THE EFFECTS OF MESOSCALE SURFACE HETEROGENEITY ON THE
FAIR-WEATHER CONVECTIVE ATMOSPHERIC BOUNDARY LAYER

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Song-Lak Kang

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The thesis of Song-Lak Kang was reviewed and approved* by the following:

Kenneth J. Davis
Associate Professor of Meteorology
Thesis Advisor
Chair of Committee

John C. Wyngaard
Professor of Meteorology, Mechanical Engineering,
and GeoEnvironmental Engineering

Nelson L. Seaman
Associate Professor of Meteorology

David R. Stauffer
Associate Professor of Meteorology

Yvette P. Richardson
Assistant Professor of Meteorology

Andrew M. Carleton
Professor of Geography

William H. Brune
Professor of Meteorology
Head of the Department of Meteorology

*Signatures are on file in the Graduate School
ABSTRACT

New insight into the structure of the fair-weather daytime ABL (about 1130-1430 LST) over heterogeneous surfaces at a 10-km-order scale is obtained from observational and numerical studies of data collected by aircraft and at surface flux sites during the International H2O Project (IHOP_2002) and using a recently-developed large eddy simulation (LES). From these studies, the key findings to the ABL structure over the mesoscale surface heterogeneity are the following.

First, the ABL over mesoscale surface heterogeneity has a considerably different structure, depending on the heterogeneity intensity. In the ABL with a low amplitude of surface heat flux variation, the microscale (turbulent) variances and fluxes, after filtering out the mesoscale contribution, can fit to the values predicted by the mixed-layer similarity which is built based on the assumption of a homogenous ABL. However, the ABL with a high variation amplitude can contain temporally oscillating mesoscale horizontal flows and thus does not satisfy the quasi-steady state condition.

Second, even in the ABL over mesoscale surface heterogeneity, the microscale (turbulent) vertical flux is found to be much more significant than the mesoscale vertical flux, the product of mesoscale fluctuations of vertical velocity and a scalar (potential temperature or water vapor mixing ratio), due to the absence of mesoscale fluctuation of vertical velocity. However, as the amplitude of surface heat flux variation increases, the interscale component of the vertical flux, the product of microscale (turbulent) vertical velocity and mesoscale fluctuation of potential temperature or water vapor mixing ratio,
becomes significant and can be comparable to the microscale (turbulent) vertical flux at certain times only in the non-quasi-steady state ABL.

Third, the advection of moisture by the surface-heterogeneity-induced mesoscale horizontal flows can be significant from a perspective of the environment for moist convection. Even for the ABL with a low amplitude (here 100 Wm$^{-2}$ or lower) of surface heat flux variation, the amount of moisture transported from the cooler region (the region over the surface having heat flux below the domain average) to the warmer region (the region over the surface having heat flux above the domain average) is as much as 130% of the amount of moisture supplied from the surface in the warmer region.

**Key words:** daytime atmospheric boundary layer, mesoscale surface heterogeneity, mixed-layer similarity, absence of mesoscale vertical velocity, mesoscale vertical flux, advection of moisture
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Chapter 1

Introduction

1.1 Background

1.1.1 Homogeneous versus heterogeneous ABL

The portion of the troposphere directly influenced by the earth’s surface is defined as the atmospheric boundary layer (ABL) (Stull, 1988; Sorjan, 1989; Arya, 2001). Although the ABL is a thin layer of just a few kilometers, it has a tremendous influence on our everyday lives directly and indirectly and is therefore of great interest. First, the ABL itself is an essential element for diagnosing and predicting the local environment. Environmentalists, for example, have been interested in the wind and turbulence fields within the ABL, which have a profound influence on the transport and dispersal of pollutants (Wyngaard, 1988). Accurate forecasts of low-level ABL conditions, such as cloud ceiling and visibility, are crucial to air traffic safety and operational efficiency (Bergot et al., 2005). Second, the ABL is an indispensable component of weather and climate systems, which are commonly emulated by numerical models for academic and/or operational purposes. Numerous studies have suggested that the ABL parameterization is one of the most sensitive factors that are particularly related to
precipitation physics in global circulation models (GCM), as well as in mesoscale modes
(MM) (e.g., Wisse and Arellano, 2004; Byun and Hong, 2004; Martin et al., 2000; Basu
et al., 2002; Hong and Pan, 1996). Accordingly, the accurate and efficient representation
of the ABL, both in an MM and a GCM, is one of the most rapidly growing research
topics in the numerical modeling community.

Micrometeorologists have largely described the homogeneous ABL structure through
application of the similarity theories. For example, Monin-Obukhov similarity (e.g.,
Monin and Obukhov, 1954; Wyngaard, 1973) has been applied for the surface layer, and
the mixed-layer similarity (e.g., Deardorff, 1972; Deardorff et al., 1980) has been used to
describe mixed layers that are in a state of free convection. Thus, atmospheric numerical
models have employed ABL parameterizations based on the similarity theories. The
reason why micrometeorologists apply similarity theories, instead of direct solution of the
exact governing equations for the ABL structure, is the complexity of turbulence, which
dominates the flow in the ABL. With the horizontally homogeneous ABL assumption,
these similarity theories have been constructed from universal relationships between
meteorological parameters, which have been empirically determined based on the results
of field experiments and numerical simulations. However, the condition of horizontal
homogeneity is satisfied only for some exceptional cases in the real atmosphere. There
are various causes that disrupt the homogeneous ABL. One of the primary factors is the
earth’s surface, which is always heterogeneous. Thus, it is of importance to investigate
the modification of ABL properties, which have been understood based on the
assumption of the homogeneous ABL, by heterogeneous land surfaces.
1.1.2 Heterogeneous surface

Many causes of surface heterogeneity that create heterogeneous structure in the ABL can be largely categorized based on two factors: mechanical factors such as surface roughness and elevation, and thermal-hydrological factors such as surface temperature and wetness. In fact, the mechanical and thermal-hydrological factors occur together and, due to the nonlinearity of the atmosphere system, the individual effects may not be additive (Arya, 2001). However, to avoid the complexity of ABL structures which the combined mechanical and thermal-hydrological factors would create, the effects of these factors have been studied separately (Arya, 2001).

Although the effects of mechanical factors can be of more significance in the stable ABL (e.g., Holden et al., 2000; Mason, 1987), the effects of thermal-hydrological factors may be dominant in the convective ABL. An example of weather phenomena associated with thermal-hydrological factors is the sea breeze, which is generated by the extreme difference of surface heat fluxes between land and ocean. Although the phenomena of sea breeze has been observed and studied since ancient times (Neumann, 1973), research on the effect of land surface heterogeneity related to the thermal-hydrological factors has rapidly grown in recent years (for a comprehensive literature review, see Pielke, 2001) compared with those related to mechanical factors.

The increasing interest in the impact of human modifications on the weather and climate system has made the topic of the effect of surface heterogeneity related to thermal-hydrological factors particularly popular recently (e.g., Weaver and Avissar, 2001). Thermal-hydrological factors are linked to the partition of surface energy fluxes...
into sensible and latent heat fluxes. This partition is a function of the surface properties, which can be affected by human activity. In other words, modulation in the earth’s surface can result in significant changes in the surface energy budget. For example, a reduction of the sensible heat flux can be caused by the conversion of land to agriculture (e.g., Lyons et al., 1996) or the leafing out of vegetation (e.g., Fitzjarrald et al., 2001). This change in the surface energy budget is likely to influence the thermodynamic potential for deep cumulus convection (Segal et al., 1995). Also, the alteration of boundary-layer moisture structure by a modified surface moisture budget, which may also result from a change of surface land use, can connect to the development of moist convection (e.g., Findell and Eltahir, 2003).

1.2 Review of previous studies

The effect of thermal-hydrological surface heterogeneity on ABL structure has been studied using theoretical, observational, and numerical approaches. In fact, many of the previous studies have used a numerical model, an MM or an LES. Most studies on the effect of surface heterogeneity at a 100-km-order scale have utilized an MM whereas most studies on the effect of surface heterogeneity at a 10-km-order scale have utilized an LES. This section surveys previous studies from the perspective of two different scales of surface heterogeneity: one at a 100-km-order scale, and the other at a 10-km-order scale.
1.2.1 At a 100-km-order scale

Some theoretical studies (e.g., Dalu and Pielke, 1993) suggested that the local Rossby radius of deformation is the optimal scale of surface heterogeneity to generate sea-breeze-like mesoscale circulations. The Rossby radius is estimated to be about 100 km with typical atmospheric parameters (Pielke, 2001). Considering the typical effective resolution of a GCM, the effect of surface heterogeneity at a 100-km-order scale is a subgrid problem in a GCM. Thus, as a subgrid problem, the effect of surface heterogeneity at a 100-km-order scale has been investigated by using a mesoscale model (MM).

These MM-based studies (e.g., Segal et al., 1988; Avissar and Chen, 1993; Chen and Avissar, 1994a, b; Avissar and Liu, 1996) have demonstrated the generation of sea-breeze-like circulations by surface heat flux variation. They concluded that the mesoscale vertical flux, the product of mesoscale fluctuations of $w$ and a scalar, is often greater than the turbulent flux (e.g., Chen and Avissar, 1994a). Some studies (e.g., Weaver and Avissar, 2001) further suggested that the negligence of the mesoscale fluxes, which are thought by some scientists to be important for cloud formation and precipitation (e.g., Chen and Avissar, 1994b), may cause a serious failure in a GCM used to predict climate change. Thus, some scientists (e.g., Liu et al., 1999; Arola, 1999; Lynn et al., 1995) have already attempted to parameterize the mesoscale vertical fluxes in a GCM to account for the effect of surface heterogeneity on a subgrid scale. Most MM-based studies, however, have prescribed idealized surface flux conditions having extreme contrast, such as on the boundary of land/sea, irrigation/non-irrigation, urban/rural, or bare/vegetated ground.
Zhong and Doran (1997, 1998) also performed numerical studies with an MM, which provided different conclusions. They argued that, with variation at a 100-km-order scale, the extreme contrast used in previous MM-based studies is rare in real surface heat flux distribution. Thus, they utilized observation-based surface flux distributions instead of idealized conditions. Zhong and Doran (1997, 1998) conclude that the magnitude of mesoscale vertical velocities generated by convergence and divergence is overestimated in previous research where idealized surface flux conditions are used. Thus, they suggest that the mesoscale vertical flux is not as significant as asserted in previous numerical model studies.

1.2.2 At a 10-km-order scale

The ABL structure over surface heterogeneity at a 10-km-order scale has been addressed by some observational studies (e.g., Mahrt et al., 1994a, b; LeMone et al., 2002, 2006; Kang et al., 2006), although there have been just a few such studies due to the difficulty in obtaining appropriate observations. Unlike the conclusions suggested from some MM-based studies (e.g., Chen and Avissar, 1996 a, b), the observational studies presented very weak mesoscale vertical velocity, which is associated with convergence or divergence. LeMone et al. (2002) analyzed data from aircraft and radar wind profilers. They observed a whole ABL circulation, which is described by near-surface convergence and upper-level divergence over a warm surface, and near-surface divergence over a cool surface. Although the observed mesoscale flows seem to be organized, the vertical velocity computed from the observed convergence or divergence
is 1.5 cm$^{-1}$ at the middle of the ABL and 0.4 cm$^{-1}$ in the upper ABL, which are one or two orders smaller than the values obtained from the MM-based studies with idealized surface flux conditions (e.g., Chen and Avissar, 1994a, b). Mahrt et al. (1994b), analyzing data from repeated low level aircraft runs over alternating regions of irrigated and nonirrigated surfaces on the length scale of 30 km, concluded that the mesoscale modulation of the turbulent flux is more important than the direct mesoscale flux. At least on the scale of 10-30 km, the observed ABL structure over a heterogeneous surface is different from what is expected from the MM-based studies where idealized surface flux conditions are used. From a perspective of the significance of mesoscale fluxes, the observed ABL structure is close to what is described by Zhong and Doran (1997, 1998).

However, due to the limitations of observational data, these observational studies could not present in detail the ABL structure and associated atmospheric processes over a heterogeneous surface. In order for the effect of surface heterogeneity at a 10-km-order scale to be fully examined, LES can be an appropriate tool. From the perspective of an MM, which has an effective resolution of the order of 10 km, the effect of surface heterogeneity at this scale should be treated as a subgrid problem. However LES can explicitly resolve energy-containing eddies in the convective ABL that would be parameterized in an MM. Thus, the turbulence-related issues, which cannot be investigated by an MM, such as mesoscale modulation of the turbulent flux and variance (Kang et al., 2006; Mahrt et al., 1994; LeMone, 1984), can be studied with LES. Although LES is an appropriate tool to examine the effect of surface heterogeneity at a 10-km-order scale, most previous LES-based studies have focused on the effect on the ABL due to surface flux variations at a scale similar to the ABL depth (on the order of 1
km) (e.g., Hadfield et al., 1991, 1992; Shen and Leclerc, 1995; Raasch and Harbusch, 2001; Kim et al., 2004). Some recent LES-based studies (e.g., Avissar and Schmidt, 1998; Patton et al., 2005) have included numerical experiments with surface forcing at a 10-km-order scale. However, these LES-based studies focused on finding the scales of surface heterogeneity that are optimal for generating sea-breeze-like circulations, instead of investigating the ABL structure, which is influenced by surface heterogeneity at a 10-km-order scale, and associated atmospheric processes.

The LES-based studies, including the experiments with mesoscale forcing observed a well-organized mesoscale flow pattern existing in a space and time averaged flow field over a heterogeneous surface at a 10-km-order scale. Specifically the scale is 20-40 km in Avissar and Schmidt (1998) and 5-15 km in Patton et al. (2005). Letzel and Raasch (2003), with a numerical experiment design similar to Avissar and Schmidt (1998), reported that mesoscale flows can undergo a temporal oscillation. In contrast, Patton et al. (2005), which indirectly prescribe surface heterogeneity with much less contrast variation than that in Avissar and Schmidt (1998), found no temporal oscillations in the mesoscale flows. Although the suggested optimal scales and the organized flow patterns differ, both Avissar and Schmidt (1998) and Patton et al. suggest that, outside the optimal scale range, surface-heterogeneity-induced mesoscale flows coexist with persistent turbulent flows.
1.2.3 Summary and critique

Once a sea-breeze-like circulation is generated by extremely contrasted surface conditions, the effect of the mesoscale circulation would be significant. Thus, the impact on the ABL structure should be considered both in a GCM and in an MM. Based on theoretical studies and MM-based studies with idealized surface conditions, it has been demonstrated that the surface heterogeneity at a 100-km-order scale can easily generate the sea-breeze-like circulation. However, some MM-based studies using realistic surface conditions concluded that the extremely contrasted surface heterogeneity at a 100-km-order scale are occasional and the parameterization of the surface heterogeneity effect in a GCM does not need immediate attention.

