



The Persistent Challenge of Surface Heterogeneity in Boundary-Layer Meteorology: A Review

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Received: 18 March 2020 / Accepted: 9 July 2020
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Abstract

Atmospheric boundary-layer dynamics over heterogeneous surfaces is significant to a wide array of geophysical and engineering applications. Yet, despite over five decades of intense efforts by the research community, numerous open research questions remain. This underlines the complexity of the physical processes that are excited by heterogeneity, the multitude of patterns and manifestations that it can display, and the importance of the implications to research in the atmospheric sciences and beyond. Here, existing knowledge is reviewed and a path forward for research is proposed, starting with the smaller scales near a surface transition and proceeding to the influence on large-scale dynamics and their forecasting.

Keywords Heterogeneity · Internal boundary layer · Internal equilibrium layer · Secondary circulations · Land-surface models

1 Introduction

Land-surface heterogeneity comes in many varieties and occurs over a wide range of spatial scales. As such, it has continued to defy a generalized approach or theory for “characterizing” its impact on the atmospheric boundary layer (ABL) (Mahrt 2000). Its multiscale nature also makes it practically impossible to “resolve” all the relevant spatial information directly in geophysical simulations (Giorgi and Avissar 1997) or through observations (see *Boundary-Layer Meteorology* special issue, Vol. 121, on the LITFASS–2003 campaign). The scales that would have to be accommodated (i.e., those that may interact with turbulent eddies and the mean flow) range from ~ 1 mm (Kolmogorov microscale) to spatial scales that may be much larger than all integral scales of turbulence ($\sim$ several km or more). Heterogene-

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ity therefore poses a challenge that we might never completely “solve”. However, it is a ubiquitous characteristic of the real-world and it is further intensified by anthropogenic land-use modification. It is thus a key determinant of microscale to synoptic-scale atmospheric dynamics, particularly in regions (agricultural, urban, etc.) where climatic projection and meteorological forecasting are most critical for human activities and well-being. As such, the challenge posed by surface heterogeneity for boundary-layer meteorology is not one that can be ignored or circumvented (Pitman 2003; Prueger et al. 2012). Important examples where the heterogeneity is a first-order determinant of ABL development include: land–sea breezes, urban–rural transitions and breezes, wind-farm boundary layers, airflow and evaporation over lakes and reservoirs, and land-surface exchanges over patchy agricultural terrain or over fractured sea-ice, among others. This importance is indeed reflected in the wide array of studies that have attempted to advance the understanding of the ABL over heterogeneous, relatively flat, terrain (the added complexity of hilly terrain is not covered in this review).

Boundary-Layer Meteorology has played a pivotal role in disseminating advances on the impact of heterogeneous surfaces on the lower atmosphere. In fact, a paper on this topic appeared in the very first issue of the journal (P.A. Taylor: *Boundary-Layer Meteorology*, 1970, 1, 18–39). The research efforts have mainly aimed at, (i) understanding the microscale effects of various patterns of heterogeneity through local-scale experiments or simulations that can resolve them, e.g., internal boundary layers (Garratt 1990), the role of advection (Rao et al. 1974; Fontan et al. 2013; Higgins et al. 2013), the onset of secondary circulations (Raasch and Harbusch 2001; Fontan et al. 2013), the impact on surface fluxes and their measurements (Esau 2007; Kenny et al. 2017), and (ii) applying this microscale understanding to developing surface models at weather and climate simulation grid scales that can implicitly and efficiently parametrize the unresolved scales of surface variability (e.g., Giorgi and Avissar 1997; Bou-Zeid et al. 2004; Ament and Simmer 2006; Stoll and Porté-Agel 2009; Li et al. 2013).

The microscale effects, defined here as the well-delineated and rapid changes occurring in the ABL near a discontinuity in the boundary condition, lend themselves to some generalizations. Multiple studies, for example, have looked at the common problem of an abrupt transition in surface properties (roughness, temperature, humidity) along an interface that is either perpendicular or parallel to the prevailing surface wind direction. These set-ups lead to the formation of internal boundary layers or to the onset of secondary circulations (e.g., land–sea breeze), and the physics of flow features excited by such transitions are reviewed in Sect. 3. Studies on oblique transitions remain scarcer, but indicate that the flow and transport may not simply be a fusion of the parallel and perpendicular transition counterparts (Raasch and Harbusch 2001; Anderson 2020). The interaction of multiple heterogeneity interfaces is an even more daunting challenge since the simpler microscale flow features can merge and interact, which they almost always do in real-world scenarios. Such interactions could be numerically simulated at fine scales (e.g., using large eddy simulation, LES) or probed using intensive field campaigns or scanning lidars. However, the uniqueness of each heterogeneity pattern requires some simplified approaches to enable the parametrization of the aggregate effect of surface variability at regional scales in coarser atmospheric models. These efforts are reviewed in Sect. 4, while Sect. 5 presents a summary and recommendations for future work. For context and comparison, we first propose a simple classification of heterogeneous surfaces that only considers the surface characteristics, independently of the flow aloft.

2 Classes of Heterogeneity

We focus on heterogeneity over relatively flat, non-mountainous terrain, and are concerned primarily with variability in land use that occurs at scales commensurate with tens of metres or larger (about 50 m for a reference numerical value). We therefore exclude the small-scale variability created by individual roughness elements such as buildings (e.g., Kastner-Klein and Rotach 2004; Li and Bou-Zeid 2019) or trees (e.g., Bailey and Stoll 2013), or by mild variability in the roughness properties at the scale of a few roughness elements (e.g., Bohrer et al. 2009). These smaller-scale variations give rise to the “roughness sublayer”, and have been extensively studied to examine flow variability in that layer and the role of dispersive stresses or fluxes (Finnigan 2000; Poggi et al. 2004; Poggi and Katul 2008; Li and Bou-Zeid 2019).

