending height based on the heterogeneity length scale; Miller and Stoll (2013) extended the work of Bou-Zeid et al. (2004, 2007) to stable conditions. While the aforementioned studies focused on the impacts of one-dimensional strips of differing heterogeneity length scales, Nakao and Hattori (2019) investigated the impacts of two-dimensional patterns of idealized forest on the bulk drag coefficient in neutral conditions. Englberger and Dörnbrack (2018) examined the wake characteristics of wind turbines over an idealized heterogeneous surface in comparison with that of a homogeneous one given a diurnal cycle, but they did not vary the heterogeneity length scale. In their review, Bou-Zeid et al. (2020) emphasize that studies using models coupled with full land surface models and portraying the effect of flow aligned in an oblique manner to the surface transitions will be important for the next steps in modeling surface heterogeneity. In addition, Bou-Zeid et al. (2020) mention a continued need for the study of land–atmosphere interactions over unstructured, or more realistic, varied landscapes.

While the impacts of heterogeneous surfaces have been investigated previously, we are not aware of studies that have investigated the impact of the length scale of tiled roughness heterogeneity within a subarctic environment influenced by a semi-idealized diurnal cycle of insolation; this includes evolving stratification and effects such as an inertial oscillation with associated variations in surface wind direction. In short, how does the effective roughness and blending height respond to an evolving diurnal cycle? Also, how does the effective roughness change as a function of heterogeneity length scale under these conditions and how does this influence wind speed at the height of a wind turbine? Understanding the behavior of the blending height and effective roughness length will be beneficial for coarse-resolution models that simulate diurnally evolving atmospheres over complex regions of land cover with wide ranges of heterogeneity length scales as long as these effects are analyzed above the blending height.

This study will use LES combined with a land–vegetation model to examine the impacts of tiled land-cover heterogeneity in an environment including a semi-idealized diurnal cycle of solar irradiance in dry conditions with a series of two-dimensional patterns varying in roughness over an otherwise flat surface. Section 2 will describe relevant theory. Section 3 will introduce the LES model used as well as the experimental setup. The results of the study concerning the changes in blending height and effective roughness length are presented in section 4, in addition to the effect of surface roughness on the inertial oscillation and characteristics of spatial spectral densities. Section 5 will then discuss the aforementioned results in the light of previous research. Section 6 will state the conclusions.

2. Theory

MOST remains the cornerstone of atmospheric surface-layer theory; the surface layer being the lowest turbulent layer of the atmosphere where the vertical derivative of the fluxes is negligible (Stull 1988). While MOST assumes a flat, homogeneous surface, numerous studies (Wood and Mason 1991; Bou-Zeid et al. 2004, 2007; Miller and Stoll 2013) extend MOST to homogenize heterogeneous surfaces of varying roughness by determining an effective aerodynamic roughness length due to momentum $z_{\text{eff},m}$ and heat $z_{\text{eff},h}$. Specifically, the effective roughness lengths can be estimated from the log-law relation

$$\kappa \frac{\langle U(z) \rangle}{\langle u_* \rangle} = \ln \left(\frac{z}{z_{\text{eff},m}} \right) - \left[\Psi_m \left(\frac{z}{L} \right) - \Psi_m \left(\frac{z_{\text{eff},m}}{L} \right) \right] \quad \text{and} \quad (1)$$

$$\kappa \frac{\langle \theta(z) \rangle - \langle \theta_0 \rangle}{\langle \theta_* \rangle} = \ln \left(\frac{z}{z_{\text{eff},h}} \right) - \left[\Psi_h \left(\frac{z}{L} \right) - \Psi_h \left(\frac{z_{\text{eff},h}}{L} \right) \right], \quad (2)$$

where the wind speed $U = \sqrt{u^2 + v^2}$ (u and v are the magnitude of the horizontal wind components) and potential temperature θ are domain averaged at a given time—from now on, angle brackets will indicate area averages and overbar denotes time averages. The friction velocity u_* is calculated by the land surface model (LSM), θ_0 is the surface potential temperature, L is the Obukhov length,

$$L = \frac{\langle \theta \rangle \langle u_* \rangle^3}{\kappa g \langle \theta_* \rangle}, \quad (3)$$

$\bar{\theta}$ is the mean reference potential temperature, $\kappa = 0.41$ is the von Kármán constant, g is the acceleration due to gravity, and θ_* is a scale temperature taken as $-(H_s)/\langle u_* \rangle$, where H_s is the sensible heat flux calculated by the LSM.

The integrated form of the stability corrections for momentum and heat are denoted by Ψ_m and Ψ_h , respectively. For stable conditions Högström (1996) used the forms

$$\Psi_m = -b_s \zeta, \quad \zeta \geq 0, \quad \text{and} \quad (4)$$

$$\Psi_h = -b_t \zeta, \quad \zeta \geq 0, \quad (5)$$

where b_s and b_t are empirical constants shown by Högström (1996) to be 5.3 and 8, respectively. For unstable conditions,

TABLE 1. Stability conditions used in this analysis and number N of diagnostic time steps realizing these domain-mean values. Absolutely stable and unstable conditions are not listed because they were not present in the simulations.

Stability condition	ζ	$N, L_f = 20 \text{ m}$	$N, L_f = 120 \text{ m}$	$N, L_f = 300 \text{ m}$	$N, L_f = 1500 \text{ m}$	$N, L_f = 3000 \text{ m}$
Weakly unstable	$-0.2 \leq \zeta < 0$	2315	2324	2325	2330	2330
Near neutral	$0 \leq \zeta < 0.02$	192	168	164	161	160
Weakly stable	$0.02 \leq \zeta < 0.6$	5750	5765	5768	5766	5767

we use a form of Ψ_m and Ψ_h found by Wilson (2001) to fit well with previous data:

$$\Psi_m = a \ln\left(\frac{1 + \phi_m^{-1}}{2}\right) \quad \text{and} \quad (6)$$

$$\Psi_h = a \ln\left(\frac{1 + \phi_h^{-1}}{2}\right), \quad (7)$$

where the nondimensional gradient for wind and temperature takes the forms

$$\phi_m = (1 + \gamma_m |\zeta|^{2/a})^{-1/2} \quad \text{and}$$

$$\phi_h = (1 + \gamma_h |\zeta|^{2/a})^{-1/2},$$

respectively. Wilson (2001) used $a = 3$ and constants $\gamma_m = 3.6$ and $\gamma_h = 7.9$.

Next, we introduce the height of the atmospheric boundary layer z_i , which in our analysis is defined as the height at which the magnitude of the horizontal shear stress $\tau = \sqrt{u'w'^2 + v'w'^2}$, where the prime denotes a deviation from the time-averaged value, reached 5% of its surface value; this is the same definition used in Miller and Stoll (2013).

The term $\zeta = z/\langle L \rangle$ is used as a stability parameter. Table 1 lists the stability criterion used in the analysis, where the reference height $z = 12 \text{ m}$ is approximately the mean height of the surface layer in stable conditions through the experiments described later this study when using the $0.1z_i$ scaling proposed by Holtslag and Nieuwstadt (1986).

Over regions of heterogeneous surface roughness, the internal boundary layer is the response to the convergence of low-level flow at the edge of an area of increased roughness or the divergence of flow over an area of relatively low roughness. The depth of the internal boundary layer can be determined as the level at which vertical diffusion and horizontal advection balance each other (Savelyev and Taylor 2005). A common topic in the literature is the concept of a blending height h_b at which the flows become spatially homogeneous and no longer feel the effect of the heterogeneous surface below it. Many studies have defined h_b using the internal boundary layer height of wind (Bou-Zeid et al. 2004; Miller and Stoll 2013). In an LES study of a stable boundary layer, Miller and Stoll (2013) noted that the vertical extent of the internal boundary layer due to temperature is roughly 1.5 times the height of that due to momentum and subsequently defined a separate blending height due to temperature $h_{b,r}$. To combine the effective roughness and blending height into a parameterization, Bou-Zeid et al. (2004) assumed that, following a roughness

change, internal boundary layers extend roughly 2 times the horizontal length of a roughness patch, before being diffused into the boundary layer flow aloft. From that assumption, they derived an expression that is similar to that of Panofsky and Dutton (1984):

$$h_b [\ln(h_b/z_{\text{eff},m}) - 1] = 1.7\kappa L_f. \quad (8)$$

A result of the combined effect of surface–atmosphere interactions and diurnal changes in stratification are inertial oscillations. Inertial oscillations are the result of the variable effect of the stratification on the turbulent drag leading to imbalances among drag, the pressure gradient acceleration, and the Coriolis force. The start of the oscillation is typically described as when the thermal contribution to the turbulent drag seizes at nighttime and pressure gradients accelerate the flow to supergeostrophic (Stull 1988). The result of this oscillation is the nocturnal low-level jet that has been well documented (Blackadar 1957; van de Wiel et al. 2010).