Compared with MM-based studies, LES-based studies on the effect of surface heterogeneity has a short history. In addition, most previous LES-based studies have focused on the effect of surface heterogeneity on the 1-km-order scale or finding the optimal scale range of surface heterogeneity to generate sea-breeze-like circulations. Based on extensive LES experiments, Avissar and Schmidt (1998) suggested that surface heterogeneity at a length scale larger than 5-10 km can generate surface-heterogeneity-induced mesoscale circulations. Since then, although the details are somewhat different, some LES-based studies (e.g., Letzel and Raasch, 2003; Patton et al, 2005) agreed that the impact of the surface heterogeneity at a scale larger than 5-10 km may be significant.

Recently, Baidya Roy et al.(2003) suggested that the atmosphere is most responsive to surface forcing at a scale in the 10-20 km scale range, irrespective of the domain scale of the surface heterogeneity. They obtained this result by running a high-
resolution MM with multiple-scale surface forcing based on observations. In addition, Zhong and Doran (1998), disagreeing with the significance of the effect of surface heterogeneity at a 100-km-order scale, also assented to the significant response of the ABL to the surface heterogeneity at a 10-km-order scale. Although some of the significant effects have been suggested by previous studies, the ABL structure and associated processes over surface heterogeneity at a 10-km-order scale have not been studied extensively.

For the heterogeneous structure of the ABL to be adequately characterized in a larger model (an MM or a GCM), the ABLs containing well-organized mesoscale flows or less-organized mesoscale flows (coexisting with turbulent eddies) must be systematically described and illustrated with respect to the features of heterogeneous surfaces. However, even the term “well-organized mesoscale circulation” (WOMC), which is commonly used in the literature (e.g., Avissar and Schmidt, 1998; Patton et al., 2005), has not been well defined. The term WOMC has often been used to refer to a sea-breeze-like mesoscale flow pattern existing in a space and time averaged flow field over a heterogeneous surface. Patton et al. (2005) used the term WOMC to describe the stationary organized flow generated by a heterogeneous surface. However, the mesoscale flows demonstrated by Avissar and Schmidt (1998), which they labeled a WOMC, has been revealed by Letzel and Rassch (2003) to be temporally oscillatory flows. Furthermore, compared with ABLs containing WOMC, the ABLs containing less-organized mesoscale circulations (LOMC) (coexisting with turbulent eddies) have been much less studied. Thus, there are numerous questions on the ABL containing LOMC, which need to be answered. One of the questions is about the modulation of turbulence
by mesoscale flows, which has been observed (e.g., LeMone, 1976; Mahrt et al., 1994a; Kang et al., 2006), but has not been investigated in detail.

1.3 Thesis Objectives and Organization

The main goal of this study is to characterize the ABL structures over heterogeneous land surfaces. Specifically, this study would focus on ABL structures influenced by mesoscale surface heterogeneity. The detailed structures of heterogeneous ABLs are to be investigated as a first step to the development of an appropriate treatment of surface heterogeneity effects in a larger scale model (an MM or a GCM), especially in the context of surface-ABL-moist convection interactions.

The main goal is to be achieved by:

- Observed structures of fair-weather convective ABLs over a surface heterogeneity at a 10-km-order scale are to be investigated. As a function of the characteristics of surface heterogeneity and background weather condition, the effect of surface heterogeneity on the ABL structure is to be described.
- Fair-weather convective ABLs simulated with various intensities of surface heterogeneity at a 10-km-order scale are to be used to investigate the structures of the ABLs and their associated processes. Furthermore, the modifications of the ABL structure over the surface heterogeneity are to be explored from a perspective of the environment for moist convection
This dissertation research primarily attempts to answer the following key questions:

- Do mixed-layer similarity relationships, which are constructed based on the assumption of quasi-steady and horizontally homogeneous ABL, still work in the ABL over the mesoscale surface heterogeneity? What is the range of heterogeneity intensity in the ABL over mesoscale surface heterogeneity, in which mixed-layer similarity still works?

- Does the mesoscale vertical flux, the product of mesoscale fluctuations of vertical velocity and a scalar (Chen and Avisar, 1994a), become comparable to or more significant than the turbulent flux? If not, is there any other component in the decomposed vertical flux that becomes more significant than the turbulent flux in the ABL over a heterogeneous surface?

- Depending on the heterogeneity intensity, does the ABL over the surface heterogeneity at a 10-km-order scale have a somewhat different structure?

Chapter 2 documents the influence of surface heterogeneity at a 10-km-order scale on ABL structure under different background weather conditions using observational datasets collected during the International H2O Project (IHOP_2002), in which multiple research components utilized various enhanced observational instruments. In addition to answers to the general questions throughout this dissertation, this chapter, specifically, includes the answers to the following questions:

- What is the primary cause of mesoscale surface heterogeneity over the study area?
• Under what kind of background weather conditions can the impact of surface heterogeneity be significant to the whole ABL structure?

Chapter 3 investigates the influence of mesoscale surface heterogeneity on simulated dry ABLs as a function of heterogeneity intensity (the amplitude of surface heat flux variation). Dry ABLs are simulated using the Bryan-Fritsch model (Bryan and Fritsch, 2002), a compressible non-hydrostatic model developed as a cloud-resolving model using LES techniques. The specific questions to be answered in this chapter are:

• What is the principal mechanism governing horizontal flows generated by mesoscale surface heterogeneity?

• Is the surface-heterogeneity-induced mesoscale flow quasi-stationary or temporally-oscillating? In other words, can the assumption of the quasi-steady ABL still be applied in the ABL over the mesoscale surface heterogeneity?

In Chapter 4, the contribution of mesoscale surface heterogeneity to moist convection initiation will be discussed also based on simulated moist ABLs using the Bryan-Fritsch model (Bryan and Fritsch, 2002). From the perspective of modification of the environment for moist convection, the primary question to be answered in this chapter is:

• Can horizontal transport of moisture by the generated mesoscale horizontal flows be significant from a perspective of the potential for moist convection initiation?

Chapter 5 provides a summary of conclusions and offers the future plans of this study.
Chapter 2

Observation

2.1 Chapter Introduction

Predicting the timing and location of convective cloud development is a fundamental challenge in the study of meteorology. Spatial variability in the atmospheric boundary layer (ABL) is closely related to the development of moist convection (e.g., Crook, 1996; Rozoff et al., 2003; among many papers in the literature). The heterogeneity of the ABL is known to be caused by multiple factors including surface forcing and above-ABL conditions. The influence of land-surface heterogeneity on ABL structure and moist convection has been studied by numerous investigators using numerical models and observational data (for comprehensive review, see Pielke, 2001).

Mesoscale numerical model studies of the effects of land-surface heterogeneity on the ABL structure and moist convection have focused on the generation of well-organized circulations with prescribed surface sensible and latent heat fluxes at the meso-β scale, which ranges from 20 to 200 km (e.g., Segal et al. 1989; Chen and Avissar, 1994 a, b; to name a few). Much of the earlier research involved idealized conditions, while Zhong and Doran (1998) used a real-data case and found that heterogeneous conditions result in a very similar situation as that simulated with homogeneous surface conditions. These results suggest that well-organized meso-β scale circulations generated by land-surface heterogeneity may be difficult to find outside of areas of
extreme contrasts, such as on the boundary of irrigated/non-irrigated, urban/rural, or bare/vegetated ground.

In order to fully simulate the effects of surface heterogeneity on ABL structure at the meso-$\gamma$ scale (2-20km), large eddy simulations (LES) have been used (e.g., Shen and Leclerc, 1995; Avissar and Schmidt, 1998; Patton et al., 2005; to name a few). Many LES results have indicated that meso-$\gamma$ scale land-surface heterogeneity can significantly influence the entire ABL structure. Most of the LES results, however, are limited by their use of idealized conditions for surface heat and moisture fluxes, mean winds, temperature and moisture soundings, lateral boundary conditions, and terrain.

There have been some observational studies that address the influence of land-surface heterogeneity at the meso-$\gamma$ scale on ABL structure, including the existence of mesoscale circulations generated by land-surface heterogeneity (e.g., Mahrt et al., 1994a; Shaw and Doran, 2001; LeMone et al., 2002; LeMone et al., 2006). However, compared to the numerical studies, there have been fewer observational studies. This is primarily due to the difficulty in obtaining appropriate observations that describe the ABL heterogeneity in a statistically robust manner. However, improvement in technology is expanding our ability to observe the heterogeneous ABL structure.

The International H2O Project (IHOP_2002; Weckwerth et al., 2004), in which multiple research components utilized various enhanced observational instruments, provided an opportunity to observe the heterogeneous ABL structure. Observational datasets collected in this field experiment can be used to assist the research toward a better understanding of the ABL structure over a heterogeneous land surface.
The goal of this study is to document the influence of meso-\(\gamma\) scale land-surface heterogeneity on the ABL structure under different background weather conditions using observational datasets collected from IHOP_2002. This research includes the investigation of the primary causes of surface heterogeneity over the study area. Datasets used for this research will be delineated in the next section, along with a description of the study area. Section 2.3 will discuss the causes of surface heterogeneity over the study area mainly using measurements at the surface flux sites. Section 2.4 will describe the influence of land-surface heterogeneity on the low-level ABL structure under different background weather conditions by applying statistical and spectral analysis methods to the data from repeated low-level aircraft passes. In section 2.5, the influence of the land-surface heterogeneity on upper-level ABL structure is qualitatively assessed based on normalized variance profiles and observed spatial variations of the ABL depth. Section 2.6 summarizes the results and presents conclusions.

2.2 Data

The IHOP_2002 field experiment was conducted from 13 May to 25 June 2002 in the Southern Great Plains (SGP) of the United States with the aim of improving our understanding of convective precipitation, including the development of heterogeneous boundary layer (BL) structure that might contribute to this understanding (Weckwerth et al., 2004). Numerous mobile observing systems, ground-based observation facilities, and six research aircraft were used in this project.
As part of the IHOP_2002 field experiment, the University of Wyoming King Air (UWKA) flew over three tracks, the western track, the central track, and the eastern track to obtain data for investigating the influence of heterogeneous land surface on daytime ABL structure. (In this study, we will focus on the western track. For more information on the other two tracks, refer to LeMone et al. (2006) and/or Weckwerth et al. (2004)). The UWKA was equipped with sensors that measure the state and motion of the ABL. Detailed information on the instrumentation of the UWKA is available online at http://flights.uwyo.edu/base. Only the aircraft-borne instruments which provide the data used in this study are briefly introduced here.

For the aircraft’s position and motion, outputs from the Honeywell Laseref SM Inertial Reference System (IRS) were corrected based on the Global Positioning System (GPS). For measurements of static pressure, temperature, and surface radiation temperature, a Rosemount 1201, a reverse-flow thermometer, and a downward-looking Heiman KT-19.85 radiometer were used, respectively. Potential temperature was derived from the measurement of static pressure by a Rosemount 1201 and from the measurement of temperature by a reverse-flow thermometer. For water vapor mixing ratio, a Lyman-alpha hygrometer, whose long-term mean is adjusted to that of a chilled-mirror hygrometer, was used. Since the measurement instruments on the aircraft were not colocated, the lags between sensors have been taken into account (LeMone et al., 2003; LeMone et al., 2006). Thus, based on the peaks of the cross correlations between the measured variables, the observations of static pressure, temperature, potential temperature, and water vapor mixing ratio are adjusted to reconcile a 0.08 second lag as compared to the wind observations.
The western track, a 60-km north-south oriented strip from approximately 36.4°N to 37.0°N along 100.6°W (Fig. 1), shows the most heterogeneous surface temperature distribution along with the most sparse vegetation distribution as illustrated in Fig. 9 of Weckwerth et al. (2004). Over this track, the UWKA flew a total of 61 legs on the five boundary layer heterogeneity (BLH) mission days: 19, 20, 25, 29 May, and 7 June 2002. Of the 61 legs, 30 are low-level legs, where the aircraft altitude was approximately 65 m above ground level (AGL). Those repeated low-level passes were done to obtain statistically reliable measurements (see Mann and Lenschow, 1994). The observations were recorded at 25 Hz, which corresponds to a spatial resolution of about three meters, between approximately 17 and 20 UTC (11 and 14 CST). General information regarding the UWKA flights and the ABL characteristics on the five days is summarized in Table 2.1. In addition, to characterize the ABL on each day, the vertical soundings of potential temperature and water vapor mixing ratio from the rawinsonde released over Homestead site (36.55°N, 100.6°W) at about flight start time are plotted in Fig. 2.2.

The surface elevation of the track from US Geological Survey (USGS) global 30 second dataset varies within a range of 150 m is shown in Fig. 1. The lowest elevation region, which is around 36.83°N, is occupied by the Beaver River Floodplain that is about 2-km wide. The vegetation cover over the track is estimated by the normalized difference vegetation index (NDVI; LeMone et al., 2006), which was measured from a downward-looking radiometer during the low-level passes of the UWKA flights (Fig. 1). The NDVI indicates sparse vegetation that became greener with time along the track except for the relatively green vegetated area that covers approximately 5 km of the track in the vicinity of the Beaver River. Thus, both terrain elevation and vegetation cover
likely make up a relatively homogenous surface condition. However, as shown in Fig. 1, the State Soil Geographic (STATSGO) map exhibits widely varying soil types along the track.

The weather over the track on the five BLH days was generally fair. The mean direction of wind at 65 m AGL was usually southerly except on 25 May and the mean wind magnitude varied from 2.5 m s$^{-1}$ to 13.2 m s$^{-1}$ (see Table 2.1). For each low-level leg, the relative importance of buoyant production compared to mechanical production of turbulence is compared through Obukhov length (Stull, 1988). Here, Obukhov length is obtained from leg-averaged values of momentum flux, buoyancy flux, and virtual potential temperature. The range of the Obukhov length calculated for each low-level leg on a given day is shown in Table 2.1. Despite their larger surface heat fluxes, 19 and 20 May have relatively large Obukhov lengths due to the strong mean winds. Conversely, 25 and 29 May have relatively small Obukhov lengths due to the relatively weak mean winds although the surface sensible heat fluxes are smaller than those on the previous two days.

In addition to UWKA, other research instruments were concentrated around the track during IHOP_2002 (for the detailed experimental array, see Fig. 4 of Weckwerth et al., 2004). Among them were three Integrated Surface Flux Facility (ISFF; Chen et al., 2003) sites, and the differential absorption lidar (DIAL; Poberaj et al., 2002) aboard the Deutsche Luft und Raumfahrt (DLR) Falcon.