The framework we adopt instead seeks to identify patches of ‘nearly-uniform’ bulk surface properties (geometry/roughness, temperature, humidity, etc.), each patch having its own length scale L_p (again ≥ 50 m). This framework presupposes that the properties of the roughness elements within each patch are statistically uniform in space (e.g., tree height or urban skin temperature distributions), but vary significantly between patches. One plausible classification system that can then be proposed, covering a wide range of real-world terrain, consists of the four types illustrated in Fig. 1 (Mahrt 2000; van Heerwaarden et al. 2014; see also Avissar and Schmidt 1998; Shen and Leclerc 1995):

(I) Semi-infinite interfaces such as the land–sea boundaries or the edges of large forests. Practically, these are very large patches (> 100 km, or 100 times the ABL depth $\delta \approx 1000$ m) such that the flow perturbation at their interface is locally relevant and very consequential, but at much-larger scales the atmosphere is predominantly interacting with (statistically) homogeneous surfaces where the ABL has equilibrated with the underlying surface over much of the patch area. Classic laws and their findings (e.g., the flow statistics mainly vary with the vertical coordinate z) apply except near the transition, where the flow is highly disturbed. These interfaces were the subject of various studies that proposed scaling arguments, parametrizations, and linear theories for sea and land breezes (e.g., Haurwitz 1947; Rotunno 1983; Porson et al. 2007a; Crosman and Horel 2010). For this class, the interest is in the interfacial area (that might extend to tens of kilometres in the case of land–sea breezes), how the flow transitions across the interface, and the secondary circulations that can be driven by differences in heat flux or surface stress and their effect on the local micrometeorology and hydrometeorology.

(II) Statistically-homogeneous patches of land. These are locally-variable surfaces, but with statistical properties that are homogeneous at regional ($\geq$ meso) scales (Brutsaert 1998). Examples include agricultural surfaces for which the mean roughness or patch-size variance (among others) is spatially homogeneous when averaged at scales $\geq 10\delta$. If these patches have (micro) scales ~ 50 m, the height at which the surface heterogeneity effect is homogenized or blended out is low (~ 20 m) and as a result the mean flow is horizontally homogeneous over most of the depth of the atmospheric surface layer (ASL) (Mahrt 2000; Bou-Zeid et al. 2004). This prompted Brutsaert (1998) to distinguish between surface variability and ABL heterogeneity: while the surface flow is spatially variable for this class, the ABL can be considered effectively homogeneous (see also Parlange and Brutsaert 1993). This type has been extensively studied over the past two decades with the primary aim of developing parametrizations for its interaction with the atmosphere (Parlange et al. 1995; Bou-Zeid et al. 2004; Patton et al. 2005; Lopes et al. 2015; Kanani-Sühring and Raasch 2017). These parametrizations usually take advantage of the flow homogeneity and low blending height,

above which classic laws such as the Monin–Obukhov similarity theory (MOST) apply, to represent the patches as a homogeneous surface with “equivalent” or “effective” properties (Parlange and Brutsaert 1989; Albertson and Parlange 1999; Bou-Zeid et al. 2007, see Sect. 4).

(III) Large, isolated patches (1D or 2D) with land uses different to their surroundings. Examples abound, including cities, deforested patches, lakes, islands, wind farms, rivers, or polar polynyas and leads. This type is more challenging than the previous classes since, as various studies point out (Mahrt 2000; Kang 2009; van Heerwaarden et al. 2014), the heterogeneity occurs at (γ and smaller β) mesoscales: it is neither too small for turbulence to blend its effect rapidly (as for type II) nor too large for the atmosphere to be mostly in equilibrium with the underlying surface (as for type I). The result is a breakdown of classic theories and the preponderance of mesoscale secondary three-dimensional (3D) circulations that are highly energetic and often completely dominate the dynamics (Kang 2009; Kang and Lenschow 2014; Omidvar et al. 2020). These circulations confound regional-scale parametrizations since they induce non-linear dynamical effects that are essential to capture (Giorgi and Avissar 1997; Chen et al. 2019).

(IV) Unstructured heterogeneity where the surface variability cannot be approximated as one of the previous types, because it is a mixture of multiple types. It covers large parts of the land surface. Currently, coarse atmospheric models mostly treat these patches as independent and non-interacting. Tiled or mosaic approaches (Sect. 4) allow the models to represent each patch even when its scale is smaller than the atmospheric-model grid scale, but the subgrid-scale atmospheric variability is often ignored and flux models for homogeneous surfaces are used [see discussion on the role of atmospheric variability in Bertoldi et al. (2007, 2008); Timmermans et al. (2008)]. A formal approach to understanding the complex flow patterns over such surfaces and their regionally-averaged characteristics remains a crucial research gap.

The importance of the scale of the variability is noted above. Small patches influence and interact with smaller eddies of commensurate scale and influence the lower part of the ABL, while very large patches affect the largest of scales and the Reynolds-averaged mean flow, potentially across the whole ABL depth (Shen and Leclerc 1995; van Heerwaarden et al. 2014). A natural normalization of the patch scale is therefore with the ABL depth δ , which is comparable to the vertical extent of largest eddies. The horizontal extent of these eddies is even larger, $\sim 10\delta$, and they have been shown to contribute significantly to momentum transfer and velocity variances (Katul 2019). Therefore, while our present classification intentionally focuses on the spatial scale of the surface features, the impact of heterogeneity on the ABL will be modulated by the interaction of this surface scale with the typical flow scales and how the latter vary with ABL stability (Margairaz et al. 2020, see their Fig. 1). Varying density gradients will yield differing signatures in flow statistics as airflow encounters each of these heterogeneity classes.