3. Method

a. LES model

The atmospheric model used in this study is version 5.4.3 of Meso-NH (Lac et al. 2018). Meso-NH is a nonhydrostatic anelastic numerical modeling package that includes filtered prognostic equations for heat, humidity and momentum, allowing users to simulate atmospheric processes from large-eddy scales (10s of meters) to regional scales. In this study we used the three-dimensional isotropic turbulence scheme described in Cuxart et al. (2000). As such, this allows Meso-NH to operate in an LES mode and simulate fine-scale atmospheric interactions. The turbulent mixing length is calculated from the mesh geometry, limited by whichever is smaller of the geometric average of the sides of the grid cells or the Deardorff mixing length (Deardorff 1980)

$$L_D = \left[\frac{2e}{(g/\theta_{V\text{ref}})/(\partial\theta_V/\partial z)} \right]^{1/2}, \quad (9)$$

where θ_V is the virtual potential temperature (dry potential temperature θ in our simulations), $\theta_{V\text{ref}}$ is the reference virtual temperature, g is the acceleration due to gravity, and e is the turbulence kinetic energy (TKE). A minimum limit for e was set to be $0.0001 \text{ m}^2 \text{ s}^{-2}$, because this prevented extremely small L_D in the capping inversion at the top of the domain.

The momentum advection scheme used in the simulations is the fourth-order centered scheme (CEN4TH; Lunet et al. 2017),

while scalar variables—temperature, moisture, and TKE—are advected using the monotonic version of the piecewise parabolic method advection scheme (Lin and Rood 1996). Numerical diffusion was activated with an e -folding time of 60 s. Time splitting is handled with the Runge–Kutta centered fourth-order (RK4) temporal scheme detailed in Lunet et al. (2017). Radiation is parameterized using the ECMWF radiation code, with longwave radiation parameterized with the Rapid Radiation Transfer Model (Mlawer et al. 1997) and shortwave radiation parameterized with the Morcrette (1991) scheme.

Meso-NH is particularly useful for model parameterization development, as it is capable of high-resolution simulations, while containing parts of the model system that are common to NWP systems used operationally (physical schemes, dynamical core, etc.).

b. Land surface model

The LSM used in this study is contained within version 8.1 of the Surface Externalisée (SURFEX) externalized surface scheme (Le Moigne et al. 2009). SURFEX is the package of physical parameterizations that calculate the energy exchange between the surface and the atmosphere. When SURFEX is coupled with an atmospheric model, forcing data from the lowest model level is sent to a series of models that calculate fluxes of momentum, sensible heat, latent heat, gas, and aerosol feedbacks for each of four tiles representing a different surface type: sea, lake, nature (consisting of vegetation and soil), and urban areas (Le Moigne et al. 2009). An average of each individual turbulent flux is then calculated such that it is weighted by the fraction of each surface type and the resulting flux is sent back to the atmosphere, along with information about soil moisture, roughness, albedo, emissivity, and surface temperature. Land surface processes are parameterized within the SURFEX framework using the Interactions between Soil, Biosphere, and Atmosphere (ISBA) LSM (Boone et al. 2017). ISBA calculates the impacts on surface energy fluxes due to vegetation, soil, hydrological processes, and the presence of snow cover. In this study we used the composite surface parameterization scheme described in Noilhan and Planton (1989) that includes one soil layer and a single bulk composite layer of soil, vegetation and snow. Similar bulk surface parameterization methods to that described Noilhan and Planton (1989) are used with in other operational NWP models [e.g., COSMO/Icosahedral Nonhydrostatic (ICON; Prill et al. 2020), ECMWF-IFS (ECMWF 2021)].

Also available within SURFEX is a multiple-layer canopy scheme described by Masson and Seity (2009) that uses the surface model in combination with the turbulence scheme to explicitly calculate temperature and wind at several levels within the tree canopy. Bengtsson et al. (2017) found that the Masson and Seity (2009) scheme leads to much-to-low surface temperatures in the operational configuration of their meso-scale model. Since we were not concerned with directly looking at flows within the tree canopy—only to provide the model with a wall roughness that approximated the effects of the forest aloft—we did not employ the Masson and Seity (2009) scheme in this study.

The parameters for the land-cover types used in this study are determined from lookup tables for “ECOCLIMAP II” (Faroux et al. 2007). ECOCLIMAP II combines satellite and land maps to create a database of surface descriptions at 1-km resolution for use in mesoscale models and is used within operational configurations of SURFEX. To represent low roughness values, we used the ECOCLIMAP II cover type describing low vegetation (ECOCLIMAP II COVER 123), which includes 90% bare soil and 10% grass. For high roughness, a forested cover type was used in this study (ECOCLIMAP II COVER 11) that consists of 100% evergreen needleleaf forest with a tree height h of 15 m.

c. Experimental simulations

Each LES experiment was conducted using a $300 \times 300 \times 127$ cell domain with 20-m horizontal grid spacing. The first level of the vertical grid is 0.5 m from the surface with 6.5% stretching between model levels above that until reaching 20-m vertical grid spacing at the model top. The height of the vertical grid in total is 1608.8 m. To determine the resolution to be used in this study, a sensitivity test was conducted using 10-, 20-, and 30-m horizontal grid spacing to test the performance of the model when resolving turbulence. To do this, the model was integrated in time for 12 h with the aforementioned horizontal grid spacings using the experimental model configuration—including the initial conditions that will be described below—albeit with an unvegetated lower boundary. The domain mean resolved and parameterized subgrid TKE was stored every 10 s and time-averaged for the 12-h period from 0703 to 1903 UTC. Figure 1a shows the spatial and time-averaged ratio of resolved TKE to the total TKE (resolved + subgrid TKE) as a function of height for the resolution test. Figure 1b presents the vertical grid.

Through the aforementioned resolution study, it was determined that 80% of the turbulence was resolved above a vertical level of 50.5 m using a 20-m grid spacing, which met the resolved turbulence criteria described in Pope (2004) for the vertical elevations at which modern commercial wind turbine nacelles exist. While some of the turbulence remains unresolved close to the surface, Arnqvist et al. (2019) emphasizes that using a large domain to model the fetch upstream of a wind turbine in lieu of high resolution allows for accurate predictions of cospectra and covariances over heterogeneous forest in flat terrain. The 20-m grid spacing in combination with a periodic lateral boundary was chosen in order to maximize domain size while keeping the computational expense within the available resources. A time step of 0.5 s was used in this study to avoid numerical errors. Each simulation was run for a total of 28.6 h to allow for a full diurnal cycle plus 3 h of spinup. The radiation code was called every 60 s.

To test the sensitivity of the atmospheric variables to land-cover heterogeneity, we used for a lower boundary five varying land-cover patterns with all other variables and parameters held constant, including the total domain fraction of surface cover types. The cover types were arranged into alternating chessboard patterns of squares with increasing land-cover heterogeneity length scale L_f . Here, we define the L_f as the length of one side of each square-shaped homogeneous roughness

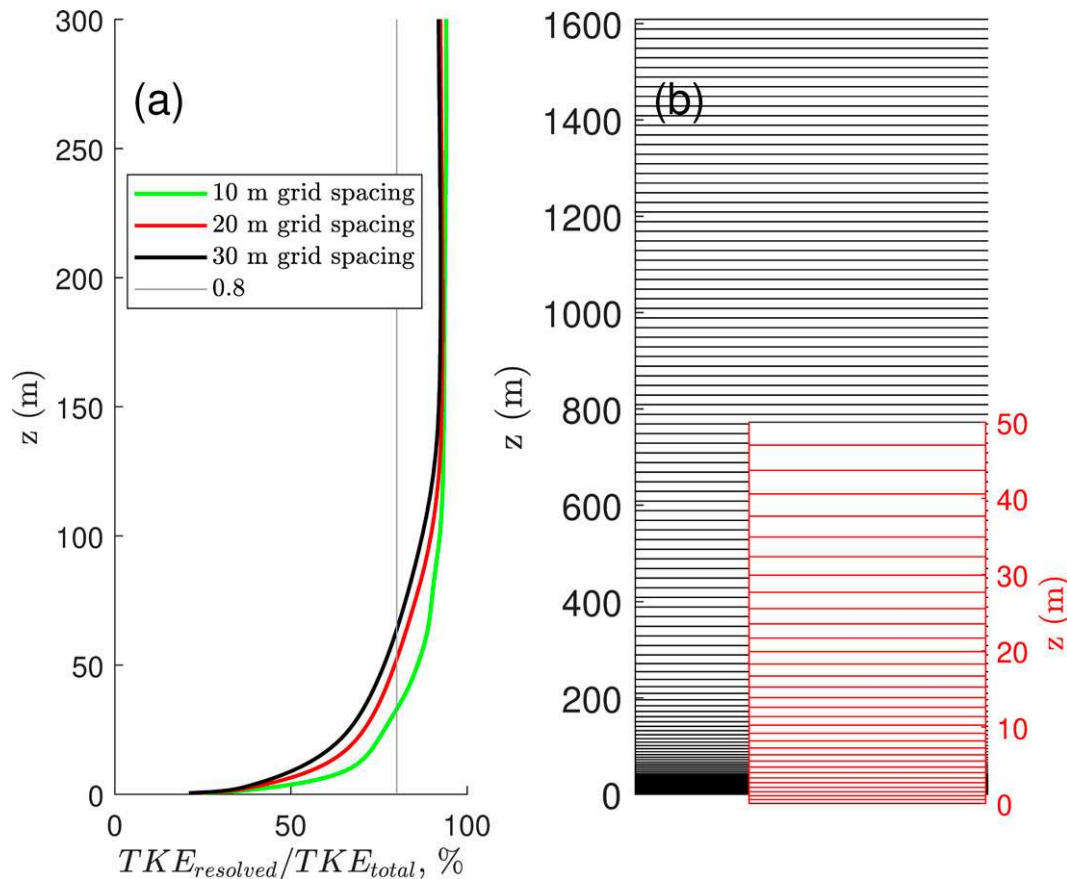


FIG. 1. (a) Horizontally and time-averaged ratio of resolved and total turbulence kinetic energy for resolution test and (b) vertical grid used in the simulations. The inset in (b) shows the detail of the vertical grid from 0 to 50 m.

patch in the chessboard. The domain land-cover patterns with $L_f = 120, 300, 1500$, and 3000 m can be seen in Fig. 2. A fifth configuration defined every other grid cell as forest or low vegetation, and this is described as $L_f = 20$ m. A network of 16 virtual towers was placed in the same location in each domain to examine local variables. The virtual towers sampled the atmosphere from the surface to the top of the vertical domain with 60-s averaging time. The virtual tower network was chosen to sample several points at different locations in each roughness patch across the domain at relatively high temporal frequency while maintaining a small dataset relative to the full gridded model output. Besides the virtual towers, 30-min-average domain-mean LES diagnostics and horizontal averages of atmospheric profiles were made and stored at 10-s intervals.