The ISFF sites 1, 2, and 3 were deployed along the track from south to north in ascending order (Fig. 1). Surface skin temperatures measured by the Everest 4000.4GL
infrared surface temperature sensors at the three ISFF sites are used for comparison with those observed by the UWKA. Tower-based eddy-covariance flux measurements are also compared with airborne eddy-covariance measurements. In addition, the surface energy balance (SEB) components, surface skin temperature, and soil temperature were all measured at the three ISFF sites.

Some segments of the UWKA flights were flown concurrently with north-south transects of the DLR Falcon. From the aerosol backscatter data of the DIAL aboard the DLR Falcon, ABL depths over the track were objectively determined by using the Haar wavelet technique of Davis et al. (2000) at a horizontal resolution of 700 to 1000 m.

2.3 Surface heterogeneity

2.3.1 Surface Skin Temperature

The surface skin temperature at a given location, defined as the temperature at the air-soil interface, depends on the radiation balance, near-surface atmospheric exchange processes, presence of vegetation cover, and thermal properties of the subsurface medium (Arya, 2001). Over surfaces with simple vegetation or bare soil, surface skin temperature can be used as an indicator for thermal heterogeneity of land surface (Mahrt, 2000). Vegetation along the track was sparse, with the soil sometimes visible from the aircraft, consistent with the low NDVI values observed along the track (Fig. 1). Thus, for the five fair-weather days, this surface skin temperature distribution may represent the surface thermal heterogeneity over the aircraft track. Fig. 2.3 shows the surface temperature
distribution, which is a composite of overlapping 4-km means at every 1 km from the radiometric surface temperatures measured by repeated low-level passes of the UWKA on a given day. Surface skin temperatures are also measured at the three ISFF sites and the daily averages covering the aircraft flight hours are shown in Fig. 2.3.

In Fig. 2.3, except on 25 May, the surface temperature distributions, which are consistent with those of surface heat flux and potential temperature at 65 m AGL (see section 2.4.1), are generally characterized by warmer conditions in the north and cooler conditions in the south with the largest horizontal gradient of 12 °C (50 km)\(^{-1}\) on 29 May and the smallest gradient of 5 °C (50 km)\(^{-1}\) on 7 June. The 25 May case differs from the other days by having an area of higher temperatures around 36.64°N with a surface temperature gradient of -6 °C (15 km)\(^{-1}\) to the north and -8 °C (15 km)\(^{-1}\) to the south. The different pattern on 25 May is caused by more heating over the region around site 2 (36.62°N), which is likely associated with relatively small soil heat flux and weak mean winds (see section 2.3.4 and 2.3.5). The high temperature over the entire leg on 25B May is due to the late observation hours, from 1852 to 2021 UTC (Table 2.1). In Fig. 2.3, one can also notice that the surface temperatures at about 36.83°N where the Beaver River is located are 0.5-1.5 °C lower than those in surrounding areas all five BLH days.

2.3.2 Rainfall, Soil Moisture, Surface Skin Temperature, and Soil Temperature

Fig. 2.4 shows the daily mean values of rainfall, volumetric soil moisture, surface skin temperature, and soil temperature at a depth of 5 cm measured at the ISFF sites 1, 2, and 3 from 13 May to 7 June 2002. There were four rainfall events: on 17 May less than
25 mm, on 24 May 3 mm, on 27 May 20-30 mm toward the north and over 80 mm to the south, and on 5 June 15-20 mm. Although the amount of rainfall differs widely, the four rainfall events always noticeably decrease both surface skin temperature and soil temperature. In spite of the variations in the surface environment, the surface and soil temperature at site 1 persistently gave the lowest values of any site, which is reflected also in the aircraft-measured surface radiation temperature (Fig. 2.3). We attempt to explain these persistent cooler conditions at site 1 in sections 2.3.4.

The significance of soil moisture in the interaction between land surface and atmosphere has been emphasized by numerous researchers (e.g. Reen et al., 2006; Chen and Avissar, 1994 a, b, among many papers in the literature). The observed volumetric soil moistures at the three sites indicate that soil hydraulic properties, along with the different characteristics depending on surface cover (e.g., runoff and infiltration), need to be considered to explain the relationship between the spatial variability in rainfall and soil moisture. At site 2, the rainfall of 36 mm (maximum rainfall rate of 10 mm/ 5 min.) on 27 May caused an increase of 29 % in the 5-cm volumetric soil moisture. However, at site 1 the 5-cm volumetric soil moisture increased only 19 % in spite of a total rainfall of 84 mm (maximum rainfall rate of 4.2 mm/ 5 min.). The difference in soil hydraulic properties between site 1 and site 2, a spatial difference of around 17 km, was seen again on 5 June (Fig. 2.4). On this day, the 12-mm rainfall at site 1 (maximum rainfall of 1.4 mm/5 min.) caused almost no change, but the 14-mm rainfall at site 2 (maximum rainfall of 1.3 mm/5 min.) increased the volumetric soil moisture by 6 %.
2.3.3 Sensible and Latent Heat Fluxes

The relationship between aircraft-measured and tower-measured sensible and latent heat fluxes is shown in Fig. 2.5. First, for the UWKA measurement, leg-averaged sensible and latent heat fluxes are computed for each low-level leg,

\[ SH = \langle \rho \rangle \langle C_p \rangle \langle w' \theta'' \rangle \]  \hspace{1cm} (2.1)

\[ LE = \langle \rho \rangle \langle L_v \rangle \langle w' q'' \rangle \]  \hspace{1cm} (2.2)

where \( \langle \rangle \) indicates leg average, \( \rho \) is air density derived from the aircraft measured variables using the ideal gas law, \( C_p \) is specific heat of moist air, and \( L_v \) is latent heat of vaporization computed as a function of temperature. The perturbations of vertical velocity (\( w' \)), potential temperature (\( \theta'' \)), and water vapor mixing ratio (\( q'' \)) are defined from the leg average as:

\[ \phi'' = \phi - \langle \phi \rangle \]  \hspace{1cm} (2.3)

where \( \phi \) represents one of the variables. The daily mean values of sensible and latent heat flux are obtained by averaging the leg averaged values over the repeated low-level aircraft passes on a given day. Second, the sensible and latent heat fluxes measured at each site during the aircraft flight are averaged to estimate the surface fluxes on a given day. The sensible and latent heat fluxes averaged over the three sites are used for the comparison with those of the UWKA. Fig. 2.5 shows that aircraft measurements give smaller values than tower measurements by about 10-20% for both sensible and latent heat fluxes, except for the cases of 29 May and 7 June. On 29 May, the aircraft latent heat
flux is 30 % higher than the tower measurements while the aircraft sensible heat flux is 10-20 % lower than the tower measurement. The sensible heat fluxes measured by an aircraft are expected to be smaller than those measured by a tower due to the aircraft flight height and the linear decrease of sensible heat flux with height to a negative value at the ABL top (Stull, 1988). The latent heat fluxes measured by an aircraft are often higher than the surface fluxes measured by a tower. On June 7, the aircraft-measured latent heat flux exceeds tower-measured latent heat flux by as much as 85 % while the tower-measured sensible heat flux exceeds aircraft-measured sensible heat flux by 30 %. This significant overestimation of aircraft-measured latent heat flux is likely caused by the considerable height dependence of this flux on this day. (see Table 2.1).

2.3.4 Surface Energy Balance

For the three surface flux sites, the diurnal curves of net radiation \( (R_{net}) \), sensible heat flux \( (H) \), latent heat flux \( (LE) \), and heat flux into the soil \( (G_s) \) from 12 to 00 UTC (from 06 to 18 CST) averaged over the five days are shown in Fig. 2.6. In Fig. 2.6a, \( R_{net} \) at site 2 is larger than those at site 1 and 3. The smaller \( R_{net} \) at site 3, compared with the \( R_{net} \) at site 2, is likely caused by more frequent cloudiness based on the observed smaller incoming shortwave radiation during the daytime. However, the smaller \( R_{net} \) at site 1 is likely associated with a relatively large albedo. The lower surface temperature at site 1 (Figs. 2.3 and 2.4) indicates smaller upwelling longwave radiation \( (Q_{\uparrow_{LW}}) \). Thus, based on \( R_{net} \approx Q_{\downarrow} (1 - A) + Q_{\uparrow_{LW}} - Q_{\downarrow_{LW}} \) (Pielke, 2001; Arya, 2001; Stull, 1988) with the
assumption of no significant spatial variability in downwelling longwave radiation \( (Q_{LW}^\downarrow) \) and insolation \( (Q_s) \) between the sites, the smaller \( R_{net} \) at site 1 implies a relatively large albedo \( (A) \) at site 1.

The daytime available energy \( (R_{net}-G_s) \) for the sensible and latent heat fluxes \( (H+L_E) \) is estimated based on the closure of surface energy balance, \( R_{net} - G_s = H + L_E \), which is balanced to within 6% residual in this dataset. In addition to the most net radiation (Fig. 2.6a), the smallest soil heat flux (Fig. 2.6d) creates the most available energy at site 2. Here, we estimate soil thermal conductivity, which determines the soil heat flux, with thermal diffusivity and heat capacity (Arya, 2001; Stull, 1988). First, soil thermal diffusivity is estimated from the theoretical solution of thermal wave propagation in soil (Arya, 2001; Stull, 1988) by using the amplitudes of the changes in soil temperature measured at 5 cm depth and in surface temperature. Second, at each site, soil heat capacity is estimated by using
\[
C_s = \frac{\rho_s}{\rho_m} \times 1.9 \times 10^6 + Q_s \times 4.2 \times 10^6 \text{ (where } Q_s \text{ is volumetric soil moisture)}
\]
with the value of soil bulk density \( (\rho_s) \) at each site and the density of mineral particles \( (\rho_m) \), which are obtained from the ISFF/IHOP_2002 web page (www.atd.ucar.edu/rtf/projects/ihop_2002/isff/). In Table 2.2, the soil thermal conductivity is usually the lowest at site 2, which is of course consistent with the smallest soil heat flux (Fig. 2.6d). Thus, although its soil moisture is always the highest, site 2 is warmer than site 1 (Figs. 2.3 and 2.4). Under relatively weak mean wind on 25 May, this local feature at site 2 is quite manifest in surface temperature distribution (Fig. 2.3). The small thermal conductivity indicates that thermal exchanges are concentrated near the
uppermost surface. This surface, with a shallow active thermal-exchange layer, experiences extreme diurnal temperature fluctuations (Zhang and Huang, 2004).

At site 3, where the soil moisture at 5 cm is the lowest (Fig. 2.4), the dry soil causes most of the available energy to be used for the surface sensible heat flux, which implies the lowest values of the surface latent heat flux (Fig. 2.6c). Thus, with the available energy similar to that at site 1, site 3 shows the highest values of the surface sensible heat flux (Fig. 2.6b) and the surface temperature among the three ISFF sites (Figs. 2.3 and 2.4). On the contrary, site 1, which distributes a significant amount of the available energy to the latent heat flux (Fig. 2.6c), shows the lowest surface heat flux (Fig. 2.6b) and the surface temperature (Figs. 2.3 and 2.4).

2.4 Stationary spatial variability

2.4.1 Decomposition

The stationary spatial variability of the flow and vertical fluxes at 65 m AGL are described by applying the decomposition method of Mahrt et al. (1994b) to the data collected from repeated aircraft passes. First, for a particular observed variable, $\phi$, the $i$th segment average over a fixed-length window in the $j$th leg, $[\phi]_{i,j}$, is computed. From the segment averages over the $j$th leg, the leg average, $\langle \phi \rangle_j$, is obtained. The spatial deviation of the $i$th segment in the $j$th leg is defined from the leg average

$$\phi^s_{i,j} = [\phi]_{i,j} - \langle \phi \rangle_j$$

(2.4)
Here, time-averaging the spatial deviations over all the repeated legs at a fixed location yields the stationary part of the spatial deviation

\[
\{ \phi^s \} = \frac{1}{J} \sum_{j=1}^{J} \phi^s_{i,j}
\]  

(2.5)

where \( \{ \} \) indicates a time average and \( J \) is the number of the repeated aircraft passes. From this stationary part of the spatial deviation, the transient part of the spatial deviation is defined as

\[
\phi^{ST}_{i,j} = \phi^s_{i,j} - \{ \phi^s \}
\]  

(2.6)

Fig. 2.7 shows the stationary spatial deviations, Eq. (2.5), of potential temperature and water vapor mixing ratio at 65 m AGL over the track. Consistent with the spatial variability of surface skin temperature (Fig. 2.3), the potential temperature at 65 m has the strongest gradient on 25A May, 1.2 K (22 km) \(^{-1}\) between 36.57°N and 36.77°N, and the weakest gradient on 7 June, 0.2 K (42 km) \(^{-1}\) between 36.50°N and 36.88°N. On 25A May, the water vapor mixing ratio has a gradient of 0.45 gkg\(^{-1}\) (24 km) \(^{-1}\) between 36.60°N around the warmer region, and 36.82°N around the cooler region, which creates warm-dry and cool-moist conditions. Similarly on 29 May, the potential temperature increases at a rate of 1.1 K (48 km) \(^{-1}\) from south to north whereas the water vapor mixing ratio decreases at a rate of 2.0 gkg\(^{-1}\) (48 km) \(^{-1}\), which also creates warm-dry and cool-moist conditions.

Fig. 2.8 exhibits the stationary spatial deviations of \( v \) and \( u \) winds. Because the mean wind is within 20 degrees of true south (see Table 2.1), \( v \) is approximately the along-wind component and \( u \) is approximately the cross-wind component for all days...
except 25 May. Excluding 25B May and 7 June, the magnitude of the horizontal wind is lower over more heated area (see Fig. 2.3). The stationary spatial deviations of vertical heat and moisture fluxes based on the leg averages at 65 m AGL are shown in Fig. 2.9. The vertical heat flux shows a correspondence with surface temperature, although less obvious than the potential temperature at 65 m.