3 The View at the Microscale

In addition to these large-scale motions, turbulence over flat terrain is inherently rich in length scales that can interact or resonate with land-surface heterogeneity. Of particular relevance to land–atmosphere interaction are the scales of eddies where turbulence is locally produced. These scales can span the distance from the ground (z , the attached eddies), all the way to $z = \delta$. Turbulence is anisotropic at those production scales, with the ratio of the integral length scale of the horizontal to vertical velocity components exceeding 10 for near-neutral

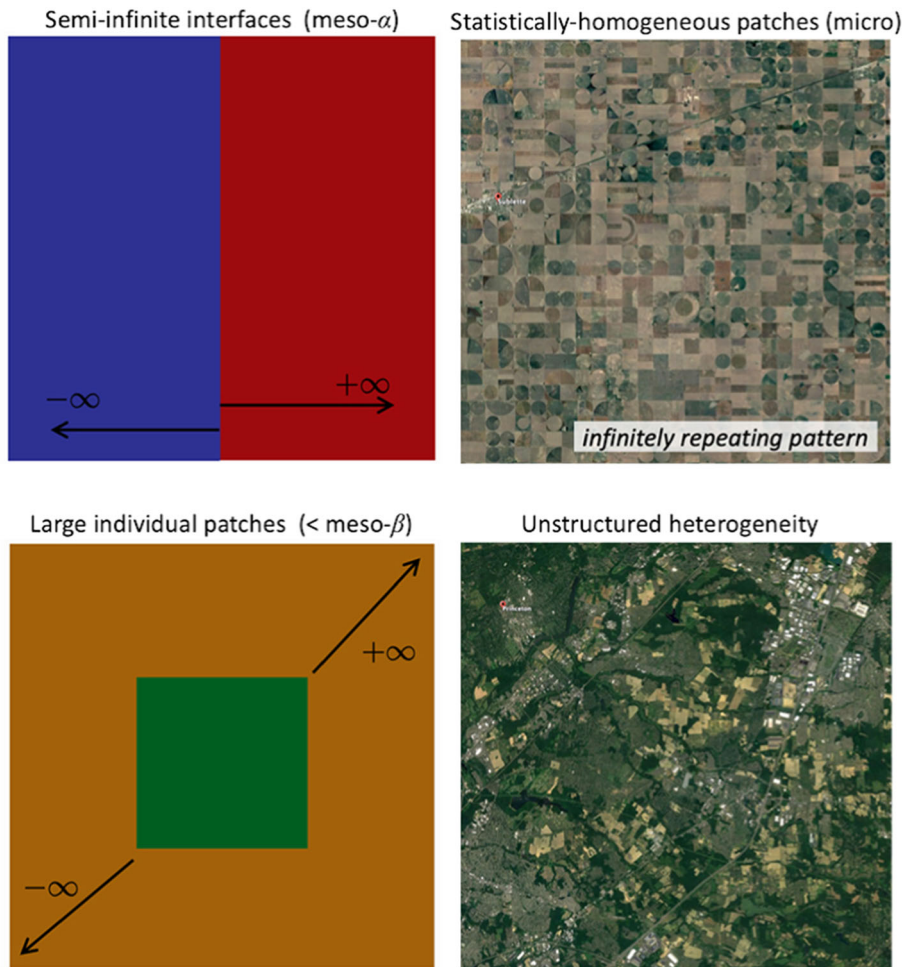


Fig. 1 The four scales/classes of heterogeneity: (I) in the top left is the canonical single transition (e.g., land–sea boundary). (II) in the top right is the infinitely repeating pattern of small patches: similarity theories could apply here above the blending height (this is the aerial view of the Kansas experiment location as it is today, scale ~ 50 km). Type (III) in the bottom left is a canonical single patch in an otherwise homogeneous surrounding (city, lake, wind farm). Type (IV) is the unstructured heterogeneity where no clear pattern can be identified. Classic similarity theories fail for types III and IV due to the complex interaction of internal boundary layers and secondary circulations

conditions (Kader and Yaglom 1990). This ratio varies appreciably (5–50) with atmospheric stability (Li et al. 2012). A large scale-separation exists between these production eddies and the smallest dissipative Kolmogorov scale motions, the latter being on the order of 0.1–1 mm in the lower atmosphere. This implies that a disturbance injected by surface variability at one temporal or spatial scale will not instantaneously propagate to all frequencies or wavelengths (Momen and Bou-Zeid 2017), and the energy cascade and spectrum might not be in equilibrium (Sreenivasan 1999). One of the primary challenges in studying the impact of surface heterogeneity on turbulent flows is that a disturbance rarely occurs in isolation or at a single wavelength, even though in many circumstances the disturbance is localized in

physical space (e.g., sharp edges delineating smooth-to-rough transitions). As is well-known, a localized disturbance in physical space is non-local in spectral (or wavenumber) space (humorously labelled as the Heisenberg ‘curse’ in some atmospheric turbulence literature, [Katul and Vidakovic (1998)]).

Therefore, near and downstream of changes in surface properties, various physical processes at multiple scales (that are absent over homogeneous terrain) are excited. Horizontal advection, secondary mesoscale circulations, and turbulence non-equilibrium (where production and dissipation processes are not in balance) are of particular importance, and tracking their effect is complicated by the potential variability in more than one surface property at a time. Prior studies have focused on a discontinuity in one of the surface properties, but simultaneous changes in multiple properties are the norm, not the exception. For example, cities are generally hotter, rougher and drier than their surroundings, while lakes are generally smoother, cooler, and more humid (during daytime). A few studies have examined the interacting and simultaneous effects of changes in multiple surface properties (e.g., Bertoldi et al. 2007, 2008); but more are needed. Coupling idealized turbulence-resolving simulations to a full surface-energy-budget (SEB) model would allow more realistic investigations. For example, thermal heterogeneity can be driven, via the SEB, by variations of soil moisture (adjacent irrigated and non-irrigated fields) and other soil characteristics or horizontal variations of stomatal control (Albertson et al. 2001; Katul et al. 2012). Variations of surface temperature and moisture content may occur with or without changes in roughness.

Following a change in properties of the underlying surface under moderate to strong winds, the mean wind field and the turbulence begin to adjust to the new boundary conditions. Very close to the surface, an internal equilibrium layer (IEL) [sometimes also referred to as the equilibrium boundary layer, EBL (Garratt 1990)], develops in which the flow (mean and higher moments) is in complete equilibrium with the new surface and the effect of upstream conditions is negligible. The depth of this layer typically scales with the distance from the surface change x , and a rough estimate under near-neutral conditions is $\delta_e \sim x/100$ (Brutsaert 1998). Further aloft, the flow has responded to the new surface but equilibrium has not been attained; this region is referred to as the internal boundary layer (IBL) and its depth $\delta_b \sim x/10$ (Brutsaert 1998; Bou-Zeid et al. 2004), again under near-neutral conditions. Above the IBL, the flow is unchanged from the upstream inflow. This idealized layered framework is sketched in Fig. 2, bottom panel.