Each simulation was initialized from 0703 UTC with a neutral potential temperature profile and a wind profile shown in Fig. 3. Several initial wind profiles were tested, including monodirectional logarithmic wind profiles with differing magnitudes of large-scale forcing from the geostrophic wind u_g (the results of all of these tests are not shown). The Coriolis force was activated in order to allow for realistic diurnal wind speed changes; in order to avoid large inertial oscillations in the surface layer, an Ekman spiral profile was chosen to obtain an initial condition relatively close to atmospheric equilibrium.

This was inspired by the inertial oscillation model in van de Wiel et al. (2010), which describes an oscillation around an equilibrium profile that includes turbulent drag effects. van de Wiel et al. (2010) suggested the Ekman spiral as an idealized equilibrium profile in their analysis as it includes wind directional changes due to turbulent drag. Tests showed the morning transition through near-neutral conditions occurring around 0703 UTC, so this was used as a starting time. With the Ekman profile and the 0703 UTC starting time, the low-level inertial oscillations were considerably dampened relative to using a monodirectional initial wind profile (see Fig. 4). It is noted that the inertial oscillation remained present above 100 m. This study will focus on the results when the simulation is initialized with an Ekman profile and $u_g = 8 \text{ m s}^{-1}$, but the results were qualitatively similar with all initial conditions tested. The potential temperature profile was initialized with random $\pm 0.5\text{-K}$ perturbations over 10×10 grid cells up to an elevation of 398 m to excite turbulent eddies (Muñoz-Esparza et al. 2014).

The simulation was forced with a diurnal cycle of insolation typical of a location in northern Sweden (64.6°N) on 18 October. This implies slightly longer night than day, facilitating the evolution and maturity of a stable boundary layer while still allowing for a diurnal cycle with distinctly different daytime and nighttime

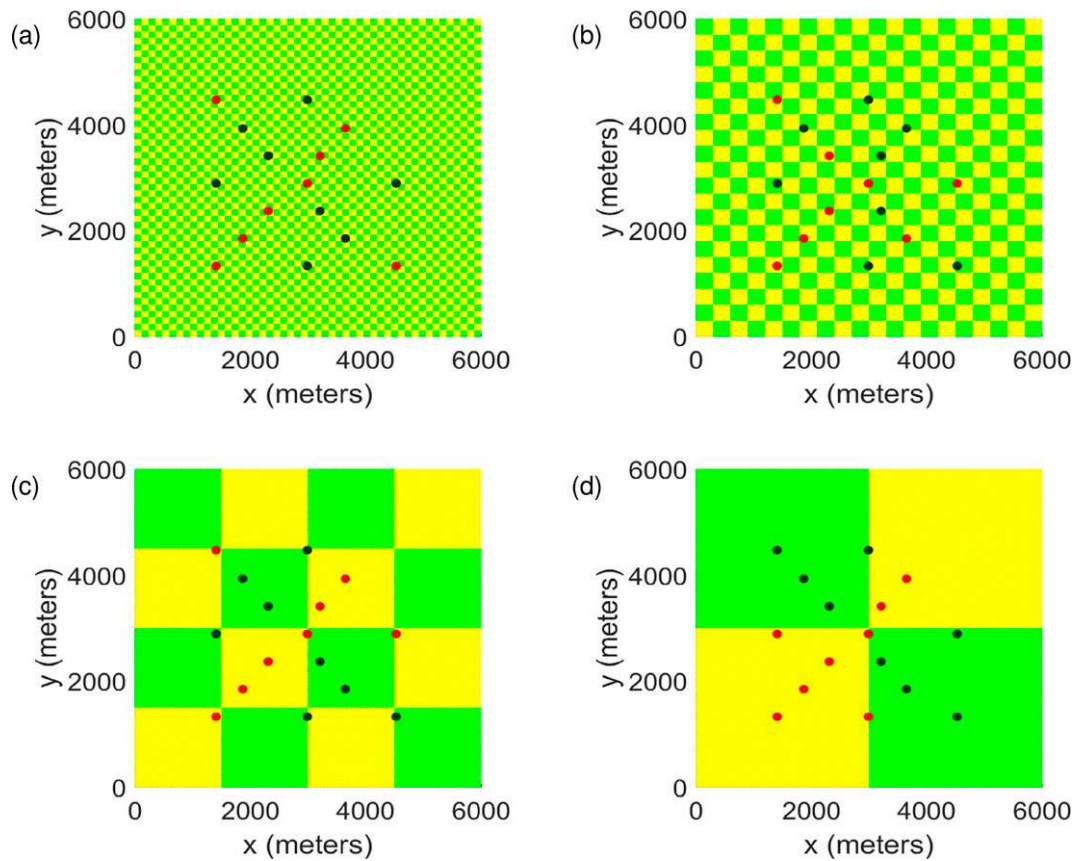


FIG. 2. Maps of the surface cover patterns for (a) $L_f = 120$ m, (b) $L_f = 300$ m, (c) $L_f = 1500$ m, and (d) $L_f = 3000$ m. Green areas indicate land cover with forest canopy, and yellow areas indicate areas with low vegetation. The dots denote the locations of virtual towers over low-vegetation (red dots) and high-vegetation (black) cover.

stratification. To illustrate the forcing used, Fig. 4 shows the evolution of the radiation budget throughout the simulation.

Table 1 shows the number N of diagnostic time steps at each stability criterion. There were no very stable values in the simulations. A brief unstable period occurred for each simulation during the spinup period, but this is not included

in the analysis. The only near-neutral periods occurred during the spinup and briefly during the sunrise/sunset transition. The largest number of diagnostic time steps were recorded during weakly stable and weakly unstable periods. From now on, the weakly unstable and stable periods will be simply referred to as unstable and stable periods.

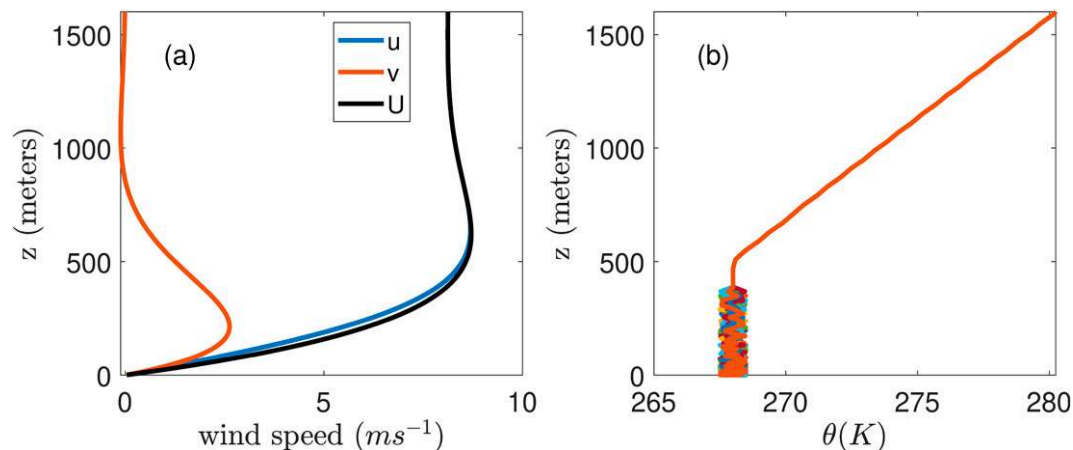


FIG. 3. (a) Initial u -wind component, v -wind component, and modulus of wind U and (b) θ profiles; the θ profile is taken from virtual towers to emphasize the extent of the initial perturbations.