2.4.2 Variance Decomposition

To quantify the relative importance of stationary spatial variance which is closely associated with surface heterogeneity to temporal and transient variances, we employed the variance decomposition of Mahrt et al. (1994b).

\[
\left\{ \left[ (\phi^s)^2 \right] \right\} = \left\{ \phi^2 \right\} + \left\{ \phi^s \right\}^2 + \left\{ \phi^{ST} \right\}^2
\]  
(2.7)

Here, the global average, \(\left\{ \left[ \phi \right] \right\} \), was previously removed from the segment averages when a formula for variance was created. For example, the leg average in the first term on the right hand side, \(\langle \phi \rangle \), is obtained from the segment averages by removing the global average. Thus, the terms on the right hand side in Eq. (2.7) were named as temporal, spatial, and transient variances, respectively (Mahrt et al., 1994b). We used this variance decomposition with the five different segment lengths of 1, 4, 6, 8, and 12 km to evaluate the relative significance of spatial variance as a function of spatial scale. Thus it can be determined at which scales the spatially stationary patterns are the most significant on a given day. This variance decomposition is applied to potential temperature, water
vapor mixing ratio, wind components, and vertical heat and moisture fluxes. Here, the deviations for vertical flux of a scalar $\phi$ have been computed relative to the leg averages.

$$w^* \phi^* = \left( w - \langle w \rangle \right) \left( \phi - \langle \phi \rangle \right)$$

(2.8)

where $\phi$ is potential temperature or water vapor mixing ratio. To avoid the complication in the interpretation that arises by introducing a different observation period and a different number of repeated flight passes, the five low-level legs between 17 and 19 UTC (11 and 13 CST) on 19, 25A May, and 7 June are used.

The variance decomposition results are shown in Fig. 2.10. For the potential temperature and the along-wind component, the spatial variance normalized by total variance reaches values greater than 0.5 on 19 May and 25A May, implying significant horizontal variability on both days. Although temporal and transient variances are not insignificant for the vertical heat and moisture fluxes, stationary spatial variances exceed temporal and transient variances at 12 km on 19 May and at 8 km on 25A May. However, on 7 June, stationary spatial variance for all the variables is smaller than temporal and transient variances at all the segment lengths, which implies that the influence of land-surface variability is relatively insignificant.

2.5 Low-level ABL properties

2.5.1 MR Spectra

Multiresolution (MR) spectrum is an alternative method to Fourier spectrum and a direct link to the definition of Reynolds averaging because it uses a wavelet basis set with
a constant basis function. One can refer to Vickers and Mahrt (2003) for a detailed description of the MR spectrum. We use these MR spectra to estimate the relative significance of the contribution of variance (and covariance) at a segment length in comparison to the total values, and to identify the physical processes of the vertical heat and moisture fluxes at each segment length in the low-level ABL over the heterogeneous land surface.

Figs. 2.11, 2.12, and 2.13 show the composites of the MR spectra (or cospectra) normalized by the total variance (or covariance) of each variable, namely \( \{ C_{w\phi}(2^m) \} \), where \( \{ \} \) implies averaging over all the low-level legs on a given day and \( 2^m \) indicates a segment length of \( 2^m \) points. In Figs. 2.11, 2.12, and 2.13, the segment lengths of \( 2^m \) points are converted into meters by using \( L_m = 2^m \times u_{ac} / f \), where \( u_{ac} \) is true air speed, which was about 80 m s\(^{-1}\) and \( f \) is the sampling frequency of 25 Hz.

The composites of the normalized MR spectra of potential temperature (\( C_{\theta\theta} \)) and water vapor mixing ratio (\( C_{qq} \)) are shown in Fig. 2.11. For potential temperature and water vapor mixing ratio, turbulence and mesoscale variances can be separated based on the spectral gap defined approximately between 3 and 15 km. On 25 and 29 May, the peaks of mesoscale variances were more significant than those of turbulence variances both for potential temperature and water vapor mixing ratio. However, the composites of the normalized MR spectra of wind components (\( C_{uu}, C_{vv}, C_{ww} \)) shows the spectral gap being about 10 km only in the along-track wind spectra (Fig. 2.12). That is to say, the spectra of cross-track wind and vertical wind have no spectral gap and mesoscale variance peak. Even for the along-track wind spectra, only on 25A May is the mesoscale
variance peak relatively significant compared to the turbulent variance peak. Thus only on 25A May the higher peaks of mesoscale variances of potential temperature and mixing ratio can be linked to the more significant peak of mesoscale along-track wind variance.

The composites of the cospectra of $w''$ and $\theta''$, $w''$ and $q''$, and $q''$ and $\theta''$ normalized by the total covariance, $\{C_{w\theta}\}$, $\{C_{wq}\}$, and $\{C_{q\theta}\}$, respectively, are shown in Fig. 2.13. In Table 2.3, the combinations of total covariances with differing signs are summarized into four modes with accompanying physical interpretations. On 19, 20, and 25B of May, heating and moistening from the surface, mode I, accounts for 97 %, 95 %, and 91 % (95, 98, and 94 %) of the total vertical heat (moisture) flux respectively. However, the contributions of mode I to the total vertical heat (moisture) flux drop to 84, 86, and 82 % (80, 88, and 83 %) on 25A, 29 May, and 7 June. Interestingly on 25A May, the negative $\{C_{q\theta}\}$ has a peak at 25-km segment length with positive $\{C_{w\theta}\}$ and a negative $\{C_{wq}\}$. In Fig. 2.11a, the variance of potential temperature at 65m AGL shows the mesoscale variance peaks at a 25-km segment length on 25A May. Similarly on 29 May the mesoscale variance peaks at a 50-km segment length in Fig. 2.11 can be linked to the peak of the negative $\{C_{q\theta}\}$ with positive $\{C_{w\theta}\}$ and negative $\{C_{wq}\}$. Considered the combination of warm and dry surface conditions with cool and moist surface conditions over the track as shown in previous figures (see Figs. 2.4, 2.5, 2.6, 2.7, and 2.9), this updraft of warm and dry air and downdraft of cool and moist air can be interpreted as a mesoscale circulation (mode III). The entrainment-drying processes (mode II) were also involved both on 25A and 29 May. On 25B May and 7 June, some portions of the vertical heat and moisture fluxes were described by the combinations of the negative $\{C_{q\theta}\}$ with
positive \( C_{u\theta} \) and positive \( C_{wq} \) (mode IV). For the case on 25B May, we hypothesized that mode IV is the consequence of the mixed signatures of heating and moistening from the surface and from mesoscale circulations. The \( C_{u\theta} \) and \( C_{wq} \) are positive due to heating and moistening by thermals from the surface whereas \( C_{q\theta} \) is negative because it is more influenced by warm and dry, and cool and moist air associated with mesoscale circulations. For the case on 7 June, mode IV seems to be the consequence of the heating and moistening from a surface mixed with dry entrainment. The strong entrainment of dry air from the free atmosphere cause \( C_{q\theta} \) to be highly negative even in the low-level ABL, while \( C_{u\theta} \) and \( C_{wq} \) are still positive due to heating and moistening from the surface.

2.5.2 Joint Probability Distributions

In order to identify similar regimes based on the distributions of the perturbations of potential temperature, water vapor mixing ratio, and vertical velocity, which were collected from the repeated aircraft passes at 65 m AGL, the joint probability distribution (JPD) technique (Holland, 1973; Mahrt and Paumier, 1984; Deardorff and Willis, 1985) is employed. For the total perturbations defined from each leg average, the occurrences of \( q^*/q^*_{ML} \) (\( q^*_{ML} \) is the mixed-layer humidity scale; Stull, 1988) and \( \theta^*/\theta^*_{ML} \) (\( \theta^*_{ML} \) is the mixed-layer temperature scale; Stull, 1988), within the grid of \( \Delta q/q^*_{ML} \) and \( \Delta \theta/\theta^*_{ML} \), were counted. The counted occurrences at each grid were normalized by the total number of points. Thus,
\[\int \int P_{q\theta} dq/q_s^{ML} d\theta/\theta_s^{ML} = 1 \quad (2.9)\]

where \( P_{q\theta} \) is the probability of the occurrence within the grid of \( \Delta q/q_s^{ML} \) and \( \Delta \theta/\theta_s^{ML} \).

Similarly, the joint probability distribution was obtained for both \( w'' \) and \( q'' \), as well as \( w'' \) and \( \theta'' \). For the scaling parameter, the value is first computed for each low-level leg. Next the average of the scales over all the low-level legs on a given day is obtained (see Table 2.4).

The JPDs categorize the six cases into two groups: 1) 19 May, 20 May, and 7 June, and 2) 25A May, 25 B May, and 29 May. Fig. 2.14 shows that on the days in the second group, the dry warm and moist cold quadrants become more significant than those in the first group. In addition, the vertical velocity-water vapor distribution looks more elliptical in the second group (Fig. 2.15). Finally, in Fig. 2.16, the JPD extends slightly more into the warm downdraft and cold updraft quadrants in the second group. These warm downdraft and cold updraft quadrants are associated with entrained free atmosphere and penetrative convection into the entrainment zone, respectively while the warm updraft and cold downdraft quadrants are related to thermals (Mahrt and Paumier, 1984; Deardorff and Willis, 1985). These JPDs imply the existence of different processes associated with vertical heat and moisture fluxes in the low-levels of the ABL over the track, between those two groups.

### 2.6 Vertical Extension

#### 2.6.1 Minimum Length Scale
Previous studies (Pielke, 2001; Avissar and Schmidt, 1998; Shen and Leclerc, 1995; to name a few) have concluded that the effect of large scale surface heterogeneity can extend upward farther into the ABL; the height in the atmosphere at which surface heterogeneity of a given spatial scale and amplitude can no longer be detected is called the blending height (Mahrt, 2000). Based on the approximate balance between the temperature advection and vertical divergence of the perturbation heat flux caused by surface heterogeneity, which is later canceled out during linearization, Wood and Mason (1991) built a thermal blending height formulation. This can be inversely used to estimate the horizontal length scale of surface heterogeneity \( L_{TH} \) that starts to influence the flow at a certain level, \( z \).

\[
L_{TH} = C_{TH} \frac{U\Theta_v}{\left(\bar{w}\bar{\theta}_v\right)_{sfc}} z, \tag{2.10}
\]

where \( C_{TH} = 3.1 \times 10^{-3} \) is the non-dimensional coefficient (Mahrt, 2000), \( U \) is the mean horizontal wind velocity, \( \Theta_v \) is the mean virtual potential temperature, and \( \left(\bar{w}\bar{\theta}_v\right)_{sfc} \) is the surface buoyancy flux (Mahrt, 2000). With respect to the distance the flow can travel during one large-eddy turnover time scale, the convective boundary layer (CBL) scaling formulation suggested by Raupach and Finnigan (1995) can be used to estimate the minimum surface heterogeneity scale that can significantly influence the whole ABL.

\[
L_{CO} = C_{CO} \frac{Uz_i}{w_*} \tag{2.11}
\]

where \( C_{CO} = 0.8 \) is the nondimensional coefficient (Mahrt, 2000), \( w_* \) is the convective velocity scale, and \( z_i \) is the ABL depth.
Using (2.10) the minimum surface heterogeneity scale which significantly influences the flow at 65 m AGL is computed for each low-level leg. Also for each low level leg, the minimum surface heterogeneity scale for the flow at the ABL top is estimated by using (2.11). The composites of the length scales over all of the low-level legs on a given day are summarized in Table 2.5. On 25A May, the estimated values are one order smaller than those on the other days. This implies that surface heterogeneity on a much smaller scale can significantly influence the whole ABL structure on 25A May.

2.6.2 Deviations from Mixed-layer Similarity

Quasi-steady and horizontally homogenous ABL conditions are assumed to obtain the vertical profiles of dimensionless variances of velocity, potential temperature, and moisture, which can be expressed by the mixed-layer similarity relationships (Lenschow et al., 1980; Stull 1988, p. 370-371). The dimensionless variance profiles are estimated from the aircraft measurements over the track for the five days. For vertical velocity, temperature, and moisture, the variances based on leg averages are normalized by their appropriate scaling parameters, $\langle \phi^2 \rangle / \phi^2$, where $\phi$ is convective velocity scale($w_*$), mixed-layer temperature scale($\theta_{ML}^*$), or mixed-layer moisture scale ($q_{ML}^*$). For horizontal and vertical wind, the convective stress velocity scale ($u_{ML}^* = u^2_{*}/w_*$; Stull, 1988) is used to normalize the variances. The scaling parameters obtained from the data collected by the repeated low-level passes are summarized in Table 2.4. For each leg, the flight height
is normalized by the ABL depth which is estimated based on the UWKA soundings and the spatial averages of the ABL depths from the DLR Falcon over the track.

As shown in Fig. 2.17, except below the height of 0.1 \( z_i \) (\( z_i \) is the ABL depth), the normalized variances of horizontal wind velocity are well fitted by

\[
\frac{\langle u'^2 \rangle}{u^*_e M^2} = \frac{\langle v'^2 \rangle}{u^*_e M^2} = \text{const}. \quad \text{(Stull, 1988)}.
\]

Normalized variances of horizontal wind on 25 and 29 May are aligned with the constant values that are nineteen times and four times larger than that on the other three days, respectively. Also, the vertical velocity variances normalized by convective stress velocity, \( \frac{\langle w'^2 \rangle}{u^*_e M^2} \), can be fitted by constant values, which are eighteen times larger on 25 May and three times larger on 29 May than that on the other days (Fig. 2.18a). However, the normalized variances of wind velocity at levels higher than 0.6 \( z_i \) on 7 June deviate from the similarity curves, which may be associated with the weak capping inversion of 2 K (200 m \(^{-1} \)) measured at the Homestead profiling site at 1856 UTC (Fig. 2.2).

In Fig. 2.19a, the variances of potential temperature on 19, 20, May, and 7 June are well fitted by the similarity curve of

\[
\frac{\langle \theta'^2 \rangle}{\theta_i^2} = 1.8(\frac{z}{z_i})^{-2/3} \quad \text{(Lenschow et al., 1980; Stull, 1988)}.
\]

except for the values near the ABL top where the effect of entrainment is significant (Wyngaard and LeMone, 1980). The deviations of the potential temperature variances from the similarity curve on 25 and 29 May are likely associated with the significant mesoscale peaks in the spectra of potential temperature at 65 m AGL (Fig. 2.11a). Moisture variances are generally larger than the expected values from the similarity curve of

\[
\frac{\langle q'^2 \rangle}{q_i^2} = 1.8(\frac{z}{z_i})^{-2/3} \quad \text{(Lenschow et al., 1980; Stull, 1988)}.
\]

In
general these large deviations can be linked to more sensitive response of moisture to entrainment or mesoscale motions than temperature and vertical wind (Mahrt et al., 1994a, b; Mahrt, 1991). Also consistent with the large mesoscale peaks in the spectra of water vapor mixing ratio at 65 m AGL (Fig. 2.11b), the deviations of the variances on 25 and 29 May are much larger than those on the other days.