If the surface change enhances turbulence (e.g., an increase in roughness or heat flux) the equilibration is accelerated (deeper IEL and IBL), while the opposite occurs under low turbulence conditions (e.g., stable flow) or when the surface change reduces turbulence (smoother or cooler surface) (Garratt 1990; Bou-Zeid et al. 2004). With multiple transitions, many IBLs and IELs develop. The elevation above which turbulence homogenizes the signature of individual patches is known as the blending height (Wieringa 1976). To model the evolution of the IEL and IBL and the ultimate blending, concepts of advection–diffusion balance are usually applied. This can be achieved by an analytical solution to the horizontal advection–vertical diffusion equation (Sutton and Brunt 1943; Yeh and Brutsaert 1971; Li and Bou-Zeid 2013). However, simpler scaling can be obtained (Miyake 1965; Panofsky and Dutton 1984; Garratt 1990; Bou-Zeid et al. 2004) by considering an air parcel (or column) advecting past the transition: its distance downwind of the transition is the product of the mean advective wind speed U_a and the time elapsed since it crossed the discontinuity t_a , while the height to which the new surface information has been transmitted by turbulent mixing (top of the IBL) is given by a (turbulent) diffusive length scale $\propto (K_t t_a)^{1/2}$, where K_t is a turbulent viscosity/diffusivity. The top interface of the IBL can then be traced as the location where $x = C(K_t t_a)^{1/2}$; that is, the IBL depth varies with x following $\delta_b = C(K_t x/U_a)^{1/2}$, where C

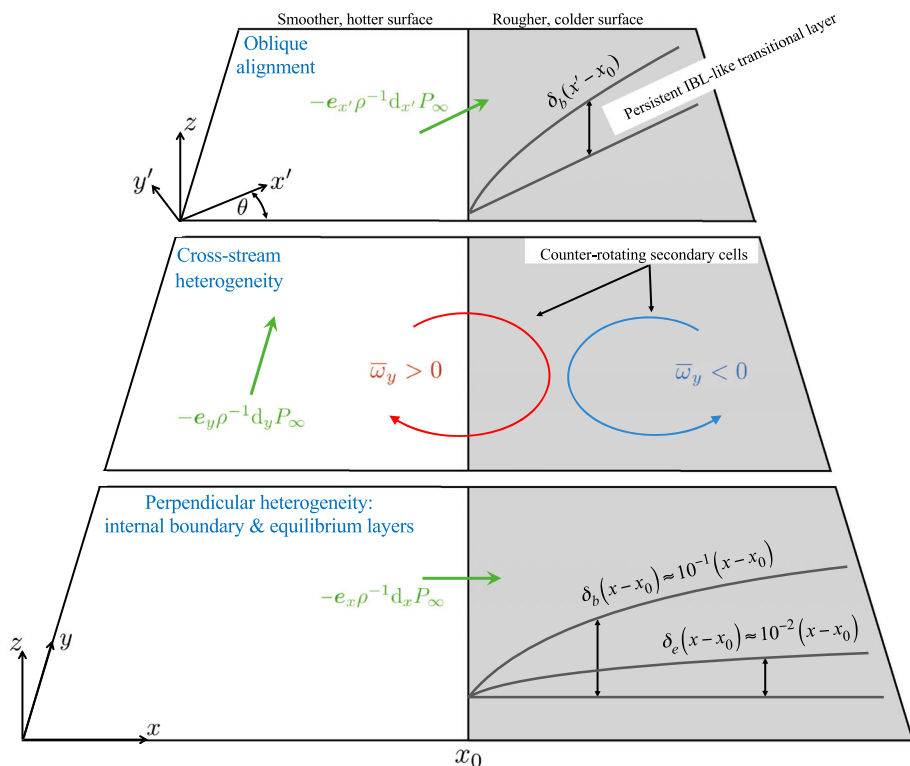


Fig. 2 Bottom panel: depiction of the IEL and IBL development over a discontinuity in surface properties. Middle panel: secondary circulations induced by surface stress or buoyancy flux cross-stream heterogeneity. Top panel: mixture of streamwise and cross-stream heterogeneity. P_∞ is the background synoptic pressure; all other variables are as defined in text

is a proportionality constant (~ 0.8), and K_t and U_a depend on surface conditions as well as on x . Over more complex surfaces, these concepts may be challenging to apply, particularly when the turbulence is not in equilibrium (as discussed later).

3.1 Secondary Circulations

When synoptic forcing and/or prevailing flow are weak (Omidvar et al. 2020), or when the mean velocity has a significant component parallel to the interface between different patches (Raasch and Harbusch 2001), large and persistent circulations develop in the ABL. The vorticity of these structures is either driven by differences in surface buoyancy flux between adjacent landscapes (thermal torque, Eq. 1), or by the divergence–convergence of the mean flow as it experiences variability in surface drag (turbulent torque, Eq. 1). For a case with cross-stream varying aerodynamic and/or energetic conditions (Fig. 2, middle panel) and weak synoptic wind forcing, this can be readily appreciated via consideration of the Reynolds-averaged transport equation for the component of vorticity aligned parallel with the heterogeneity interface ω_y ,

$$\underbrace{\overline{u}\partial_x\overline{w}_y + \overline{w}\partial_z\overline{w}_y}_{\text{Advection of } \overline{w}_y} = \underbrace{(\partial_z^2 - \partial_x^2)\overline{T}_{xz} + \partial_{xz}(\overline{T}_{xx} - \overline{T}_{zz})}_{\text{Turbulent torque}} + \underbrace{\epsilon_{yjk}\partial_j\Theta_k}_{\text{Thermal torque}}, \quad (1)$$