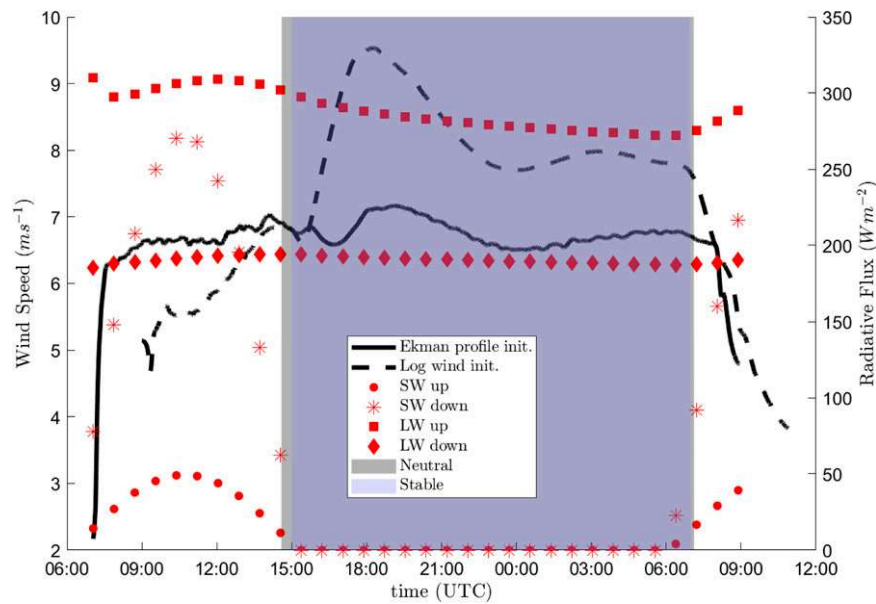


FIG. 4. Surface radiative budget for the case of $L_f = 20$ m. Gray shading shows the transition through near-neutral conditions, and lavender shading is for stable conditions. Black lines show the time series of horizontally averaged 56.3-m wind initialized with a monodirectional log profile (dashed) and Ekman profile (solid).

Table 2 shows the list of experiments in this study and a summary of relevant variables time-averaged for each stability class. The boundary layer height z_i was determined using the height at which the horizontal shear stress reaches 5% of the surface level, as stated in the theory section. The values of z_i during the stable period varied between 90 and 150 m for all experiments, with the largest mean value of 150 m for the case in which $L_f = 20$ m. The mean z_i values for the height of the boundary layer were very low for stable conditions, but $z_i \sim 100$ m are not unusual, especially for subarctic conditions (Steenefeld et al. 2007; Sinclair et al. 2022). Given the dry, cloudless boundary layer, the shallow values for z_i were not seen as unusual.

4. Results

a. Mean wind and temperature profiles

Figure 5 shows time-averaged profiles for the wind speed $\langle U \rangle$ for unstable, near-neutral, and stable conditions for each experiment. The strongest response to L_f can be seen in the

20-m case closest to the surface, with mean wind speeds systematically increasing for all stabilities as L_f increases. In stable conditions, a weak low-level jet was present in all simulations. The mean low-level jet strength averaged for stable conditions is seen in Fig. 6d.

Figure 6 shows profiles of $\langle \theta \rangle$ for each of the forest cover patterns as a function of stability and Fig. 6d shows the time series of $\langle \theta \rangle$ at the first model level. Figure 6d includes the spinup period, which was characterized by a brief imbalance before 1000 UTC resulting in a sharp increase to 270 K and decrease to 269 K. After the spinup period and following a brief near-surface temperature increase to nearly 271.3 K for $L_f = 20$ m to 271.5 K for $L_f = 3000$ m by 1230 UTC, mean surface temperatures decreased to 264.3 K by 0600 UTC in the nocturnal stable regime. Temperatures increase to approximately 268 K with increased insolation by 0900 UTC the next morning. The impact of the surface cover pattern on $\langle \theta \rangle$ is subtle, with a maximum difference of 0.15 K between the experiments seen during the first convective period. The

TABLE 2. Summary of variables for each of the model configurations for each experiment. The variables listed are as follows: L_f is the heterogeneity length scale, u_g is the geostrophic wind speed, u_* is the friction velocity, θ_* is the scale temperature, z_i is the boundary layer height, and ζ is the stability parameter. Time-averaged variables are denoted and the stability condition over which they are averaged is indicated.

L_f	u_g	$\langle \bar{u}_* \rangle, \zeta \geq 0$	$\langle \bar{u}_* \rangle, \zeta < 0$	$\langle \bar{\theta}_* \rangle, \zeta > 0$	$\langle \bar{\theta}_* \rangle, \zeta < 0$	$\langle \bar{z}_i \rangle, \zeta \geq 0$	$\langle \bar{z}_i \rangle, \zeta < 0$
20 m	8 m s ⁻¹	0.2199 m s ⁻¹	0.4435 m s ⁻¹	0.0433 K	-0.0967 K	150.6 m	669.5 m
120 m	8 m s ⁻¹	0.2147 m s ⁻¹	0.4352 m s ⁻¹	0.0445 K	-0.1033 K	145.8 m	689.3 m
300 m	8 m s ⁻¹	0.2138 m s ⁻¹	0.4343 m s ⁻¹	0.0442 K	-0.1047 K	140.2 m	679.2 m
1500 m	8 m s ⁻¹	0.2119 m s ⁻¹	0.4292 m s ⁻¹	0.0442 K	-0.1065 K	144.0 m	689.6 m
3000 m	8 m s ⁻¹	0.2099 m s ⁻¹	0.4258 m s ⁻¹	0.0438 K	-0.1075 K	144.0 m	692.0 m

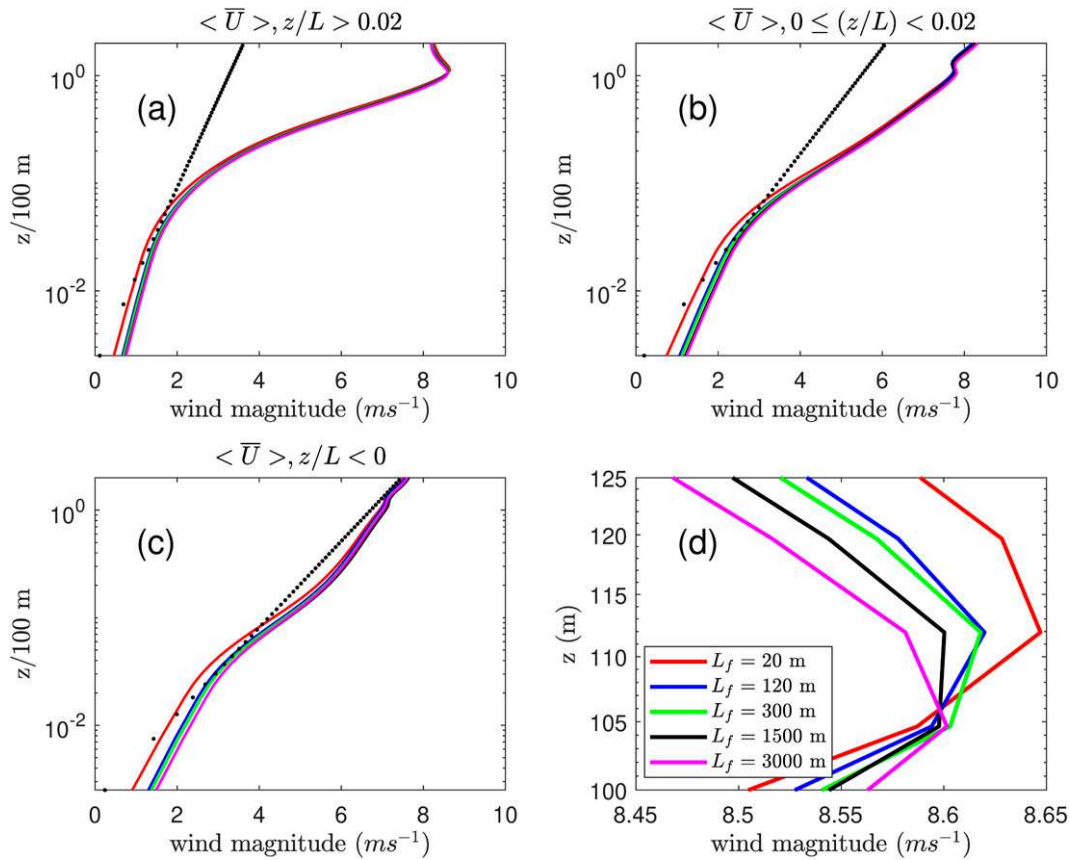


FIG. 5. Domain- and time-averaged wind speed for (a) stable, (b) near-neutral, and (c) unstable conditions. The black dotted line in (a)–(c) depicts an idealized log wind profile using the mean $\langle \bar{u}_* \rangle$ for the case of $L_f = 20$ m for each stability condition and a fixed surface roughness of 0.20 m. (d) Detail of $\langle \bar{U} \rangle$ for $\zeta > 0.02$.

domain-averaged differences in potential temperature are small during the nighttime period, which was also reflected in the mean surface sensible heat flux (not shown).

b. Effective roughness

To determine the effective roughness $z_{0\text{eff},m}$ for stable conditions, using the LES diagnostic 10 min-averaged domain mean variables, a least squares approach identical to Eqs. (8a)–(8d) in Arnqvist et al. (2015) was used by rearranging Eq. (1) to the linear expression $y = kx + m$, where

$$y = \kappa \frac{\langle \bar{U} \rangle}{\langle \bar{u}_* \rangle} - \ln(z), \quad (10a)$$

$$k = b_s / \langle \bar{L} \rangle, \quad (10b)$$

$$x = z, \text{ and} \quad (10c)$$

$$m = -b_s \frac{z_{0\text{eff},m}}{\langle \bar{L} \rangle} - \ln(z_{0\text{eff},m}). \quad (10d)$$

For each available time, the least squares fit to find k and m was performed from the surface up to 12 m. Using this approach yielded $b_s = 5 \pm 3$ during stable conditions. For

unstable conditions, $z_{0\text{eff},m}$ was determined from the wind profiles in the lowest 12 m using the stability correction proposed by Wilson (2001) and described in Eqs. (6) and (7). Here, to find $z_{0\text{eff},m}$, we solved $y = kz$, where

$$y = A^3 \exp \kappa \frac{\langle \bar{U} \rangle}{\langle \bar{u}_* \rangle} \quad \text{and} \quad (11)$$

$$A = 1 + \sqrt{1 + \gamma_m \left| \frac{z}{\langle \bar{L} \rangle} \right|^{2/3}}, \quad (12)$$

and an iterative solution is found for $z_{0\text{eff},m} = B^3/k$, where $k = B^3/\bar{u}_*$ and

$$B = 1 + \sqrt{1 + \gamma_m \left| \frac{z_{0\text{eff},m}}{\langle \bar{L} \rangle} \right|^{2/3}}. \quad (13)$$

A fixed range of z up to 12 m was chosen to be consistent with the calculations of stable conditions. The fitting procedure using z up to a dynamic height of $0.1z_i$ was also attempted with similar results, but the solution fitting the profile up to a fixed vertical level of 12 m was more stable with less noise.