2.6.3 Observed spatial variations of the ABL depth

The spatial distributions of the ABL depths over the track for the five BLH days, which are estimated from the aerosol backscatter measurements of the DIAL aboard the DLR Falcon, are shown in Fig. 2.21. On 19, 20, and 29 May, the ABL was deeper to the north and shallower to the south, consistent with the surface skin temperature distribution (Fig. 2.3). On 25 May, a different ABL depth distribution, elevated around 36.60°N and depressed around 36.80°N, was also consistent with the surface temperature distribution on that day (Fig. 2.3). Smaller scale surface heterogeneity was expected to significantly influence the whole ABL structure on 25 May based on the minimum length scale argument in section 2.6.1. For the four days, these ABL depth distributions are consistent also with the potential temperature distributions measured at 65 m AGL. On 7 June, only the first leg, flown 1651 and 1704 UTC, showed the deeper-north pattern. On subsequent legs, the ABL grew rapidly across the entire track, associated with a weak capping inversion of 2 K (200m)^{-1} measured at Homestead profiling site at 1856 UTC (Fig. 2.2); and ABL heights were randomly distributed along the track. Thus one can conclude that the features of the surface heterogeneity over the track influenced the entire ABL.
structures except on 7 June. In other words, over the track for the four days among the five days, the blending height determined from the surface heterogeneity exceeded the ABL depths and the ABL established equilibrium with the local surface condition, a regime called macroscale heterogeneity (Mahrt, 2000).

### 2.7 Summary and Conclusions

This study analyzes data collected by aircraft on five fair-weather days during IHOP_2002 to investigate ABL structures over a heterogeneous land surface along the track (36.4°N-37.0°N, 100.6°W; Fig. 1) under different background weather conditions. The tower- and aircraft- measured surface skin temperature distributions both show a warm-north and cool-south pattern over the track. The surface skin temperature was increased by about 10 °C or more from southern end to the northern end of the aircraft track for four of the five days, whereas on 7 June the surface temperature at the northern end of the track was slightly less than 5 °C higher than at the southern end. On 25 May, a warm region around 36.64°N was superimposed on the warm-north and cool-south pattern.

Measurements at the surface flux sites suggest that the surface temperature distribution is mainly caused by the combination of soil thermal properties and soil moisture. With a similar amount of available surface energy (Rn-G) along the track, site 3 located around the northern end of the track (36.86°N) used more of the available energy to produce sensible heat flux due to dry soil, as shown in Fig. 2.6. At site 2 (36.62°N), the heat flux into ground (G) is less due to relatively small soil thermal conductivity.
(Table 2.2). Thus, even with similar amount of latent heat flux, site 2 could produce more sensible heat flux than site 1 due to more available surface energy. On 25 May, surface temperature around site 2 can rise more rapidly due to the shallower active thermal-exchange layer that is associated with the smaller soil thermal conductivity, especially under relatively calm winds. This additional heating around site 2 causes the small scale heterogeneity superimposed on the large scale pattern.

The stationary spatial variability described by the data collected from the repeated aircraft passes at 65 m AGL shows a slowdown of the along-wind component and warmer air temperatures over warmer surfaces. High vertical flux is also associated with warmer surfaces, but the association is less robust due to considerable temporal and transient variability.

Mesoscale circulations (warm, dry updrafts and cool, moist downdrafts) generated by the land-surface heterogeneity can be seen through MR cospectra and JPDs on 25 and 29 May. On these two days, the ABL is likely dominated by buoyancy-generated turbulence, given the smallness of the Obukhov lengths, and the relatively small scale surface heterogeneity needed to influence the whole ABL. On 25 May, MR spectra show a significant peak of mesoscale along-track wind variance (Fig. 2.12b), which is linked to the significant peaks of mesoscale variances of potential temperature and water vapor mixing ratio. Also on these two days, the vertical profiles of dimensionless variances of wind components, potential temperature, and moisture (Fig. 2.17, 2.18, and 2.19) exhibit significant deviations from the similarity curves which are based on the assumption of homogeneous surface conditions. However, when we used high-pass
spatial filters, these dimensionless variances fit the similarity curves better (Fig. 2.20), which also implies the existence of mesoscale circulation on 25 and 29 May.

On 25 and 29 May, however, the lack of a high amplitude mesoscale peak in the wind spectra (Fig. 2.12), except for the along-track wind component on 25A May, makes the existence of well-organized mesoscale circulations doubtful. Considering the scales of the surface heterogeneity feature which can be seen in the ABL depth distribution on these two days, the observed ABL seems to be in the mixed-scale ABL regime of turbulence and less-organized mesoscale circulations. With similar horizontal scales of surface heterogeneity (20-40km), LES studies (Patton et al., 2005; Avissar and Schmidt, 1998) have shown persistent turbulent thermals co-existing with less-organized mesoscale circulations over a sinusoidal-shaped heterogeneous land surface. However, they have mostly focused on the well-organized mesoscale circulations. In future research, we plan to simulate the ABL under the mixed-scale regime which was likely observed on 25 and 29 May by using LES with the background weather conditions measured at the Homestead profiling site (Fig. 1) and the observed surface heat fluxes at the three ISFF sites. These simulations will allow us to investigate ABL structure in detail under the mixed-scale regime in detail and its potential link to convective cloud development.
Fig. 2.1: (Top) Topography along the western track of the UWKA during IHOP, which come from USGS global 30 second dataset. (Middle) Composite structure of NDVI with 4-km means at every 1 km from the repeated passes of the aircraft at 65 m AGL at a given day:0519(Solid), 0520(Dotted), 0525A(Dashed), 0525B(Dash-dotted), 0529(Dash-two-dotted), 0607(Long dashed). (Bottom) STATSGO soil map around the western track. The locations of the ISFF site 1 (rectangle), 2 (circle), and 3 (triangle) are marked. The symbol of X indicates the location of the Homestead profiling site. The color bar indicates the soil categories:1 (Sand), 2 (Loamy Sand), 3 (Sandy Loam), 4 (Silt Loam), 5 (Silt), 6 (Loam), 7 (Sandy Clay Loam), 8 (Silty Clay Loam), 9 (Clay Loam), 10 (Sandy Clay).
Fig. 2.2: Vertical profiles of (a) potential temperature and (b) water vapor mixing ratio from the rawinsonde released over Homestead profiling site (Fig. 2.1; 36.55°N, 100.6°W) at 1740, 1731, 1733, 1741, and 1856 UTC on 19, 20, 25, 29 May, and 7 June, 2002, respectively.
Fig. 2.3: Composite structure of surface radiation temperature with 4-km means at every 1 km from the repeated passes of the aircraft at 65 m AGL at a given day. The rectangles, circles, and triangles indicate the averages of the surface skin temperatures measured at the ISFF site 1, 2, and 3 over the aircraft flight hours at the written day beside the symbols. There were no surface radiation temperatures available on 7 June from the UWKA due to the malfunction of the measurement instrument (downward-looking Heiman KT-19.85 radiometer).
Fig. 2.4: Daily mean values of precipitation (solid line and filled symbol), volumetric soil moisture (dotted line and filled symbol), surface skin temperature (solid line and unfilled symbol), and soil temperature at a depth of 5 cm (dotted line and unfilled symbol) measured at the ISFF site 1 (square), 2 (circle), and 3 (triangle) from 13 May to 7, June 2002.
Fig. 2.5: (Top) Relationship between averages of sensible heat fluxes derived from the data collected by the repeated low-level aircraft passes versus average of the sensible heat fluxes measured at the three ISFF sites during the aircraft flights. (Bottom) Same as (Top) but for latent heat fluxes. Unfilled star denotes the relationship between the aircraft values and the spatial averages of the values at the three sites. The error bar over the star indicates standard error defined as the standard deviation between the low-level legs divided by the square root of the number of the low-level legs (Mahrt, 1998). Here standard error is computed by dividing standard deviation by the square root of the number of the repeated low-level passes. Square, circle, and triangle represent the relationship between the aircraft values and the values measured at site 1, 2, and 3, respectively.
Fig. 2.6: The surface energy balance components measured at the ISFF site 1(solid), 2(dotted), and 3(dashed) averaged for the five days. $R_{\text{net}}$, $H$, $LE$, and $Gs$ represent net radiation, sensible heat flux, latent heat flux, and soil heat flux at a depth of 5 cm. Here, the soil heat flux was measured with a heat flux plate at a depth of 5 cm. For $R_{\text{net}}$ and $Gs$, positive sign is used for downward flux. However, for $H$ and $LE$, negative sign is used for upward flux.
Fig. 2.7: Composite structures of potential temperature, $\{\theta^s\}$, and water vapor mixing ratio, $\{q^s\}$, using 4-km means at every 1km from the repeated passes of the aircraft at 65 m AGL over the western track on (a) 19 May, (b) 20 May, (c) 25A May, (d) 25B May, (e) 29 May, and (f) 7 June, 2002. The error bar indicates standard error defined as the standard deviation between the low-level legs divided by the square root of the number of the low-level legs (Mahrt, 1998).
Fig. 2.8: Composite structures of along-track wind, \( \{ \bar{\nu} \} \), and cross-track wind, \( \{ \bar{\mu} \} \), using 4-km means at every 1 km from the repeated passes of the aircraft at 65 m AGL over the western track on (a) 19 May, (b) 20 May, (c) 25A May, (d) 25B May, (e) 29 May, and (f) 7 June, 2002. The error bar indicates standard error defined as the standard deviation between the low-level legs divided by the square root of the number of the low-level legs (Mahrt, 1998).
Fig. 2.9: Composite structures of vertical heat fluxes, \( \{w^* \theta^*\} \), and moisture fluxes, \( \{w^* q^*\} \), using 4-km means at every 1 km from the repeated passes of the aircraft at 65 m AGL over the western track on (a) 19 May, (b) 20 May, (c) 25A May, (d) 25B May, (e) 29 May, and (f) 7 June, 2002. The error bar indicates standard error defined as the standard deviation between the low-level legs divided by the square root of the number of the low-level legs (Mahrt, 1998).
Fig. 2.10: The ratio of spatial variance to total variance in the variance decompositions of (a) potential temperature and water vapor mixing ratio, of (b) along-track and cross-track winds, and of (c) vertical heat and moisture fluxes. \( \text{th}, \text{q}, \text{v}, \text{u}, \text{wth}, \text{and} \) \( \text{wq} \) represent potential temperature, mixing ratio, along-track wind, cross-track wind, vertical heat flux, and vertical moisture flux, respectively. Here, 1-, 4-, 6-, 8-, 12- km non-overlapping segments are used for the data from the five low-level legs between 17 and 19 UTC in 19 May (square), 25A May (diamond), and 7 June (triangle).
Fig. 2.11: The composite of the normalized MR spectra of (a) potential temperature (\( \theta \)) and (b) water vapor mixing ratio (\( q \)) over all the low-level legs on 19, 20, 25A, 25B, 29 May, and 7 June. The error bar indicates standard error defined as the standard deviation between the low-level legs divided by the square root of the number of the low-level legs.
Fig. 2.12: The composite of the MR spectra of horizontal ($u$, $v$) and vertical velocity ($w$) components over all the low-level legs on 19, 20, 25A, 25B, 29 May, and 7 June. The error bar indicates standard error.
Fig. 2.13: The composites of the normalized MR cospectra of $w''\theta''$ (solid), $w''q''$ (dotted), and $q''\theta''$ (dashed) over all the low-level legs on (a) 19 May, (b) 20 May, (c) 25A May, (d) 25B May, (e) 29 May, and (f) 7 June. The error bar represents standard error. The vertical solid line indicates boundary between mode I and other modes. The mode I, II, III, and IV are explained in Table 2.3.
Fig. 2.14: Joint probability distributions of the perturbations of water vapor mixing ratio and potential temperature on (a) 19 May, (b) 20 May, (c) 25A May, (d) 25B May, (e) 29 May, and (f) 7 June. The color bar indicates the range of the probability in percent within a grid of $\Delta q^*/q^*_ML$ and $\Delta \theta^*/\theta^*_ML$. Here $\Delta q^*$ is $12 \text{ g/kg}$, $\Delta \theta^*$ is $6 \text{ K}$. For each day, the values of $q^*_ML$ and $\theta^*_ML$ are given in Table 4.
Fig. 2.15: Joint probability distributions of the perturbations of vertical velocity and water vapor mixing ratio on (a) 19 May, (b) 20 May, (c) 25A May, (d) 25B May, (e) 29 May, and (f) 7 June. The color bar indicates the range of the probability in percent with \(\Delta w^*/w_*^{ML}\) and \(\Delta q^*/q_*^{ML}\). Here \(\Delta w^*\) is 0.8 ms\(^{-1}\), \(\Delta q^*\) is 12 g/kg. For each day, the values of \(w_*^{ML}\) and \(q_*^{ML}\) are given in Table 2.4.
Fig. 2.16: Joint probability distributions of the perturbations of vertical velocity and potential temperature on (a) 19 May, (b) 20 May, (c) 25A May, (d) 25B May, (e) 29 May, and (f) 7 June. The color bar indicates the range of the probability in percent within a grid of $\Delta w'/w_{s,ML}'$ and $\Delta \theta'/\theta_{s,ML}'$. Here $\Delta w'$ is 0.8 $ms^{-1}$, $\Delta \theta'$ is 6 K. For each day, the values of $w_{s,ML}'$ and $\theta_{s,ML}'$ are given in Table 2.4.
Fig. 2.17: The vertical profiles of the normalized horizontal velocity variances. The dotted lines are the curves of (a) $\langle u'^2 \rangle / u_*^{ML^2} = \text{const.}$ and (b) $\langle v'^2 \rangle / u_*^{ML^2} = \text{const.}$. Here the convective stress velocity scale, $u_*^{ML}$, is defined as $u_*^2 / w_*$, where $u_*$ is friction velocity and $w_*$ is convective velocity scale (Stull, 1988). The constants are 30, 110, and 580 (a) for the along-track wind variance, and 30, 110, and 800 for the cross-track wind variances.
Fig. 2.18: The vertical profiles of the normalized vertical velocity variances normalized by (a) convective stress velocity scale, $u_*^{ML}$, and (b) convective velocity scale, $w_*$. The dotted lines are the curve of (a) $\langle w'^2 \rangle / u_*^{ML} = \text{const.}$ and (b) $\langle w'^2 \rangle / w_*^2 = 1.8(z/z_i)^{2/3} (1 - 0.8 z/z_i)^2$ (from Lenschow et al., 1980). The constants are 30, 90, and 550 for the vertical velocity variances normalized by convective stress velocity.
Fig. 2.19: The vertical profiles of the normalized potential temperature and water vapor mixing ratio variances. The dotted lines are the curve of (a) $\frac{\theta'^2}{\theta^2} = 1.8(z/z_i)^{-3}$, and (b) $\frac{q'^2}{q^2} = 1.8(z/z_i)^{-3}$ (from Lenschow et al., 1980).
Fig. 2.20: The vertical variance profiles of (a) vertical velocity, (b) potential temperature, and (c) water vapor mixing ratio normalized by convective velocity scale, mixed-layer temperature scale, and mixed-layer moisture scale, respectively. For vertical velocity and water vapor mixing ratio, 1-km spatial filter are used. For potential temperature, 4-km spatial filter are used. The dotted lines are the curve of (a) \[ \left\langle w'^2 \right\rangle / w^* = 1.8 (z/z_i)^{2/3} \left( 1 - 0.8 z/z_i \right)^2 \], (b) \[ \overline{\theta'^2}/\theta^* = 1.8 (z/z_i)^{-2/3} \], and (c) \[ q'^2/q^* = 1.8 (z/z_i)^{-2/3} \] (from Lenschow et al., 1988).
Fig. 2.21: The estimated ABL depths based on the aerosol backscatter measurements of the DIAL aboard the DLR Falcon flown over the western track on (a) 19, (b) 20, (c) 25, (d) 29, May, and (e) 7 June, 2002. The circles, filled circles, squares, and filled squares indicate first, second, third, and fourth legs. The first leg on 19 May is flown at 1650-1704 UTC. The first and second legs on 20 May are flown at 1920-1932 UTC and 1941-1953 UTC, respectively. The first, second, and third legs on 25 May are flown at 1745-1758, 1807-1818, and 1844-1857. The first, second, third, and fourth legs on 29 May are flown at 1806-1820, 1829-1839, 1904-1918, and 1927-1938 UTC. The first, second, third, and fourth legs are flown at 1651-1704, 1754-1803, 1829-1835, and 1932-1944 UTC.
Table 2.1: Information regarding the UWKA flights and the ABL characteristics over the western track for the five days. Here L shows the range of the Obukhov length calculated for each low-level leg.