where the overbar denotes a Reynolds-averaged quantity, u , v , and w are the streamwise (x), cross-stream (y), and vertical (z) components of the velocity vector u_i , turbulent (Reynolds) stresses are denoted by $\overline{T}_{ij} = \overline{u'_i u'_j}$, and ϵ_{ijk} is the alternating unit tensor. In the last term, $\Theta_k = -g_k (\delta\theta/\theta_0)$, denotes the buoyancy force due to large thermal gradients, where g_k is the downward-pointing gravitational acceleration vector, θ_0 the reference potential temperature, and $\delta\theta$ is the potential temperature perturbation driving the buoyant forcing. Note that the buoyancy term for the considered alignment reduces, upon expansion of the alternating unit tensor and given that g_k is aligned in the wall-normal direction, to $-g\partial_x(\delta\theta/\theta_0)$. As per the underbrace terms in Eq. 1, the first and second right-hand side terms represent the contribution to $\overline{w}_y$ from spatial heterogeneity of the Reynolds stresses, while the last emanates from the cross-stream heterogeneity of buoyancy forces. The Reynolds-stress induced secondary circulation is known as Prandtl's secondary flow of the second kind (Prandtl 1952; Bradshaw 1987) (note that the 'first kind' production, $\overline{\omega} \cdot \nabla \overline{u}$, vanishes in this example due to the absence of streamwise heterogeneity). These circulations and their forcing have only been recently studied in the context of boundary-layer meteorology (Willingham et al. 2013; Anderson et al. 2015, 2018; Awasthi and Anderson 2018; Anderson 2019).

The most prominent example of thermal circulations induced by the last term in Eq. 1 is the land-sea breeze, which is driven by strong temperature contrasts that enable it to persist even under strong winds driven by synoptic pressure gradients. As previously discussed, various studies have developed scaling arguments, parametrizations, and linear theories for sea and land breezes. Steyn (1998, 2003) and Porson et al. (2007a, b) proposed a dimensional analysis of the problem and evaluated it against mesoscale model simulations; the driving dimensional parameters for the circulations being the land-sea temperature contrast as well as the integrated heat from the surface to the atmosphere. Haurwitz (1947) and Rotunno (1983) sought analytical solutions of the momentum and heat budget equations to describe the circulation. Crosman and Horel (2010) reviewed numerical studies of sea and lake breezes and the dimensionless parameters proposed in the literature. What is still needed are studies to contrast and unify these various scaling and theoretical approaches using LES for evaluation since Reynolds-averaged closures employed in mesoscale models might have difficulty resolving temporally- and spatially-evolving turbulence fields. Less studied, too, is how such circulations respond to synoptic conditions: it is expected that strong synoptic forcing will modulate and may weaken the secondary structures [as suggested by the theoretical model of Haurwitz (1947)].

While land and sea breezes have been the most widely studied, secondary thermal circulations are common in many other set-ups (in classes I or III of Fig. 1) and the physics and potential models are quite similar across these flows. Similar breezes can for example develop around lakes or rivers (Wrenger and Cuxart 2017). The 3D flow structures around a city have consequential impacts on air quality and excessive heat affecting citizens; they have been examined in many studies (Delage and Taylor 1970; Ryu et al. 2013; Fan et al. 2016, 2017, 2018; Liang et al. 2018; Omidvar et al. 2020). The flow patterns around polynyas or individual leads (these are warmer polar water surfaces surrounded by colder sea-ice) can alter the transport of atmospheric aerosols and trace gases, and thus modulate cloud formation over large regions (Khvorostyanov et al. 2003; Esau 2007). The flow inside and downwind of a wind farm is self-evidently critical to research on that version of the single-patch problem (Calaf et al. 2010; Hezaveh and Bou-Zeid 2018). As such, understanding and predicting the flow details of these circulations have myriad applications.

Despite these advances, there remains a clear deficit in understanding the resulting flow when the mean flow is oblique to the transition as depicted in the top panel of Fig. 2 [inducing a mixture of internal boundary layers and secondary circulations, see Anderson (2020)], or when both Reynolds stress and thermal forcing combine to modulate these secondary flows simultaneously.

3.2 Implications for Measurements

Secondary circulations and IBLs challenge meteorological eddy-covariance flux observations: a fine balance is needed between placing the instrument at a higher elevation to have a larger, more representative footprint and the need to sample the fluxes from the surface of interest underneath the sensor. If the instrument is outside the IEL, advection and non-equilibrium conditions will influence the measurements. Observational sites are often at homogeneous flat terrain with short vegetation, and more significant vegetation or topography upwind can induce the horizontal advection of temperature and other quantities into the observational site (Cuxart et al. 2016) even for disorganized heterogeneity (Mahrt 2017; Mahrt et al. 2018; Simó et al. 2019). Such temperature advection distorts the flux profiles. Moreover, classic tests for stationarity and equilibrium may be misleading. A sensor placed in the IBL but above the IEL might report stationary statistics, but the turbulence would in fact not be in equilibrium when viewed in a Lagrangian framework as a parcel adjusts to the new boundary conditions [hence stationarity and equilibrium are distinct attributes as further underlined in Mahrt and Bou-Zeid (2020)]. A production–dissipation balance might be misleadingly suggested by the measurements, but the tendency, horizontal transport (advection, turbulence and pressure) and vertical turbulent and pressure transport will be significant (Yang et al. 2006; Bou-Zeid et al. 2009) and may also nearly balance each other independently of production and dissipation.

4 Integrating the Effects to Regional Scales

Building on the advances reviewed above in understanding the local circulations and dynamics of the ABL over heterogeneous surfaces, a wide array of studies have aimed to parametrize these microscale physics for atmospheric models with coarser resolutions (Pitman 2003; de Vrese et al. 2016). The primary goal of these schemes is to provide the atmospheric model with the correct fluxes of heat and trace gases, and impose on the ABL flow the correct surface drag. Current schemes still overwhelmingly rely on homogeneous-surface frameworks such as MOST, with modifications. A derived benefit is the potential (not for all schemes) to obtain information about surface properties at scales smaller than the atmospheric model resolution (Li et al. 2013). There has been multiple approaches to formulate such parametrizations (Giorgi and Avissar 1997; de Vrese et al. 2016). One class treats the heterogeneous surface as a homogeneous equivalent surface, and another maintains information on the spatial heterogeneity for use in the scheme. These two classes are briefly covered in the next two subsections, and a brief discussion of novel approaches and open challenges follows.