Figure 7 shows the results for $z_{0\text{eff},m}$, time-averaged by stability condition and plotted as a function of L_f . The mean $z_{0\text{eff},m}$ values

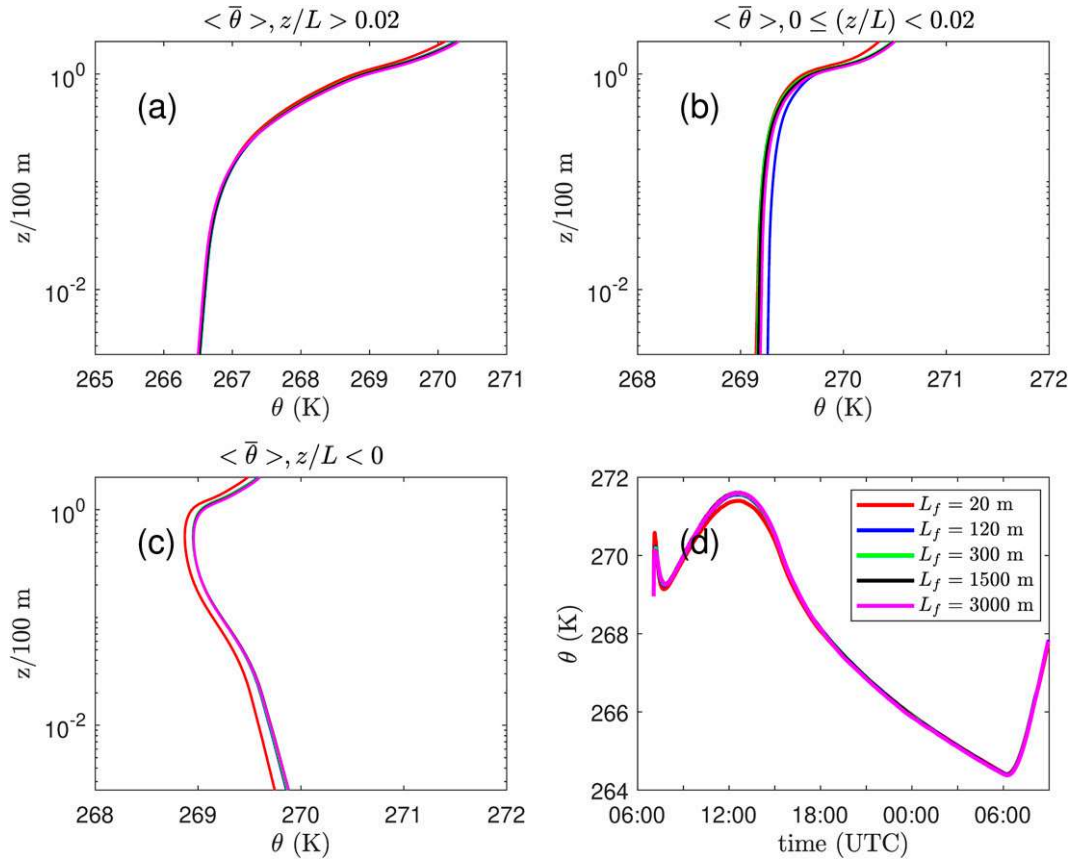


FIG. 6. Domain- and time-averaged $\bar{\theta}$ for (a) stable, (b) near-neutral, and (c) unstable conditions. (d) Time series of $\langle \bar{\theta} \rangle$ at the first model level.

for weakly stable and near-neutral conditions are similar for all L_f , but the mean $z_{0\text{eff},m}$ values for weakly unstable conditions are generally smaller than the near-neutral values by 14%–21%; the largest spread in the stability-averaged $z_{0\text{eff},m}$ values is for the

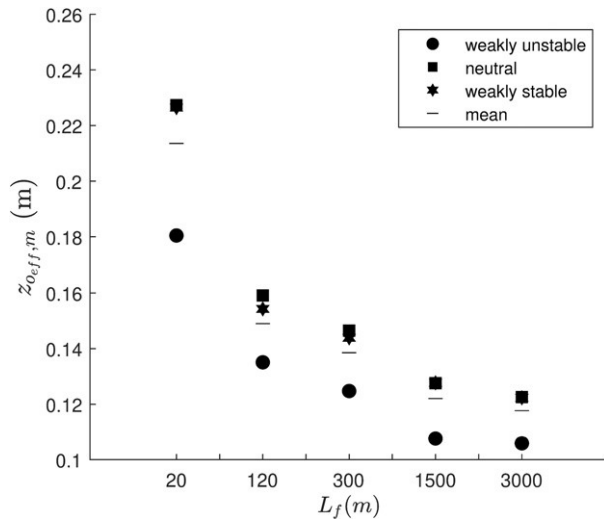


FIG. 7. The $z_{0\text{eff},m}$ time-averaged by stability condition and for each L_f .

case of $L_f = 20$ m. Despite the spread in $z_{0\text{eff},m}$, the decreasing trend as a function of L_f is the same for all stability conditions. Here we see a response following the wind speed differences seen in Fig. 5, with a 61% decrease in the mean value of $z_{0\text{eff},m}$ between the shortest and longest L_f . The largest decrease in roughness was calculated between $L_f = 20$ and 120 m, suggesting a more rapid response in roughness for shorter length scales of heterogeneity. The overall trend of decreasing $z_{0\text{eff},m}$ with increasing L_f follows the results seen in Bou-Zeid et al. (2004), reflecting that as patch size increases, the contribution from roughness changes between patches to the overall roughness decreases. The SURFEX subgrid parameterization of roughness in our domain is nearest to the case with $L_f = 20$ m, which suggests that the LSM aggregation represents a surface of evenly distributed land cover in contrast to large homogeneous patches.

c. Impact of $z_{0\text{eff},m}$ on inertial oscillation

As mentioned above, a prominent feature of all experiments mentioned here was an inertial oscillation and low-level jet above 100 m. The amplitude of the oscillation in wind speed in the residual layer appears to have a weak correlation to L_f (Fig. 5d), which is consistent with the effect of $z_{0\text{eff},m}$ on the mean near-surface flow. Using the conceptual model

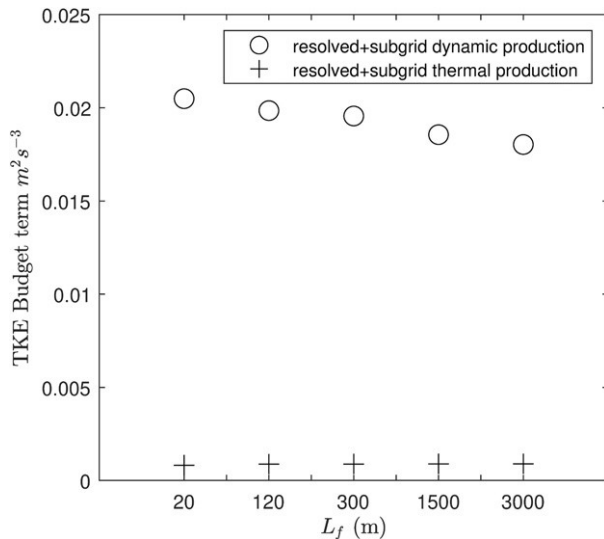


FIG. 8. Grid-averaged dynamic and thermal production of TKE, time averaged for 0703 UTC until the near-neutral period around sunset.

developed by van de Wiel et al. (2010) with an Ekman spiral solution for an equilibrium wind profile, a qualitatively similar amplitude in the inertial oscillation was found using the model output near-neutral profile from the evening transition for each L_f , suggesting that the amplitude of this oscillation is physical. To determine the source of turbulent friction in the preceding daytime period, the relative mean contribution of dynamic and thermal production to the TKE budget was analyzed (Fig. 8). Relative to the case with the $L_f = 20$ m surface cover pattern, there was a very weak correlation between the daytime thermal production of TKE and L_f . In the case described in this study, with $u_g = 8 \text{ m s}^{-1}$, secondary circulations were not present and the sensitivity of land-cover heterogeneity to the thermal production of TKE during daytime was inconclusive. In contrast, there was a stronger negative correlation between dynamic production of TKE and L_f , roughly indicative of the trend in $z_{0\text{eff},m}$. Therefore, the change in effective surface roughness due to land-cover heterogeneity appears to have a slight impact on the strength of the low-level jet.

d. Blending height

To examine the behavior of the internal boundary layer height and to determine the blending height h_b , the nondimensional vertical wind gradient anomaly $[(\partial \langle U \rangle_{y,t} / \partial z) - (\partial \langle U \rangle_{x,y,t} z / \partial z)] \langle z_i \rangle_{x,y,t} / \langle u_* \rangle_{x,y,t}$ was calculated and plotted from gridded wind speed data collected every 30 min between 1900 and 0705 UTC, following Fig. 5 in Bou-Zeid et al. (2007) (here, Fig. 9).