<table>
<thead>
<tr>
<th>Date</th>
<th>The UWKA flights</th>
<th>The ABL characteristics</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Flight time (HHMMSS)</td>
<td># of low-level legs/ # of total legs</td>
</tr>
<tr>
<td>0519</td>
<td>170309-192022</td>
<td>6/10</td>
</tr>
<tr>
<td>0520</td>
<td>172236-193758</td>
<td>2/10</td>
</tr>
<tr>
<td>0525</td>
<td>170110-202123</td>
<td>10/13</td>
</tr>
<tr>
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<td>170110-185010</td>
<td>5/7</td>
</tr>
<tr>
<td>0525B^1</td>
<td>185242-202123</td>
<td>5/6</td>
</tr>
<tr>
<td>0529</td>
<td>164007-202405</td>
<td>5/15</td>
</tr>
<tr>
<td>0607</td>
<td>165037-195859</td>
<td>7/13</td>
</tr>
</tbody>
</table>

^1Note that 25 May is sometimes separated into 25A and 25B May due to the slightly eastward-shifted flight track for the last five legs.

^2This mean wind is computed from the low-level legs at each case.

^3These fluxes of heat and moisture at the surface and the ABL top are estimated based on linear fits to the composite vertical profiles of the sensible and latent heat fluxes from the in situ measurements at surface flux sites and aircraft.
Table 2.2: Estimated soil thermal conductivity ($Wm^{-1}K^{-1}$)

<table>
<thead>
<tr>
<th></th>
<th>May 19</th>
<th>May 20</th>
<th>May 25</th>
<th>May 29</th>
<th>June 7</th>
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</thead>
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<tr>
<td>Site 1</td>
<td>0.57</td>
<td>0.41</td>
<td>0.75</td>
<td>(6.3×10⁴) *</td>
<td>1.34</td>
</tr>
<tr>
<td>Site 2</td>
<td>0.23</td>
<td>0.21</td>
<td>0.30</td>
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<td>Site 3</td>
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<td>0.34</td>
<td>0.61</td>
<td>0.77</td>
<td>0.60</td>
</tr>
</tbody>
</table>

Table 2.3: Combinations of the composited MR cospectra of $w''$ and $\theta''$, $\{C_w\}$, $w''$ and $q''$, $\{C_wq\}$, and $q''$ and $\theta''$, $\{C_{q\theta}\}$ ({} is omitted in the table).

<table>
<thead>
<tr>
<th>Mode</th>
<th>Combinations</th>
<th>Description</th>
<th>Physical Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>I</td>
<td>$C_{w\theta} &gt; 0, C_{wq} &gt; 0, C_{q\theta} &gt; 0$</td>
<td>Updraft: Warm moist air Downdraft: Cool dry air</td>
<td>Heating and moistening from surface</td>
</tr>
<tr>
<td></td>
<td>$C_{w\theta} &lt; 0, C_{wq} &gt; 0, C_{q\theta} &lt; 0$</td>
<td>Downdraft: Warm dry air</td>
<td>Entrainment from free atmosphere</td>
</tr>
<tr>
<td>II</td>
<td>$C_{w\theta} &gt; 0, C_{wq} &lt; 0, C_{q\theta} &lt; 0$</td>
<td>Updraft: Warm dry air Downdraft: Cool moist air</td>
<td>Mesoscale circulation</td>
</tr>
<tr>
<td>III</td>
<td>$C_{w\theta} &gt; 0, C_{wq} &gt; 0, C_{q\theta} &lt; 0$</td>
<td>-</td>
<td>Mixed signatures</td>
</tr>
</tbody>
</table>
Table 2.4: The mixed-layer scaling parameters on the five case days. For the definitions of these scaling parameters, refer to Stull (1988)

<table>
<thead>
<tr>
<th></th>
<th>$u_*$</th>
<th>$w_*$</th>
<th>$u_*^{ML}$</th>
<th>$\theta_*^{ML}$</th>
<th>$q_*^{ML}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>0519</td>
<td>0.76</td>
<td>2.23</td>
<td>0.259</td>
<td>0.139</td>
<td>0.013</td>
</tr>
<tr>
<td>0520</td>
<td>0.76</td>
<td>2.00</td>
<td>0.285</td>
<td>0.145</td>
<td>0.012</td>
</tr>
<tr>
<td>0525</td>
<td>0.27</td>
<td>1.66</td>
<td>0.044</td>
<td>0.119</td>
<td>0.026</td>
</tr>
<tr>
<td>0529</td>
<td>0.39</td>
<td>1.61</td>
<td>0.093</td>
<td>0.088</td>
<td>0.081</td>
</tr>
<tr>
<td>0607</td>
<td>0.54</td>
<td>1.65</td>
<td>0.176</td>
<td>0.104</td>
<td>0.068</td>
</tr>
</tbody>
</table>

Table 2.5: The minimum horizontal length scale of surface heterogeneity, which can significantly influence on the flow at 65 m AGL, estimated from (2.10) and at the ABL top, estimated from (2.11) by using the data from the low-level aircraft passes.

<table>
<thead>
<tr>
<th></th>
<th>May19</th>
<th>May 20</th>
<th>May 25A</th>
<th>May 25B</th>
<th>May 29</th>
<th>June 7</th>
</tr>
</thead>
<tbody>
<tr>
<td>$L_{TH}$</td>
<td>5.21</td>
<td>4.59</td>
<td>0.33</td>
<td>1.33</td>
<td>2.28</td>
<td>4.19</td>
</tr>
<tr>
<td>$L_{CO}$</td>
<td>2.85</td>
<td>2.92</td>
<td>0.35</td>
<td>1.02</td>
<td>2.40</td>
<td>3.63</td>
</tr>
</tbody>
</table>
Chapter 3

Dry ABL Simulation

3.1 Chapter Introduction

The atmospheric boundary layer (ABL) by definition is directly influenced by the earth’s surface, which is continuously heterogeneous. The ABL significantly influences the initiation and development of moist convection (for a recent literature review, see Pielke, 2001). Similarity theory is currently used to describe the ABL. This theory is constructed based on the assumption that the ABL is homogeneous and stationary (Stull, 1988). For a realistic description of the ABL over a heterogeneous surface, the ABL response to the surface heterogeneity needs to be fully investigated. It is difficult to obtain appropriate observations that describe ABL structures over heterogeneous surfaces in a statistically robust manner. Thus, the study of the heterogeneous surface’s effect on the ABL often incorporates numerical models. Due to the extensive scale range of surface heterogeneity, encompassing multiple orders of magnitude, research of this sort has utilized both mesoscale models and large eddy simulation (LES).

Published studies that have applied mesoscale numerical models have focused on sea-breeze-like mesoscale circulations (e.g., Segal et al., 1989; Chen and Avissar, 1994a, b; Reen et al., 2006). The optimal scale of surface heterogeneity needed to generate these mesoscale circulations is suggested by some theoretical studies to be the
local Rossby radius of deformation (e.g., Dalu et al., 1996). This radius is estimated to be about 100 km with typical atmospheric parameters (Pielke, 2001). These mesoscale numerical model studies have suggested that the vertical heat and moisture fluxes directly associated with mesoscale circulations (mesoscale fluxes) are often greater than the turbulent fluxes (e.g., Chen and Avissar, 1994a). They further demonstrated that mesoscale circulations can have a significant effect on cloud development and precipitation (e.g., Chen and Avissar, 1994b; Weaver and Avissar, 2001).

Zhong and Doran (1997, 1998) also performed numerical studies with a mesoscale model; however, they utilized observation-based surface flux distributions instead of idealized conditions. They conclude that the magnitude of the vertical velocities generated by convergence and divergence is overestimated in previous research where idealized surface flux conditions had been used. Thus, they suggest that the mesoscale flux is not as significant as asserted in previous numerical model studies. Zhong and Doran (1997, 1998) also discuss the rarity of extreme contrasts in surface heat flux with variation at scales on the order of 100 km. They suggest that the ABL may be affected by surface heat flux variations on scales on the order of 10 km. These apparently contradictory findings stimulated this research, which uses LES for investigating the ABL over a heterogeneous surface.

LES can explicitly resolve energy-containing eddies in the convective ABL that would be parameterized in a mesoscale model. Some LES studies have investigated modifications of the turbulence statistics in the ABL over heterogeneous surfaces where the heterogeneity is at a scale similar to the ABL depth (e.g. Hadfield et al., 1991, 1992; Shen and Leclerc, 1995; Raasch and Harbusch, 2001; Kim et al., 2004). Other LES-based
research on the heterogeneous ABLs includes numerical experiments with surface forcing at a scale one order of magnitude greater than the ABL depth (e.g. Avissar and Schmidt, 1998; Letzel and Raasch, 2003; Patton et al, 2005).

Avissar and Schmidt (1998) show that a sinusoidally prescribed surface heat flux variation with a wavelength of 20-40 km generates well-organized mesoscale circulations. Based on results from an LES coupled with a land surface model, Patton et al. (2005) suggest that a wavelength of 5-15 km is the optimal scale range of surface heterogeneity to generate well-organized mesoscale circulations. Although the suggested optimal scales differ, Avissar and Schmidt (1998) and Patton et al. (2005) agree that over these scales surface-heterogeneity-induced mesoscale circulations coexist with persistent turbulent flows. Roy and Avissar (2000) show that the wavelet spectra of the horizontal wind from Avissar and Schmidt (1998) have two peaks; one at a scale of 1.5 times the ABL depth representing turbulent flows (Kaimal et al., 1972), and the other at the scale of the prescribed heterogeneity, structure of the representing mesoscale flows. Although they do not describe in-depth the ABL containing the mesoscale circulations coexisting with turbulent eddies, Avissar and Schmidt (1998) suggest that both the horizontal pressure gradient caused by the surface heterogeneity and buoyancy must be considered for this ABL structure to be adequately characterized.

Recently Letzel and Rassch (2003), with a numerical experiment design similar to Avissar and Schmidt (1998), reported that surface-heterogeneity-induced mesoscale circulations undergo a temporal oscillation in the volume-average of mesoscale and turbulent kinetic energies, which they call kinetic perturbation energy. However, Patton et al. (2005) observe no temporal oscillations in the kinetic perturbation energy. In
addition, previous studies appear to be inconclusive regarding the effect of mesoscale circulations on entrainment. Avissar and Schmidt (1998) and Letzel and Raasch (2003) assert that mesoscale circulations increase entrainment in the convective ABL, whereas Patton et al. (2005) contend that turbulence dominates the entrainment process even over a heterogeneous surface. These inconsistent conclusions may be due to the selected amplitudes of surface heat flux variation. Avissar and Schmidt (1998) and Letzel and Rassch (2003) specify an amplitude of 0.1-0.2 Kms$^{-1}$ (about 120-250 Wm$^{-2}$) over wavelengths of up to 40 km. Patton et al. (2005) indirectly prescribe an amplitude of 0.03 Kms$^{-1}$ (about 40 Wm$^{-2}$) over wavelengths of up to 30 km.

We simulate fair-weather convective ABLs (1130-1430 LST) over heterogeneous land surfaces by using a recently developed compressible nonhydrostatic numerical model (Bryan and Fritsch, 2002) as an LES. Previous studies have focused on the determination of the scales of surface heterogeneity that are optimal for generating well-organized mesoscale circulations. This investigation focuses on the effect of the intensity of the given mesoscale surface heterogeneity. This new approach will reveal information regarding the apparently inconsistent conclusions of the previous LES studies of the ABL in the presence of mesoscale surface heterogeneity. In addition, the modulation of turbulence statistics in the heterogeneous ABL will be investigated as a function of the magnitude of the surface flux heterogeneity.
3.2 Numerical Experiment

3.2.1 Observational Background

This numerical study is partially motivated by the fair-weather convective ABL structures observed over a heterogeneous land surface during the International H20 Project (IHOP_2002; Weckwerth et al., 2004). The heterogeneous surface that is the focus of the present study is a 60-km north-south oriented strip from approximately 36.4 °N to 37.0 °N along 100.6 °W. Although there was no sea-breeze-like circulation observed, the ABL structures over this heterogeneous surface are obviously different from those over a homogenous surface (for detailed description, see Kang et al. (2006)).