Characterizing the different patches requires suitable fine-scale observations of the surface. Current land-use and land-cover data are available at about 30-m resolution in many locations (e.g., the National Land Cover Database for the USA: <https://www.usgs.gov/centers/eros/science/national-land-cover-database>), but translating that to actual surface properties may not be always straightforward. This process can be complex and its details are beyond the

scope of this review. An additional challenge is the characterization of subsurface properties and hydrological flows that may significantly impact the surface, inducing heterogeneity even over seemingly homogeneous landscapes (Rihani et al. 2015).

4.1 The Equivalent Surface Approaches: Parameter Aggregation

The most straightforward method to account for heterogeneity is to derive an equivalent homogeneous surface that would produce the same grid-cell-average fluxes to/from the atmosphere. Heterogeneity information is then included implicitly, only in a cell-averaged sense. A severe reductionist application of this method is to consider only one type of land use, the most dominant in a given grid cell, completely ignoring unresolved heterogeneity. Seth et al. (1994) and Li et al. (2013), among others, have demonstrated the limits of this approach, particularly in areas where the land is highly heterogeneous at scales smaller than the typical weather- or climate-model grid cell. These include the all-important urban and agricultural areas in which the most-common land use could, for example, represent at little as 30% of the cell (not really a dominant land use). Despite these proven limitations, this ‘dominant approach’ is still widely used; for example, it is at present the default scheme in the Weather Research and Forecasting model when used with the popular Noah land-surface model (Wang et al. 2019).

The ‘effective parameters approach’ aims to represent the heterogeneous land surface as an equivalent homogeneous surface with properties (roughness, temperature, etc.) that would result in the correct regional-averaged fluxes when coupled to an atmospheric model. It is thus sometimes called the parameter aggregation approach (Claussen 1995). An important question is then how to obtain such properties. Let us consider the computation of the kinematic heat flux ($H = \overline{w'\theta'}$) as an example, where w' is the turbulent vertical velocity perturbation and θ' the turbulent temperature perturbation. MOST remains the dominant model for these schemes whence,

$$H = \frac{1}{\kappa} u_* (\bar{\theta}_s - \bar{\theta}) \left[\ln(z_1) - \ln(z_{0s}) - \psi_h \left(\frac{z}{L} \right) \right]^{-1}, \quad (2)$$

where u_* is the friction velocity, κ is the von Kármán constant, the displacement height is assumed negligible compared to the elevation of the first grid cell, z_1 ; $\bar{\theta}_s$ is the ‘aerodynamic’ surface temperature, $\bar{\theta}$ is the air temperature, z_{0s} is the thermal roughness length, and ψ_h is the the stability function for heat. The overbar denotes Reynolds averaging, while angle brackets denote grid-cell averaging. If one aims to compute $\langle H \rangle$, ‘effective’ values for u_* , θ_s , z_{0s} , and ψ_h are needed (if atmospheric heterogeneity is taken into account, an effective θ is also needed). If these parameters all lack spatial correlation to each other, the answer would simply be $\langle u_* \rangle$, $\langle \theta_s \rangle$, $\langle \ln(z_{0s}) \rangle$, and $\langle \psi_h \left(\frac{z}{L} \right) \rangle$. The stability correction $\langle \psi_h \rangle$ then depends on its functional form and the grid-averaged value of $\langle L \rangle$. However, these parameters are in fact correlated in space. This is why efforts to determine an effective roughness length (André and Blondin 1986), for example, often note that its value depends on flow characteristics such as stratification (Zilitinkevich et al. 2008) and wind direction, and on the typical size of the patches (Bou-Zeid et al. 2004, 2007). Moreover, the edges of tall vegetation or of sea-ice floes may significantly increase the area-averaged stress even when occupying only a small surface fraction (Lüpkes et al. 2012, 2013).

The primary approach for investigating how these spatial correlations influence the determination of effective surface parameters is through fine-scale modelling that resolves this heterogeneity. A full examination of these studies is beyond the limits of our review, but

see the many studies that have adopted this approach (Claussen 1990, 1991; Bou-Zeid et al. 2007; Bertoldi et al. 2007, 2008; Miller and Stoll 2013; Kröniger et al. 2019).

A related method is often called the ‘statistical approach’; it uses the probability density functions (p.d.f.s) of surface properties that are relevant in modelling land–atmosphere fluxes, and integrates these fluxes (e.g., Eq. 2) or the whole SEB across the p.d.f.s. This approach may allow, with some simplifications, a theoretical derivation of some of the needed equivalent surface properties (Avissar 1992; Giorgi and Avissar 1997).

An ongoing challenge for this class of schemes is the development of generalized methods for aggregating the surface properties for all patch scales and for all covariances of surface parameters. This is not a trivial goal particularly for unstructured heterogeneity patterns, where individual patches may not even be easy to delineate, and where some land uses may not be uniformly distributed over the area. For such surfaces, some of the open questions include: can well-posed effective parameters be backed out from spatially-averaged observed fluxes? Can general theories or models be developed for the needed parametrizations? Will that lead to significantly-improved performance of the bulk formulae?

4.2 The Multi-surfaces Approaches: Flux Aggregation

Other approaches such as tiling or mosaic (the definitions are not uniform across the literature) aim to model the interaction of each surface type or patch with the atmosphere separately, and then to aggregate (sum) the fluxes from each patch into the ABL (Avissar and Pielke 1989). The primary question then is whether to assume the ABL is horizontally homogeneous over the whole grid cell or not. If all patches ‘see’ the same horizontally-homogeneous air layer aloft, the surface fluxes from each patch can be computed (via MOST for example) using the cell-averaged atmospheric conditions that the atmospheric model provides. This is an acceptable assumption when the turbulence intensity and mixing are strong and the patches are small, yielding a blending height below the first grid cell level (Fig. 3—left panel). One can then group all patches of similar land use into a single clump and model the surface of, and fluxes from, each clump as it interacts with the ABL (e.g., Li et al. 2013). If this assumption is made, there is no longer a need to account for the spatial distribution of the patches.