In Figs. 9b–9e, structures associated with the surface heterogeneity pattern clearly extend from the surface. Note that the case for $L_f = 20$ m as shown in Fig. 9a does not show clear internal boundary layer structures extending from the surface; this is to be expected, given the horizontal resolution of the LES and the

small patch size. Nonetheless, it is included here as a reference. While the plots in Fig. 9 present a qualitative picture of the blending height, it is not straightforward to convert this into an objective quantitative measure of h_b . The ambiguity of the h_b in Fig. 9 may be due to the nature of the tiled heterogeneity pattern used in this study as compared with previous studies using alternating strips of roughness normal to the streamwise wind. An attempt was made to determine h_b from virtual tower wind profiles using the technique described in Bou-Zeid et al. (2004), in which they find the elevation of a local minimum in the difference of the vertical wind gradient from average, but this was also inconclusive. In a similar attempt to determine the blending height through the local minimum of the spatial perturbation of the vertical wind gradient, Lopes et al. (2015) also had difficulties determining h_b using this technique.

To find a clear signal for the IBL height we determined at which height the difference between TKE over forested and low vegetation fell below a certain fraction of the difference in TKE at ground height (Fig. 10).

Identifying the height at which the TKE difference over forested and low vegetation areas is minimized is not trivial. At low elevations, TKE is expected to be higher over rough surfaces than over smooth. However, at higher elevations the profiles often intersected the effects of the internal boundary layer upstream, resulting in a sign change in the TKE difference. This effect was especially evident in the case in which $L_f = 120$ m. To find the absolute height at which all upstream internal boundary layers were blended the TKE profiles were averaged by land-cover type and a 10-min rolling average of the TKE difference of forested minus low vegetation profiles was stored, normalized by the surface value. From this, the blending height was defined as the lowest elevation of five consecutive model levels at which the forested to bare ground TKE difference fell below 0.001 of the surface value. In Fig. 10, we see a systematic increase in the height at which the virtual towers over rough and smooth surfaces collapse as a function of L_f . The h_b for stable conditions is plotted in Fig. 11.

For small L_f up to 300 m, h_b generally followed the trend calculated using Eq. (8) in near-neutral and stable conditions, albeit 25% higher at $L_f = 300$ m. However, for larger L_f , the h_b as calculated by Eq. (8) is much higher than the IBL height determined by the virtual tower TKE, which is limited to z_i . Given the dependence of h_b on z_i for large L_f in our results, a modification of Eq. (8) was developed. To do this, we propose a simple asymptotic relation, using an inverse summation method similar to the combined length scale model described in Gryning et al. (2007). Here, as h_b approaches and exceeds z_i , the new blending height h_{bi} will be limited by z_i . When $z_i < h_b$ in stable conditions, h_b [as in Eq. (8)] tends to be too large (relative to the LES results), and the addition $[(1/h_b) + (1/z_i)]^{-1}$ limits the new h_{bi} for the amount z_i differs from h_b . This leads to an expression for h_{bi} :

$$h_{bi} = \left(\frac{1}{h_b} + \frac{1}{z_i} \right)^{-1}, \quad (14)$$

where h_b is the blending height calculated using Eq. (8). In stable conditions for large patch sizes, we found that the

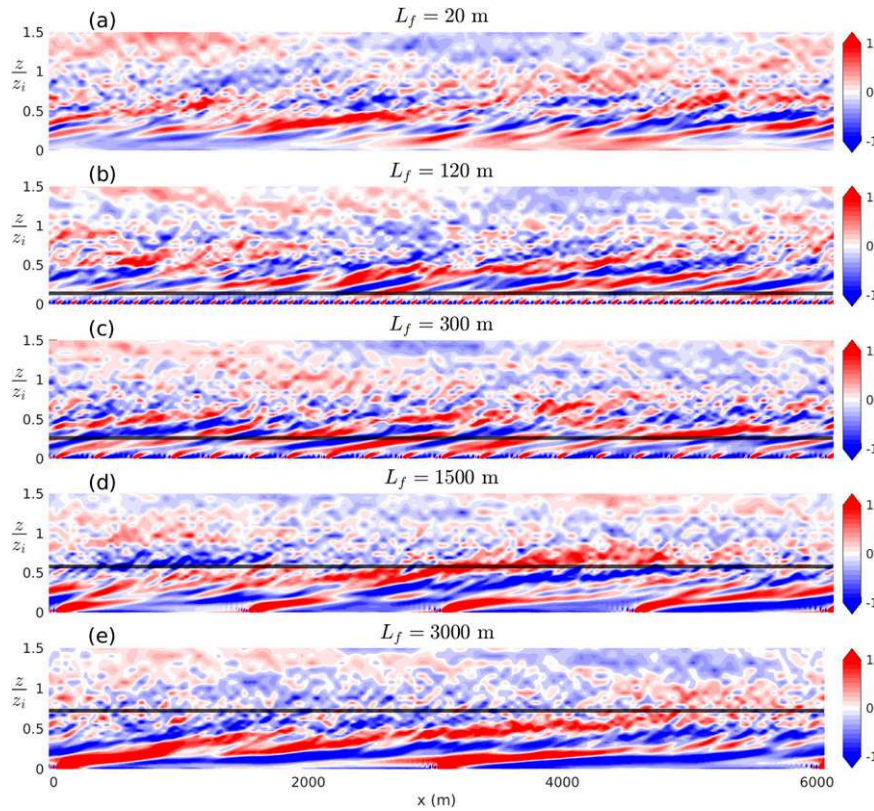


FIG. 9. Plots of $[(\partial \langle U \rangle_{y,t} / \partial z) - (\partial \langle U \rangle_{x,y,t} / \partial z)] \langle z_i \rangle_{x,y,t} / \langle u_* \rangle_{x,y,t}$ to denote internal boundary layer extent, where subscripts denote the direction of averaging, time averaged from 1900 to 0705 UTC and scaled by boundary layer height z_i . The black line is the value of the blending height calculated with Eq. (14).

method from Bou-Zeid et al. (2004) would overestimate h_b . In the case of coarse-resolution model grids containing large homogeneous patches of forest, the proposed adjustment is necessary when coupling LSMs in stable conditions. The h_b

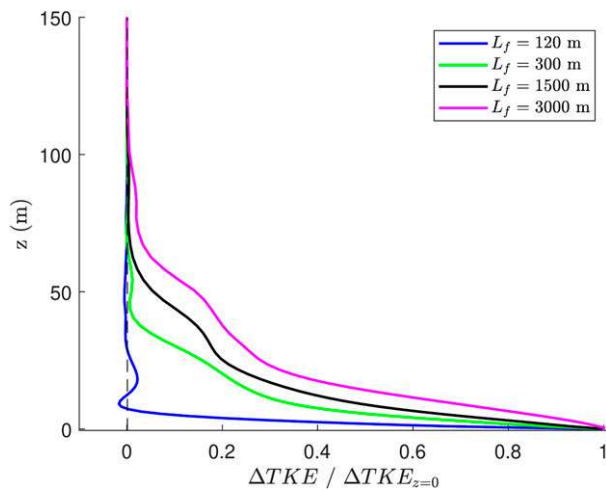


FIG. 10. Plot of $[\text{TKE}_{\text{forest}}(z) - \text{TKE}_{\text{lowveg}}(z)] / [\text{TKE}_{\text{forest}}(z=0) - \text{TKE}_{\text{lowveg}}(z=0)]$ averaged from 1900 to 0705 UTC. The dashed line denotes the 0.001 threshold.

determined using Eq. (14) is plotted on Fig. 9 and one can see that this matches very well the IBL patterns that were simulated. The blending height proposed here allows for an elegant solution for coupling LSMs to NWP systems while taking into account the response of the IBLs to shallow boundary layers.