Over the 16-km transect of this heterogeneous surface, from Integration Surface Flux Facility (ISFF) site 1 (36.47 °N) to site 2 (36.62 °N), repeated aircraft flights at approximately 65 m above ground level for the five fair-weather days persistently showed that surface heat flux increased linearly from south to north (Kang et al., 2006). Based on the measurements at these two surface sites during fair-weather midday (from 1130 to 1430 LST) conditions, the probable range of sensible heat flux gradient between these two sites is estimated in Table 3.1. Soundings of pressure, density, and potential temperature from the rawinsonde released over the Homestead site (36.55 °N, 100.6 °N) in the midst of this heterogeneous surface at about 1130 LST on 25 May are selected for the initial conditions for the LES runs. The initial potential temperature is constant (296 K) at altitudes less than 626 m, and increases at the rate of 34.8 $Kkm^{-1}$ between 626 and 806 m. Above 806 m, the potential temperature increases at the rate of 3.8 $Kkm^{-1}$.
3.2.2 Surface heat flux variation

As in many previous numerical studies of the ABL over a heterogeneous land surface (e.g., Schmidt and Avissar, 1998; Letzel and Raasch, 2003), we shall prescribe a surface heat flux variation that is sinusoidal with mean value $\langle F_{w\theta} \rangle(z = 0)$, amplitude $A_{w\theta}$, and wavelength $\lambda$:

$$F_{w\theta}(x_1, x_2; z = 0) = \langle F_{w\theta} \rangle(z = 0) + A_{w\theta} \sin\left(\frac{2\pi}{\lambda} x_1\right) \quad (3.1)$$

This sinusoidal variation of surface heat flux generates sea-breeze-like circulations as schematically illustrated in Fig. 3.1 (e.g., Letzel and Raasch, 2003; Avissar and Schmidt, 1988; Hadfield et al., 1991). Based on the potential temperature and wind velocity distributions along the $x_1$ axis expected from this sinusoidal surface heat flux variation, the ABL is divided into two pairs of regions: (1) warmer ($0 \leq x_1 < \lambda/2$) and cooler regions ($\lambda/2 \leq x_1 < \lambda$), and (2) middle ($\lambda/4 \leq x_1 < 3\lambda/4$) and edge regions ($3\lambda/4 \leq x_1 < \lambda$ and $0 \leq x_1 < \lambda/4$). In this study, $x_1$ is the streamwise direction (and south to north to match our IHOP_2002 case studies), and $x_2$ is the crosswind (west-east) direction.

In Eq. (3.1), $\langle F_{w\theta} \rangle(z = 0)$ is set at 0.21 Kms\(^{-1}\) (260 Wm\(^{-2}\)), which is estimated based on the composite heat flux profile obtained from the measurements at the surface flux sites and aircraft over the 60-km strip-like land surface over 1130-1430 LST on 25 May during IHOP_2002 (Kang et al., 2006). The wavelength $\lambda$ in Eq. (3.1) is determined to be 32 km to emulate the slope between the monotonically varying surface heat flux variation observed over the 16-km transect between site 1 and site 2. In order to
investigate the effect of wavelength on the results found with $\lambda = 32 \, km$, simulations with $\lambda = 16 \, km$ also are performed. We report on the simulation results with $\lambda = 16 \, km$ when they differ significantly from the $\lambda = 32 \, km$ results. The amplitude $A_{\omega\theta}$ in (1) is varied extensively from 0 to 0.2 Kms$^{-1}$. For a heterogeneous surface, the amplitude starts with 0.02 Kms$^{-1} (25 \, Wm^{-2})$, approximately the average value of the observed sensible heat flux differences between the two surface sites (Table 3.1). The amplitude increases to 0.2 Kms$^{-1} (250 \, Wm^{-2})$, which is the maximum amplitude used in Avissar and Schmidt (1998) and Letzel and Raasch (2003). The simulations performed with various values of $A_{\omega\theta}$ and $\lambda$ are summarized in Table 3.2.

3.2.3 Model description and setup

This study utilizes the compressible nonhydrostatic numerical model of Bryan and Fritsch (2002) as an LES. The Bryan-Fritsch model, developed as a cloud-resolving model using LES techniques, has been used to solve many different nonlinear moist convection problems (e.g., Fanelli and Bannon, 2005; James et al., 2006). In addition, from an LES perspective, this Bryan-Fritsch model has been applied to investigate the appropriate spatial resolution for the simulation of deep moist convection (Bryan et al., 2003). The utility of this Bryan-Fritsch model as an LES for the convective ABL study is demonstrated by producing turbulence statistics of the ABL over a homogeneous land surface. The expected turbulence statistics are well-known from both observational and numerical studies (e.g. Lenschow et al., 1980; Nieuwstadt et al., 1991). Some of these...
turbulence statistics will be presented in section 3.5, along with the results from numerical experiments for the ABLs over heterogeneous surfaces.

This Bryan-Fritsch model integrates the filtered compressible Navier-Stokes equation using third-order Runge-Kutta time differencing and fifth-order spatial derivatives for the advection terms. This coupled numerical scheme has been evaluated as the most accurate finite-difference solution for a highly nonlinear flow simulation by Wicker and Skamarock (2002). The turbulence kinetic energy scheme of Deardorff (1980) is employed for a subfilter-scale parameterization (Bryan and Fritsch, 2002).

In an LES experiment the large, energy-containing turbulent eddies have to be explicitly computed. Given an effective resolution of $6\Delta$ (where $\Delta$ is a grid mesh size; Bryan et al., 2003) and a typical size of energy-containing turbulent eddies ($l \approx 1.5z_i$, where $z_i$ is the ABL depth; Kaimal et al., 1976), the grid mesh size has to be in the range of $\Delta < 0.25z_i$. Thus, considering the initial ABL depth of about 700 m (section 3.2.1), the grid spacing is 100 m in the horizontal. While this is relatively coarse resolution as compared to some state-of-the-art simulations of the ABL, our goal is to study the mesoscale organization of ABL convective turbulence. Therefore, we sacrifice fine-scale resolution in exchange for a large spatial domain, while retaining enough resolution to simulate well the dominant convective eddies.

Due to the strong height dependence of the vertical velocity length scales and the weakness of subfilter-scale parameterization, the LES fidelity in the surface layer is always questionable. Wyngaard et al. (1998) suggest the adoption of a high grid aspect ratio (horizontal dimension/ vertical dimension). This high aspect ratio would produce a
mean surface-exchange coefficient from many coupling eddies instead of from one
coupling eddy. Thus, the grid spacing in the vertical is 10 m up to 100 m above the
ground, linearly increases from 10 m to 40 m between 100 m and 1900 m above ground,
and then remains constant at 40 m up to the model top of 3500 m.

The model domain is 32 km (or 16 km) in length in the $x_1$ (here, south-north and
streamwise) direction, and is 5 km in length in the $x_2$ (here, west-east and crosswind)
direction. Without the Coriolis force, mesoscale flows driven by the surface forcing
prescribed with (1) are likely to be generated only in the $x_1$ direction, as shown in
Fig. 3.1. Thus, in the $x_2$ direction, it is assumed that there are only turbulent flows. In
both horizontal directions the lateral boundary conditions are periodic. The vertical
extent of the domain is 3.5 km. The upper boundary is a flat, rigid wall with a Rayleigh
damping layer (Durran and Klemp, 1983) occupying 1 km beneath the model top. The
lower boundary is also a flat, rigid surface. Unlike the prescribed surface heat flux, the
surface momentum flux is derived from a simple surface drag parameterization (Stull,
1988).

The mesoscale surface heat flux variation is activated upon initiation of the
simulation. Also, at this initial time, random perturbations of 0.1 K that initiate
development of three-dimensional turbulent flows are superimposed on the potential
temperature at the lowest atmospheric level. Considering that a model spin-up time is
less than 1 h and that this is a midday ABL simulation (1130-1430 LST), the 4 h
integration time is sufficient. Thus, we use the temporally-fixed surface heat flux
distribution and neglect the Coriolis force. Although occasionally we use the results from
the model run up to 9 h, the discussion is typically based on the results from the model run between 1 h and 4 h. For all the cases, the integration time step is set to 1 second and the model output was saved every 100 seconds.

3.3 Horizontal averaging and Wave-cutoff filter

3.3.1 Horizontal averages

At a given height \((z)\) and time \((t)\), one of the variables from an LES result is denoted by \(\phi(x_i; z, t)\) where \(i = 1, 2\). An LES variable is horizontally re-filtered as

\[
\overline{\phi}(x_i; z, t) = \int_{-\infty}^{\infty} G(x_i - x'_i) \phi(x'_i; z, t) dx'_i
\]

(3.2)

where \(G(x_i - x'_i)\) is a horizontal spatial filter function and \(x_i\) is limited to values of the form \(x_i(n) = nS_i - S_i/2\) \((n = 1, 2, \ldots, L_i/S_i)\). Here, \(L_i\) is the size of the model domain in the \(x_i\) direction and \(S_i\) is a filter scale in meters, which can be represented as \(S_i = N_i \times \Delta x_i\) (where \(N_i\) is the number of LES grid points to be included in the filter scale and \(\Delta x_i\) is the LES grid mesh size, here 100 m). The range of the filter scale, \(S_i\), is \(\Delta x_i \leq S_i \leq L_i\). We use a filter function \(G\) that is a simple spatial average defined as

\[
G(x_i - x'_i) = \begin{cases} 
1/(S_i \times S_2), & |x_i - x'_i| \leq S_i/2 \\
0, & |x_i - x'_i| > S_i/2
\end{cases}
\]

(3.3)

In this numerical experiment, surface heat flux variation is only prescribed in the \(x_1\) direction, therefore we expect a homogeneous turbulent field in the \(x_2\) direction.
Thus, the filter scale in the $x_2$ direction is always the size of the model domain in that direction ($S_2 = L_2$). However, to investigate the generated mesoscale flows and their influences on the turbulent flows, several different filter scales are used for the $x_1$ direction. When $S_1 = \Delta x_1$, the re-filtered variable is denoted by $\phi(x_1; z, t)$. With $S_1 = \lambda/2$, the variables averaged over the warmer, cooler, middle, and edge regions defined in section 3.2.2 are indicated by $\phi^w(z, t)$, $\phi^c(z, t)$, $\phi^m(z, t)$, and $\phi^e(z, t)$, respectively. When $S_1 = L_1$, the domain-averaged variable is denoted by $\langle \phi \rangle(z, t)$. From this horizontal domain average, the total fluctuation is defined as

$$\phi'(x_1, x_2; z, t) = \phi(x_1, x_2; z, t) - \langle \phi \rangle(z, t) \quad (3.4)$$

3.3.2 Wave-cutoff filter

To extract the mesoscale component from an LES variable, the variable $\phi(x_1; z, t)$ (which represents a simple cross-wind mean value) is re-filtered with a one-dimensional, low-pass wave-cutoff filter function $G'$ in the $x_1$ direction.

$$[\phi](x_1; z, t) = \sum_{x_1 - N_1/2}^{x_1 + N_1/2} G'(x_1 - x_1') \phi(x_1'; z, t) \quad (3.5)$$

where $N_1 = S_1 / \Delta x_1$. In Eq. (3.5), the wave-cutoff filter $G'$ is

$$G'(j) = C^j \quad (3.6)$$
where \( j \) is a location relative to the re-filtered variable location. In Eq. (3.6), \( C^j \) are the weighting coefficients at \( j \) and these coefficients are determined by the choice of cutoff wavelength \( S_c \) (Tong et al., 1998). For the four cutoff wavelengths of \( \lambda/2 \), \( \lambda/4 \), \( \lambda/8 \), and \( \lambda/16 \), Fig. 3.2 shows the Fourier transform of the filter function \( \hat{G} \), referred to as the filter response function by Horst et al. (2004),

\[
\hat{G}(\kappa_i) = 2\sum_{j=1}^{N/2} C^j \cos(\kappa_i j)
\]

where \( \kappa_i \) is the wavenumber in the \( x_i \) direction. We use this low-pass wave-cutoff filter to decompose the LES variables into a low-pass filtered component and a residual,

\[
\phi(x_1, x_2; z, t) = [\phi(x_1; z, t) + \phi^\sigma(x_1, x_2; z, t)]
\]

We shall refer to the first and second terms in the right-hand side of Eq. (3.8) as mesoscale and microscale (turbulent) components, respectively.

### 3.3.3 Variance and vertical heat flux decomposition

By applying Eq. (3.8) to Eq. (3.4), the total fluctuation can be written as:

\[
\phi'(x_1, x_2; z, t) = [\phi(x_1; z, t) - \langle \phi \rangle(z, t) + \phi^\sigma(x_1, x_2; z, t)]
\]

Here, we define a mesoscale fluctuation as

\[
\phi^\mu(x_1; z, t) = [\phi(x_1; z, t) - \langle \phi \rangle(z, t)]
\]

Using this in Eq. (3.9), the total variance of an LES variable is separated into three terms:

\[
\langle \phi'^2 \rangle = \langle \phi^\mu^2 \rangle + \langle \phi^\sigma^2 \rangle + 2\langle \phi^\mu \phi^\sigma \rangle
\]
We shall refer to these three terms on the right-hand side (rhs) of Eq. (3.11) as mesoscale, microscale (turbulent), and interscale components, respectively.

Similarly, the total vertical heat flux is separated into four terms, in which case

\[ \langle w' \theta'' \rangle = \langle w^M \theta^M \rangle + \langle w'' \theta'' \rangle + \langle w^M \theta'' \rangle + \langle w'' \theta^M \rangle \] (3.12)

The four terms in the rhs of Eq. (3.12) will be referred to as the components of mesoscale, microscale (turbulent), interscale between \( w'' \) and \( \theta^M \), and interscale between \( w^M \) and \( \theta'' \), respectively.

### 3.4 Surface-heterogeneity-induced mesoscale horizontal flows

In this section, we attempt to explain the mechanism behind the mesoscale fluctuations which are generated by a given mesoscale surface heat flux variation by using the low-pass filtered components defined in Eq. (3.8).

#### 3.4.1 \([u_i]\)-momentum and \([\theta]\) equations

The filtered Navier-Stokes equation in a dry atmosphere, neglecting the Coriolis force and viscous terms, is written as

\[ \frac{\partial u_i}{\partial t} + u_j \frac{\partial u_i}{\partial x_j} = -\frac{1}{\rho_0} \frac{\partial p}{\partial x_i} - \frac{\rho'}{\rho_0} \delta \frac{\partial}{\partial x_i} + \frac{1}{\rho_0} \frac{\partial \tau_{ij}}{\partial x_j} \] (3.13)

where \( p \) is pressure, \( \rho_0(z) \) is the base-state density satisfying hydrostatic balance, \( \rho' \) is the perturbation from the hydrostatic reference state, \( \delta \) is Kronecker delta, and \( \tau_{ij} \) are the
subfilter-scale fluxes produced by LES spatial filtering. In Eq. (3.13), \( u_i (i = 1, 2, 3) \) are the velocity components in the \( x_i \) direction, which are also denoted by \( v, u, \) and \( w, \) respectively.