Another possibility is to account for atmospheric heterogeneity. This is particularly important if the height of the first grid level that interacts with the surface is below the blending height (Fig. 3—right panel), which occurs for relatively large patches or high vertical-resolution models. This first grid level is then inside the IBL of each surface or inside secondary circulations that are generated by surface heterogeneity. Intermediate nodes that lie inside the IEL then need to be added for each patch to represent the air properties right above that patch. With a heterogeneous atmosphere, one can also run the surface model on a finer grid (since it is cheaper computationally than the atmospheric solver), accounting for the location of each patch (Seth et al. 1994). The availability of spatial information then allows some accounting for IBLs, blending heights, and secondary circulations. Alternatively, the spatial patch distribution and associated atmospheric heterogeneity can be captured more simply using some characteristic (or effective) heterogeneity scale with grouping of similar land-use patches. The clumping is here justified by the assumption that the intermediate nodes above similar patches must have the same properties, if they lie inside the supposedly similar IELs.

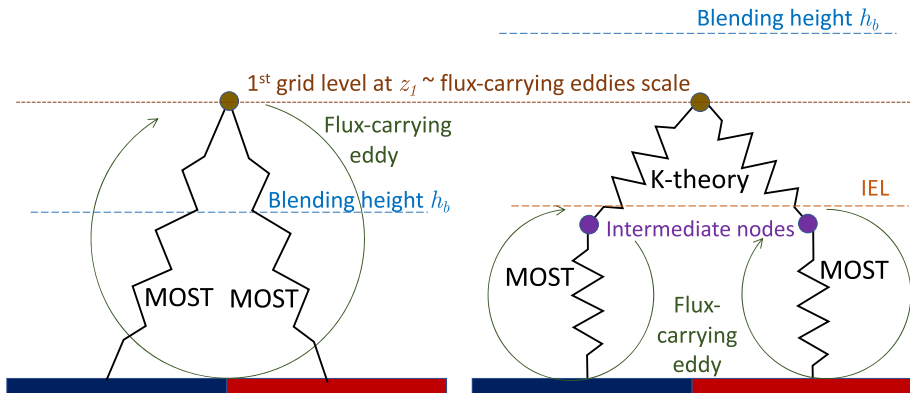


Fig. 3 In the left panel, the blending height h_b is assumed to be below the atmospheric model's first grid node z_1 . Since the surface model should be designed to connect that node to the surface, it would mainly need to represent the attached eddy scales that dominate this interaction that are then $\sim z_1$. These eddies are sampling blended air above h_b and sweeping it towards the surface. Each patch thus interacts with this blended homogeneous ABL, and this can then be directly parametrized via a similarity theory such as MOST. In the right panel, $h_b > z_1$, and the eddies connecting the surface to the first grid node are not blended. The statistics of the attached eddies above the various surfaces are different, precluding the direct use of MOST. One can then account for this atmospheric heterogeneity by adding intermediate nodes that conceptually are inside the IEL of each patch (algorithmically, these are best handled by the surface, not the atmospheric solver). MOST (or equivalent theories) can then be used to connect these intermediate nodes to the surface, and these nodes are subsequently connected to the first grid node via some turbulence closure such as the K-theory. Since $h_b \propto \text{mean patch scale } \langle L_p \rangle$, the need to account for atmospheric heterogeneity becomes more likely as $\langle L_p \rangle / z_1$ increases (either the patches are larger or the model resolution is higher)

4.3 Non-equilibrium and Non-linear Effects

An assumption in all of the schemes above is that MOST, or some comparable scheme, can be applied over each individual patch. This, by design, ignores potential non-equilibrium conditions between the atmosphere and underlying surface. MOST postulates such an equilibrium (i.e., stationarity, planar homogeneity, lack of subsidence, and high Reynolds number), and as such its use is better justified in set-ups akin to the right panel of Fig. 3, than in those akin to the left panel. However, for multi-surfaces approaches with intermediate nodes, one also needs a closure to connect these intermediate nodes to the actual first node in the atmospheric model, and that is again a challenge given the complex flow patterns that are almost certainly generating non-equilibrium turbulence at the microscale. Even if one allows atmospheric heterogeneity in a flux-aggregation approach, the coarse atmospheric model cannot predict the non-equilibrium turbulence generated by subgrid-scale circulations and IBLs, and how they may modulate the ABL. A subgrid-scale parametrization would then be needed to account for the unresolved secondary circulations and advection. An illustration of this need can be made for stable ABLs where subgrid-scale instabilities and variability can increase turbulence intensity, and increase or decrease surface fluxes, beyond what the mean atmospheric and surface states imply (Stoll and Porté-Agel 2009; Mironov and Sullivan 2016).

Even under equilibrium conditions, the exchanges between the ABL and surface are non-linear, as illustrated by Eq. 2. This implies that the correlations of all surface and atmospheric parameters are relevant in theory (Giorgi and Avissar 1997), but maybe not in practice (Bertoldi et al. 2008). When do non-linear effects and feedbacks dominate and invalidate the

extrapolation of classic homogeneous theories to heterogeneous surfaces? When and where is non-equilibrium a first-order driver of the dynamics?

Many open questions thus persist on how to represent unresolved heterogeneity in coarse atmospheric models. Some recent studies have attempted to propose new approaches outside of the two paradigms reviewed above. For example, Kröniger et al. (2019) used cospectral budgets that maintain key features of the modes of variability over heterogeneous terrain. Big data and machine learning are also potential tools that may be useful in the future to classify and probe heterogeneity patterns and their atmospheric signatures (Chaney et al. 2018).