The TKE profile method had some difficulties determining h_b for $L_f = 20$ m (not shown) since the length scale is the same as the model resolution but determined the height of h_b for larger patch sizes that matched reasonably well with qualitative analyses. Note that only stable and near-neutral conditions are shown here. In unstable conditions, the results were noisy and inconclusive and are not included in this analysis. It is also noted that, besides a very slight increasing difference with increasing L_f , the change in $\langle \bar{\theta} \rangle$ among the virtual towers (not shown) was very small, and similar analyses to determine the h_b for temperature were also inconclusive.

e. Spectra

To further study the behavior of the dispersive stress, the spatial spectra of wind and potential temperature as well as cospectra of momentum and sensible heat flux were determined in an analysis similar to that of Margairaz et al. (2020b). First, the fluctuating fields of velocity and potential temperature were averaged during the stable period, 1903–0403 UTC. Then, the coordinate system

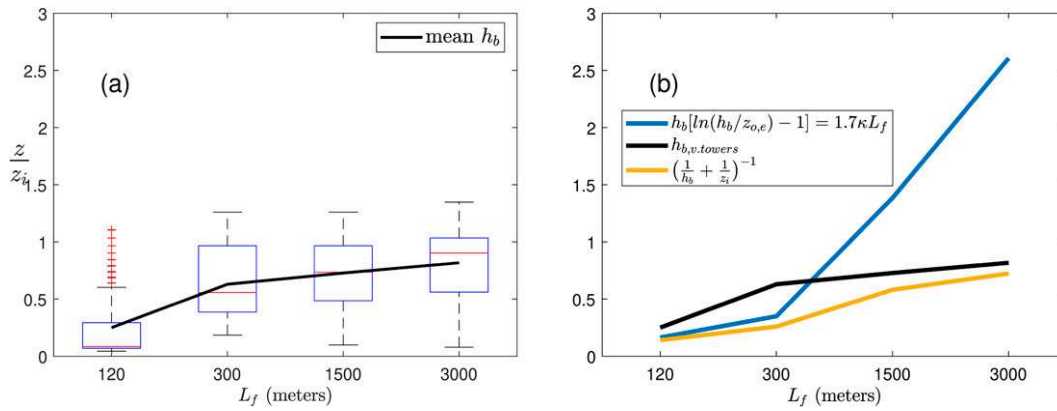


FIG. 11. (a) Boxplots of blending height calculated from virtual towers, and (b) a comparison of mean h_b calculated from Eq. (8), experimental values from virtual tower TKE, and the proposed model described in Eq. (14) from 1900 to 0705 UTC. The boxplots depict the statistical spread in values of h_b calculated from virtual towers: the horizontal red line displays the median; the bottom and top edges of the box display the 25th and 75th quartiles, respectively; whiskers show range of values not considered outliers; and red pluses show outlier values. All values are normalized by time-averaged z_i .

was oriented along the mean wind direction at each height and spatial fluctuations from the spatial mean were constructed as

$$\bar{u}'' = \bar{u} - \langle \bar{u} \rangle, \quad (15)$$

and likewise for w and θ . A fast Fourier transform was then computed along the x direction after which power and cospectral densities were constructed. The results were finally averaged along the y direction to reduce the noise.

Figure 12 shows the power spectra and cospectra at 20 m above ground. At that height we clearly see spectral peaks in all of the cospectra at the wavenumber corresponding to π/L_f , showing the contribution of dispersive fluxes. Notice that for both the heat flux and the momentum flux, the peaks for the dispersive stresses are positive, contrary to the turbulent fluxes. Above 20 m, however, these peaks diminish rapidly and become comparable to the noise level for all L_f above 50 m. For the power spectra the peaks remain visible higher up, especially for the vertical velocity at larger L_f , similar to the pattern in Fig. 9. Note that the spectral peak in the cospectra for the case with $L_f = 120$ m is relatively small at 20 m; 20 m is approximately the height of h_b for this case, so this implies that the dispersive stresses have been distributed over a wider wavenumber range or that dispersive stresses are insignificant. Even though 120 m only corresponds to 6 grid points in length, the small peak in the uw cospectra does not seem to be explained by resolution issues, since the peaks in S_{uw} , $S_{\theta\theta}$, and $CoS_{w\theta}$ are prominent also for the 120-m patch length. Similar spectral peaks can be seen in other cases, although the peaks are less prominent for L_f of 1500 and 3000 m since the peak at this wavenumber range is approaching the domain size.

f. Effect of effective roughness on wind power estimates

Earlier we showed an inverse dependence of $z_{0\text{eff},m}$ on L_f . This suggests that subgrid heterogeneity may cause the effective surface roughness to deviate significantly from fixed values

determined from lookup tables in coupled mesoscale models. To highlight the impact of this surface roughness change on wind energy applications, a reference simulation was run with otherwise the exact same setup as the other LES experiments in this study, but using the LSM to define each grid cell as 50% of the cover types used in the experimental runs. Using the effective roughness analysis described above for this simulation, the mean $z_{0\text{eff},m}$ for neutral and stable conditions is 0.2822 m, or 25% higher than $z_{0\text{eff},m}$ for the case of $L_f = 20$ m. We then used Eq. (1) to create wind profiles with the $z_{0\text{eff},m}$ of each simulation using $u_* = 0.2 \text{ m s}^{-1}$. To estimate the power produced by the wind turbine, the rotor equivalent wind speed v_{eq} described in Wagner et al. (2014) was used:

$$v_{\text{eq}} = \left(\sum_{m=1}^{n_h} U_m^3 \frac{A_m}{A} \right)^{1/3}, \quad (16)$$

where the $A = \pi R^2$ is the swept area of a turbine rotor with radius $R = 40$ m, U_m is the wind speed at the model level m within the wind turbine sweep, n_h are the total number of model levels within the turbine sweep,

$$A_m = \int_{z_m}^{z_{m+1}} c(z) dz \quad (17)$$

is the area of the wind turbine sweep between model levels m and $m + 1$, z_m is the height of model level m , $c(z) = 2\sqrt{R^2 - (z - H)^2}$, and H is the turbine hub height. Assuming a wind turbine with 80-m hub height and a rotor radius of 40 m, we calculated the estimated power produced by a wind turbine as a function of $z_{0\text{eff},m}$. These results are shown in Table 3 as relative to the roughness for the case with $L_f = 20$ m and to the effective roughness length when each grid cell was 50% of each cover type aggregated by the LSM. Here it is readily apparent that the effective roughness can impact wind power estimates

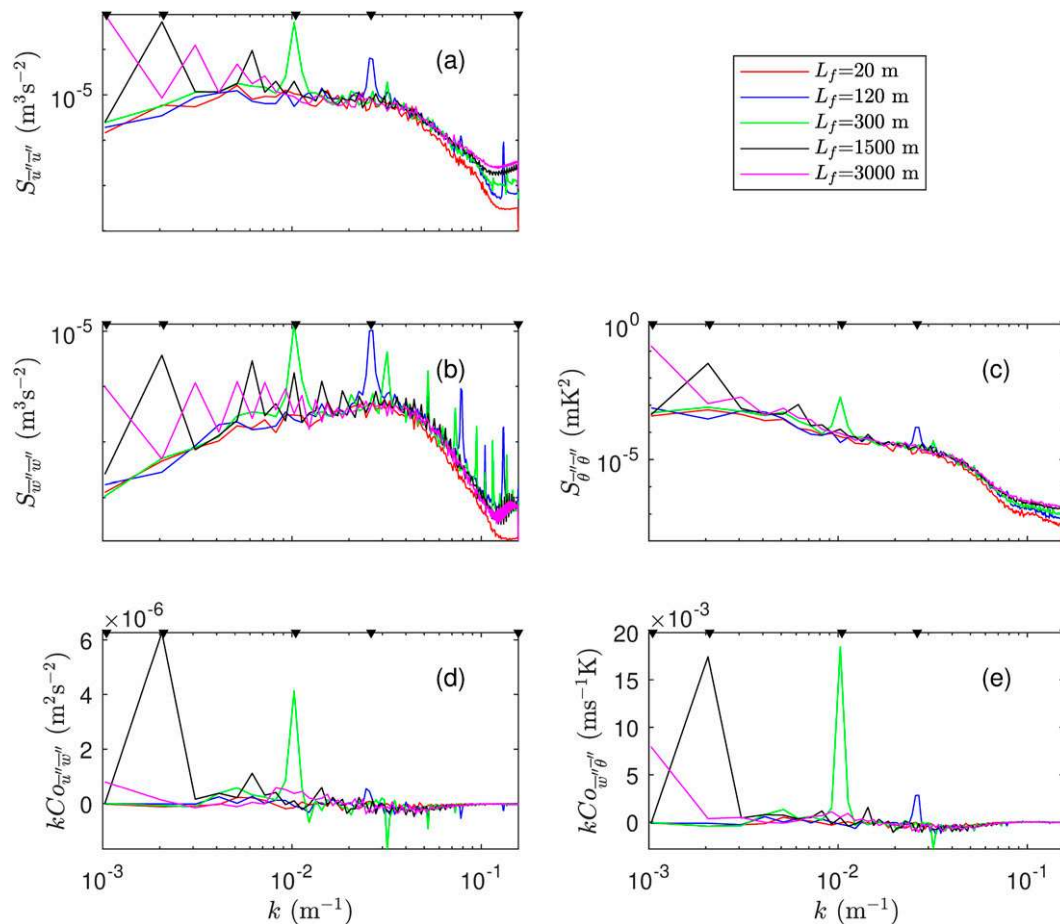


FIG. 12. Spectral densities at 20-m height: power spectra for (a) streamwise wind speed, (b) vertical wind speed, and (c) potential temperature and premultiplied cospectral density of (d) $\bar{u}''\bar{w}''$ and (e) $\bar{w}''\bar{\theta}''$; the time average was between 1903 and 0403, the spectral transform was along the x direction, and averaging of the spectra was along the y direction. Black triangles represent π/L_f .

for wind turbines of 80-m hub height; the impact is greatest for smaller values of L_f . Relative to the case with $L_f = 20$ m, the wind profile using the effective roughness calculated from the case of $L_f = 120$ m results in a 9% increase in wind power. Using wind profiles calculated using the aggregated roughness length from the LSM aggregation results in a 5% increase in mean wind power relative to the same analysis using the effective roughness calculated from the simulation with $L_f = 20$ m. For larger turbines, with hub heights significantly higher than

the blending height, the differences in estimated wind power are much smaller for different L_f .