5 Summary and Recommendations

Land-surface heterogeneity continues to be an open challenge for understanding and predicting microscale to synoptic-scale atmospheric dynamics, particularly in regions of the world (agricultural, urban, polar, etc.) where climatic projection and meteorological forecasting are most critical for human activities and well-being. Our review summarizes the current state of knowledge and open questions on the topic. Land-surface heterogeneity can be qualitatively classified into four types: (I) semi-infinite interfaces such as the land–sea boundaries or the edges of large forests, (II) statistically-homogeneous patches of land such as the mosaic of different crops in agricultural areas, the properties of which are spatially homogeneous when averaged at scales ≥ 10 km, (III) large isolated patches with different land uses than their surroundings, such as cities, deforested patches, lakes, islands, wind farms, or polar polynyas, and (IV) the rest, which here we call unstructured heterogeneity, consisting of a mixture of multiple types with irregular patches that may not even be easy to delineate. This last type remains understudied, and a formal approach to understand the complex flow patterns over such surfaces and their regionally-averaged characteristics is critically lagging.

Heterogeneity has implications at the microscales (~ 1 to 10,000 m, the scales of turbulent structures and secondary circulations in the ABL). Advection and non-equilibrium become dominant processes and features. If the flow is perpendicular to the interface of change in surface properties, internal equilibrium and boundary layers develop. If the flow is parallel to the interface, secondary circulations driven by either the heterogeneity of surface stress or buoyancy flux develop (and near the surface they also induce internal equilibrium and boundary layers). Flows that are oblique to the change in surface properties are seldom studied, and may not be simple intermediates between the two parallel and perpendicular end regimes. Understanding how the flow evolves near such interfaces remains important for a wide range of applications such as modelling energy replenishment in wind farms, evaporation from reservoirs, circulations and ventilation of cities, to name a few. Modern computational tools (LES and high resolution mesoscale modelling) have significantly advanced the understanding of the dynamics, but they too have uncertainties in their coupling between the surface and the overlying airflow (wall-models and subgrid-scale processes near the ground), that are beyond the scope of this review.

Field experiments relying on single tower measurements do not provide the needed resolution alone, but if supplemented with aircraft measurements or scanning lidars, the combined observational platforms do allow for detailed analyses of the flow evolution near surface transitions. The expanding availability of lidars and the use of novel small drones for atmospheric sensing, that can fly closer to the surface (compared to large and expensive instrumented manned planes), will continue to improve observations of the heterogeneous ABL above variable land surfaces (Reuder et al. 2012; Higgins et al. 2013; Bonin et al. 2013; Elston et al.

2015; Cuxart et al. 2019; Båserud et al. 2020). New distributed sensing systems can also provide fine resolution measurements in space and time for temperature and sometimes velocity in the ABL (Thomas et al. 2012; Zeeman et al. 2015; Pfister et al. 2019; Izett et al. 2019). Particular attention is needed for unstructured heterogeneity, transitions that are oblique to mean wind directions, and the interacting effect of transitions in multiple surface properties. Another area where our understanding remains incomplete are the limits of strongly stable or strongly unstable ABLs; that is, regimes with weak mean wind and synoptic forcing where concepts such as the internal boundary layer and blending height may not be suitable to examine the flow and turbulence.

At regional scales, an open challenge is to parametrize the heterogeneity and dynamics that cannot be resolved by the coarse models (currently, up to 10 km or so for weather models and up to 100 km or so for climate models). With the increasing resolution of these models, the landscape of this research domain is shifting. Geophysical models of the atmosphere used to be divided into three classes: global (climate), meso (weather) and micro (ABL) scales. It seems, however, that in the foreseeable future, these will coalesce into two groups: meso to global models (see Zhou et al. 2019) and micro to synoptic models (see Schalkwijk et al. (2015) who did a year-long LES forecast of weather over in the Netherlands). This rapidly-changing resolution makes parametrization of heterogeneity a moving target, because we will continually improve our ability to explicitly resolve the patches. Some of the questions we need to answer, however, are scale independent. For example, at whatever scale one is simulating the land–atmosphere coupling, the use of equilibrium frameworks such as MOST can be theoretically flawed. In fact, the problem may become more acute at higher resolutions where advection from adjacent cells is important, compared to lower resolutions where the exchanges are averaged over large areas. We then may need a new non-local framework designed for heterogeneous areas that can explicitly account for advection and for the conditions upwind of a given model grid cell. One line of reasoning would be to define a dimensionless quantity formed by the ratio of an advection time scale to a local equilibrium time scale. The advection time scale includes the mean velocity at the first grid level, $U(z_1)$, as well as the characteristic patch scale (L_p). The local equilibration time scale can be thought of as the delay needed for eddies to become in equilibrium with the local gradient ($\kappa z_1/u_*$). When $(L_p/U) < \kappa z_1/u_*$, local equilibrium will be distorted by advection and needs to be accounted for. The direction of this distortion and how to correct for it may be compared to the terra-incognita problem postulated for parameterizing turbulence in the ABL at intermediate model resolution (Wyngaard 2004). For surface heterogeneity, if the model is very coarse (~ 10 km or more grid resolution horizontally) we need to parametrize the bulk of the subgrid-scale heterogeneous structures in a statistical sense and we have experience doing that. If the model is very fine (~ 10 m resolution LES), the first grid points are predominantly inside their respective IELs, and we can use MOST confidently. But how do we proceed at intermediate scales?

As emphasized throughout this review, the ABL response to landscape heterogeneity has far-reaching implications for biodiversity, hydrology, air quality, renewable energy, and climate projections. These problems are inextricably coupled to climate change and sustainability, which are contemporary themes of great societal and scientific significance. Since its inception, *Boundary-Layer Meteorology* has figured prominently in dissemination of progress in this area. Given the multitude of open challenges, and the increasingly important implications for the scientific challenges of future decades, the journal will remain a vitally-important avenue for scholars studying the lower atmosphere.

Acknowledgements We would like to thank an anonymous reviewer whose comments helped the authors greatly in improving this review. E.B.Z. was supported by the Andlinger Center for Energy and the Environment at Princeton University and the Physical and Dynamic Meteorology Program of the National Science Foundation under AGS-1026636. G.K. acknowledges support from NSF-AGS-1644382 and NSF-IOS-1754893. Larry Mahrt was supported by Grant 1945587 from the National Science Foundation.

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