5. Discussion

Given that mesoscale models are often used in wind energy applications, we can see in the results the practical importance of taking land-cover heterogeneity into account when developing parameterization schemes for NWP models and for wind atlases.

We examined the impact of L_f on the inertial oscillation and low-level jet. The response of the low-level jet maximum speed was positively correlated to the effective roughness length. This proportional dependency of the strength of the low-level jet on surface roughness follows the results described by Heinold et al. (2015). To rule out impacts of secondary circulations during the convective period, we tested the experimental setup with $u_g = 1$ and 4 m s^{-1} , in addition to the 8 m s^{-1} large-scale forcing used in this study. Stable convective cells only developed for $u_g = 1 \text{ m s}^{-1}$. These findings were not surprising, as previous studies have shown (Kröniger et al. 2018; Margairaz et al. 2020a,b) that secondary

TABLE 3. Percent difference of wind power calculated for each $z_{0\text{eff},m}$ for near-neutral and stable conditions.

L_f (m)	Percent diff from $L_f = 20$ m case	Percent diff from SURFEX aggregated z_0
20	—	+5
120	+9	+15
300	+11	+17
1500	+14	+20
3000	+15	+22

circulations can develop over heterogeneous landscapes and contribute to the sensible heat flux during daytime, but only when geostrophic forcing is relatively weak.

For stable conditions, analysis of the spatial spectra and co-spectra confirmed existence of a spectral peak at 10 m above the surface that corresponds to the length of L_f . The spectral peak was not present (or was greatly diminished for larger L_f) at higher elevations. In a similar spectral and cospectral analysis over thermal surface heterogeneity, Margairaz et al. (2020b) showed that—for larger u_g —the effects of heterogeneous surface patterns on covariances away from the surface was stretched and redistributed to different scales. In addition to the lack of secondary circulations found in the analysis above for $u_g = 8 \text{ m s}^{-1}$, these results confirm that the effects of dispersive fluxes are under these conditions insignificant far away from the surface heterogeneity patterns.

We used virtual towers to examine the blending height as a function of L_f . Comparing the simulated virtual tower observations of TKE averaged over low vegetation minus those averaged over forest patches for $\zeta > 0.2$ (Fig. 10), we see a systematic increase in height at which the TKE is equal over those areas. The average height at which this occurs is roughly equal to the virtual tower h_b that is seen from visual inspection of Fig. 9 (horizontal black lines). When comparing the experiments as a function of L_f , we see that there is not a large difference in h_b for $L_f = 1500$ and 3000 m . In contrast, Bou-Zeid et al. (2004) suggest that h_b increases nearly monotonically with L_f , which would lead to errors in h_b for large L_f in stable conditions. The depth of the IBL is quite dependent upon the height of boundary layer z_i , as the analysis shows. In an LES study over the 7-km-long Yatir forest in Israel, Brugger et al. (2018) also found a dependence of the IBL height on z_i ; they state that energy is required for an IBL to influence the boundary layer height and that this required energy increases with stability. In their case, Brugger et al. (2018) only saw an influence of the IBL on the local z_i under weak geostrophic forcing with an unstable boundary capped by a weakly stable free atmosphere. Since the ABL shown in Fig. 9 is weakly stable under a stable free atmosphere with moderate u_g it is not surprising that the IBL height is limited by z_i . The sensitivity tests using weaker large-scale forcing of $u_g = 1$ and 4 m s^{-1} showed similar behavior of h_b and $z_{0\text{eff},m}$ as a function of L_f , even with much lower values of z_i (4 and 16 m, respectively) and less resolved turbulence.

In contrast to the analysis in stable conditions, the h_b results were noisy and inconclusive for unstable conditions. Blending-height identification in unstable conditions has been contested in earlier literature. In their study of aggregation techniques over heterogeneous surfaces, Hu et al. (1999) defined the blending height strictly over nonconvective boundary layer conditions. Using aircraft observations over thermal internal boundary layers in convective conditions, Strunin et al. (2004) concluded that the blending-height concept is unable to estimate turbulent fluxes in convective conditions over heterogeneous terrain. In an LES study of unstable atmospheric conditions over heterogeneous land cover, Sühring and Raasch (2013) were unable to find a blending height at all using correlation analysis.

Last, we compared the estimated power produced by a wind turbine calculated from idealized wind profiles using the effective roughness length as a function of L_f versus the roughness value that would be aggregated by the LSM if each grid cell contained 50% of each cover type. The surface parameterization schemes within mesoscale models regularly use values of surface roughness from lookup tables that have been shown in previous studies to misrepresent the values expected in local areas used in wind energy applications (Dellwik et al. 2014; Floors et al. 2018). Note that Basu and Lacser (2017) caution against the use of Monin–Obukhov theory at z within 2–5 times the height of the roughness elements, which becomes a problem when running LES simulations with vertical grids very close to the ground. Since our analysis was conducted using profiles up to 12 m, we avoid these errors.

In regions with heterogeneous land cover, the effective roughness of a local region obtained from LES studies or other algorithms may offer an alternative to general lookup tables when using a mesoscale model for wind energy uses. In a study using a formulation for effective roughness, Cheng et al. (2019) found that using the effective roughness reduces errors in wind speed and momentum flux when used in a mesoscale model.

Given its ability to resolve small-scale processes resulting from features such as land-cover heterogeneity, LES remains a powerful tool. It is also noted that the land-cover patterns used in this study are well-defined patches, while natural heterogeneous land surfaces may contain varying length scales of patches with different roughness and radiative characteristics. It is noted that the borders between land-cover types may not be as well defined as the idealized surfaces used in this study and the surface parameterization used here does not model complex processes such as forest canopy effects, the edges of which have been shown to have significant effects on the internal boundary layer turbulence level (Kanani et al. 2014). Another possible uncertainty in the presented results is that the grid spacing of the LES may be coarse for strongly stable conditions (Beare et al. 2006). However, Arnqvist et al. (2019) showed that LES with relatively coarse resolution can be useful at elevations considered in wind energy applications and that other issues, such as model complexity and domain size might be more critical to obtain results closer to observations.

The latest land-cover maps used in surface-layer parameterization schemes feature horizontal resolutions approaching 300 m (Tzanos et al. 2018). While this allows for a more detailed representation of surface cover than past surface cover maps, we show here that the effective surface roughness is quite sensitive to changes in patch size when L_f is small. High-resolution mapping of land-cover properties using, for example, laser scans of forests (Floors et al. 2018) combined with more realistic representations of forest canopies available for LES modeling should be used to further examine the sensitivity of atmospheric models to land-cover heterogeneity in comparison with observations taken within these areas.

6. Conclusions

In this study we used LES to examine the impacts of two-dimensional land-cover heterogeneity on the boundary layer

given a diurnal cycle. In near-neutral and stable conditions the effective roughness length due to momentum was found to decrease with increasing heterogeneity length scale; this follows previous findings in similar studies (Bou-Zeid et al. 2004; Miller and Stoll 2013) but extends the analysis to large patch sizes with a diurnal atmospheric pattern. The mean effective roughness calculated from the LES wind profiles decreases by nearly 50% from the case with roughness patches alternating every other grid cell when compared with values calculated with the algorithm that the SURFEX LSM uses to aggregate subgrid roughness length. The mean effective roughness continues to decrease by 44% from patch lengths of 20 m to patch lengths of 3000 m, with the largest changes occurring for smaller patch sizes. When estimating the wind power produced from an 80-m wind turbine using the wind profile calculated from the range of effective roughness, we found a wind power increase of up to 23% when compared with wind profiles using a default value determined from an LSM roughness aggregation algorithm. As relatively coarse mesoscale models are often used in wind energy applications for operational forecasting, the representation of surface roughness due to land surface heterogeneity within these models can have a significant impact on wind power estimates. The height of the wind speed maximum within the nocturnal low-level jet was found to decrease with increasing land-cover heterogeneity, with a slight decrease in wind speed.

We also tested the impact of heterogeneity length scales on the blending height, which was found to match the characteristics of the blending height when predicted with the Bou-Zeid et al. (2004) model for small heterogeneity length scales. For heterogeneity length scales of 1500 and 3000 m, the blending height increased asymptotically to the boundary layer height. As the blending height has been proposed as a means for coupling LSMs with the atmosphere within coarse mesoscale models, its value is important for determining the parameterization scheme used in calculating wind profiles and fluxes within these models.

The results of this study shed light on the low-level impact of a wide range of heterogeneity length scales given a diurnal insolation cycle. Future work will examine the impact of the resolution of land-cover data on the lower atmosphere and to develop strategies for coupling LSMs to atmospheric models using the insight gained about the blending height.

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Data availability statement. The numerical model simulations described in this work are too large to archive or transfer. Instead, we provide all of the information needed to replicate the simulations; we used Meso-NH, version 5.4.3, with SURFEX, version 8.1.

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