The mesoscale flows appear to be generated only in the \( x_i \) direction because the mesoscale surface heat flux variation Eq. (3.1) is prescribed only in this direction. Thus, by using Eq. (3.8) with the assumptions of \( [u_2] = 0 \) and \( [u_3] = 0 \), the \( u_1 \)-momentum equation is decomposed into

\[
\frac{\partial [u_1]}{\partial t} + \frac{\partial u_1^*}{\partial t} + [u_1] \frac{\partial [u_1]}{\partial x_1} + [u_1] \frac{\partial u_1^*}{\partial x_1} + u_j^* \frac{\partial [u_1]}{\partial x_j} + u_j^* \frac{\partial u_1^*}{\partial x_j} = - \frac{1}{\rho_0} \frac{\partial [p]}{\partial x_1} - \frac{1}{\rho_0} \frac{\partial p^*}{\partial x_1} + \frac{1}{\rho_0} \frac{\partial \tau_{1j}}{\partial x_j} + \frac{1}{\rho_0} \frac{\partial \tau_{1j}^*}{\partial x_j} \tag{3.14}
\]

After refiltering Eq. (3.14) with Eq. (3.8) and using \( [[\phi] \phi^*] = 0, \)

\( [[\phi] [\phi]] = [\phi] [\phi], \quad [\phi^*] = 0, \)

and the commutative property of spatial averaging (\( \frac{\partial \phi}{\partial t} = \frac{\partial [\phi]}{\partial t} \)), the \([u_1]\)-momentum equation is

\[
\frac{\partial [u_1]}{\partial t} + [u_1] \frac{\partial [u_1]}{\partial x_1} + \frac{\partial \left[ u_1^* \left( \frac{\tau_{1j}^* - \tau_{1j}}{\rho_0} \right) \right]}{\partial x_j} = - \frac{1}{\rho_0} \frac{\partial [p]}{\partial x_1} \tag{3.15}
\]

The zero-divergence assumption was used to obtain the third term on the left-hand side (lhs) of Eq. (3.15). This assumption is valid when the vertical displacement is small compared to the density scale height (which is on the order of 10 km).

By using Eq. (3.8) also with the assumptions \( [u_2] = 0 \) and \( [u_3] = 0 \), the filtered potential temperature equation, neglecting the molecular diffusion term, is reduced to 3.16
where $f_i$ is the subfilter-scale flux of potential temperature. Refiltering Eq. (3.15) with Eq. (3.8) yields

$$\frac{\partial \theta}{\partial t} + \frac{\partial \theta^*}{\partial t} + [u_i] \frac{\partial \theta}{\partial x_i} + [u_j] \frac{\partial \theta^*}{\partial x_j} + u_j \frac{\partial \theta^*}{\partial x_j} + \frac{\partial f_j}{\partial x_j} + \frac{\partial f_j^*}{\partial x_j} = 0 \quad (3.16)$$

3.4.2 Mesoscale temperature gradient

By assuming horizontal homogeneity in the warmer region, Eq. (3.17) can be written as

$$\frac{\partial \theta}{\partial t} + [u_i] \frac{\partial \theta}{\partial x_i} + \frac{\partial [u_j \theta^* + f_j]}{\partial x_j} = 0 \quad (3.17)$$

By integrating Eq. (3.18) vertically through the well-mixed ABL in the warmer region, one can obtain the local time evolution of the vertically averaged mixed-layer potential temperature,

$$\frac{\partial \theta_{ML}^w}{\partial t} = -\frac{\partial F_{w\theta}^w}{\partial z} \quad (3.18)$$

where $F_{w\theta}^w \equiv w^\theta \theta^* + f_3$.

By integrating Eq. (3.18) vertically through the well-mixed ABL in the warmer region, one can obtain the local time evolution of the vertically averaged mixed-layer potential temperature,

$$\frac{\partial \theta_{ML}^w}{\partial t} = \frac{F_{w\theta}^w}{z_i} (z = 0) - \frac{F_{w\theta}^w}{z_i} (z = z_i) \approx (1 - \alpha) \frac{F_{w\theta}^w}{z_i} (z = 0) \quad (3.19)$$

where $\theta_{ML}^w$ indicates a mixed-layer potential temperature and $z_i^w$ is the ABL top averaged over the warmer region. In Eq. (3.19), it is assumed that the heat flux at the ABL top is expressed as a constant fraction of the value at the surface $(\alpha = F_{w\theta}^w (z = z_i) / F_{w\theta}^w (z = 0)$; Stull, 1988). After applying the horizontal homogeneity
assumption to the cooler region, the mixed-layer potential temperature gradient between the warmer and cooler regions can be approximated as

$$
\delta \theta_{ML} = \overline{\theta^W_{ML}} - \overline{\theta^C_{ML}} \approx \alpha' A_{w\theta} \langle \theta' \rangle \frac{t}{z_i}
$$

(3.20)

where \( \alpha' \equiv 1.27(1 - \alpha) \), and 1.27 is the difference of the means of the sine function between 0-\( \pi \) and \( \pi-2\pi \). In Eq. (3.20), \( A_{w\theta} = \overline{F_{w\theta}^W (z = 0) - F_{w\theta}^C (z = 0)} \) is used and \( \langle \theta' \rangle = 0.5 \times \left( \overline{\theta^W_{i}} + \overline{\theta^C_{i}} \right) \) is assumed. This growth of potential temperature gradient between the warmer and cooler regions causes the horizontal pressure gradient to develop. According to Eq. (3.15), this pressure gradient creates mesoscale horizontal flows across the middle (or edge) region.

As the generated mesoscale horizontal flow becomes significant, the temperature advection by this generated mesoscale flow, the second term in the lhs of Eq. (3.17), is not negligible. This temperature advection reduces the temperature gradient between the warmer and cooler regions. In addition, if there is negative (or positive) \( u_i \) and negative (or positive) \( \partial[\theta]/\partial x_i \) in the lower ABL, this potential temperature advection is cold advection; whereas at the upper ABL, it is warm advection. The cold advection at the lower level and warm advection at the upper level stabilizes the ABL overall; the significance of this stabilization varies depending on the amplitude of surface heat flux variation.

We hypothesize that a balanced state between the growth of the temperature gradient by differential surface heating and the decay of the gradient by temperature advection can be reached, and express this balance as

\[ \text{balance expression} \]
Using our prior estimate of the mesoscale horizontal temperature gradient, this balance becomes,

$$-rac{\partial [F_{w \theta}]}{\partial z} \approx [u_i] \frac{\partial [\theta]}{\partial x_i} \quad (3.21)$$

where \( V \) represents a characteristic velocity scale of the generated mesoscale horizontal flows. Making use of Eq. (3.20) in Eq. (3.22), one can estimate the time required for the temperature advection to become large enough to balance the vertical flux divergence,

$$\left( F_{w \theta}(z = z_i) \right) - \left( F_{w \theta}(z = 0) \right) \approx V \frac{\delta \theta_{\text{ML}}}{\lambda/2} \quad (3.22)$$

where \( V \) is defined as the ratio of the domain-averaged surface heat flux to the amplitude of the surface heat flux variation, and \( \tau_v \) is the time required for the generated mesoscale horizontal flow to travel over a region (the advection time scale; \( \tau_v = \lambda/2V \)). Based on Eq. (3.23), one can conclude that for larger amplitude surface flux heterogeneity (larger \( A_{u \theta} \)), the time \( \tau_e \) required to reach equilibrium between temperature advection and vertical flux divergence is smaller. One can imagine that horizontal advection could go beyond a point of equilibrium and begin to attenuate the potential temperature gradient generated by heterogeneity in the surface fluxes. This attenuation of the potential temperature gradient would then weaken the mesoscale horizontal flows. The weakened mesoscale flows would then reduce the potential temperature advection and create an environment for the potential temperature gradient to be intensified again by the time-
invariant surface differential heating described by Eq. (3.1). This would result in an oscillatory mesoscale flow, as was found by Letzel and Raasch (2003).

3.4.3 Critical velocity scale, $V_c$

In this section, we attempt to obtain an estimate of the critical scale of the generated mesoscale horizontal velocity ($V$) which may initiate a temporally oscillating mesoscale flow which is also as significant in magnitude as turbulent flows. First, in terms of scaling variables, the $[u_t]$-momentum equation Eq. (3.15) is written as

$$\frac{V}{\tau} + \frac{V}{L} + \frac{u_t^2}{l} = \frac{1}{\rho_o L} \frac{P}{L}$$

(3.24)

where $\tau$ is the Eulerian time scale, $u_t$ and $l$ the velocity and size of energy-containing turbulent eddies, $P$ the characteristic pressure scale, and $L$ the characteristic horizontal scale, here $0.5\lambda$.

Assuming that the Eulerian time scale $\tau$ and the advection time scale $L/V$ ($\tau_e \equiv \lambda/2V$) are the same order of magnitude, one can write

$$\frac{V}{\tau} \approx \frac{V}{\lambda/2V}$$

(3.25)

At a given wavelength and amplitude of surface heat flux variation, the advection time scale ($\tau_e \equiv \lambda/2V$) is inversely proportional to $V$, but the Eulerian time scale, which we assume is determined by the pressure gradient force, is proportional to $V$ ($\tau \approx \rho_o \frac{VL}{P}$).

The terms $V$ and $P$ grow with time due to the continuity of surface differential heating.
which determines the pressure gradient force. At an earlier time when $V$ is smaller, the Eulerian time scale ($\tau$) is smaller than the advection time scale ($\tau_v$). In other words, the advection terms, the second terms in the lhs of both Eq. (3.15) and Eq. (3.17), can be disregarded and a linear approximation may work in this time period. However, if the amplitude ($A_{\varphi \theta}$) of surface heat flux variation is high enough, $V$ can grow to the point where Eq. (3.25) is satisfied. Thus, $\tau$ becomes larger than $\tau_v$, and the advection terms are too significant to be ignored. As discussed in the previous section, the temperature advection plays the role of attenuating the mesoscale temperature gradient, which is a necessary (though perhaps not sufficient) condition for the initiation of a temporally-oscillating horizontal mesoscale flow.

Next we hypothesize that when the advection term is equal to the turbulence term in Eq. (3.24), this defines a critical velocity, $V_c$ which must be attained in order to initiate a temporally-oscillating mesoscale flow:

$$\frac{2V_c^2}{\lambda} = \frac{u_t^2}{l}$$

(3.26)

Here, we assume that the typical size ($l$) and velocity ($u_t$) of energy-containing turbulent eddies still can be scaled with the mixed-layer depth ($z_l$) and the convective velocity scale ($w_*$), respectively. Thus, using $l = c_z z_l$ and $u_t = c_u w_*$, where $c_z$ and $c_u$ are coefficients to be determined, Eq. (3.26) can be rewritten as

$$V_c \approx c \sqrt{\frac{\lambda}{z_l}} w_*$$

(3.27)
where \( c \equiv \frac{c_u}{\sqrt{2c_z}} \). Lenschow et al. (1980) demonstrates that \( c_u = 0.4 \) and Kaimal et al. (1972) reveals that \( c_z = 1.5 \). Thus, the coefficient \( c \) in Eq. (3.27) is estimated as 0.2. If the mesoscale velocity that is generated exceeds this critical velocity, we hypothesize that an oscillatory flow can be initiated.

We additionally suggest that the magnitude of the critical velocity must be greater than the turbulent, or microscale velocity, \((V_t \geq u_t)\), for an oscillatory flow that is greater in magnitude to the microscale flow to be initiated. By substituting the critical velocity definition from Eq. (3.27), and \( u_t = 0.4w_t \) (Lenschow et al., 1980) into the inequality above, one can show that if the wavelength of surface heat flux variation is greater than or equal to four times the ABL depth \((\lambda \geq 4z_i)\), the critical velocity is greater than or equal to the characteristic turbulent velocity scale. To clarify, we suggest that the generated mesoscale velocity must exceed the critical velocity, and that the critical velocity must be greater than the turbulent velocity for an oscillatory mesoscale flow that is comparable in magnitude to the turbulent flow to develop. If these conditions are correct, this implies that wavelengths smaller than \( 4z_i \) do not produce oscillatory mesoscale flows which are comparable in magnitude to the turbulent flows. Letzel and Raasch (2003) reported that there is no clear oscillation produced in their simulations with \( \lambda = 2z_i \) and \( \lambda = 4z_i \). The magnitude of the mesoscale flow generated at these wavelengths in their simulations also appears to be small compared to the microscale flows.
3.5 Model results

3.5.1 Domain-averaged statistics

We first examine domain-averaged vertical profiles of potential temperature. The vertical profiles are computed at each hour between 0-4 h (Fig. 3.3). Whereas the vertical profiles from a homogeneous ABL (BA000) demonstrate a typical well-mixed ABL structure, the profiles from the heterogeneous ABLs exhibit a somewhat statically stable ABL structure, with a warmer upper level and a cooler lower level. Previous studies (e.g., Letzel and Rassch, 2003; Avissar and Schmidt, 1998) also have shown these statically stable ABL structures over heterogeneous surfaces. This increased static stability likely results from the potential temperature advection by the generated mesoscale flows, as explained in section 3.4.2. From the perspective of the domain-averaged statistics, this heterogeneous ABL reduces buoyancy and decreases the intensity of turbulent vertical velocity throughout the model domain (section 3.5.2 and 3.5.4).

Fig. 3.4a shows the vertical profiles of total vertical heat flux averaged over 1-4 h. In Fig. 3.4b, the temporal variability is estimated by the standard deviation of total vertical heat flux between 1-4 h. Here, standard deviation represents the distribution of profile values across time. This standard deviation is normalized by the magnitude of the temporally averaged vertical heat flux. For all cases, the vertical profiles of total vertical heat flux averaged over 1-4 h are linear. The temporal variability for the ABLs with higher $A_w$ (BA150, BA200, and BA250) is significantly larger than that for the
homogenous ABL (BA000). In contrast, the temporal variability for the ABLs with lower $A_{w\theta}$ (BA025 and BA050) is similar to that for BA000. This result implies that only the ABLs with lower $A_{w\theta}$ can satisfy the quasi-steady state condition. In Fig. 3.4, the ABLs with higher $A_{w\theta}$ have minimum values of heat flux near the ABL top that are less negative than those of the homogeneous ABL (BA000). This less-negative minimum kinematic heat flux has been illustrated in some previous LES results (e.g., Avissar and Schmidt, 1998; Patton et al., 2005). Patton et al. (2005) suggested that this less-negative minimum value is caused by the spatial averaging of heat flux in a thicker capping inversion over a heterogeneous surface. We suggest that the temporal averaging of heat flux also be considered as a source for the less-negative minimum value, based on the significant temporal variability near the ABL top shown in Fig. 3.4b.

Fig. 3.5 shows the vertical profiles of total variances of $\theta$, $v$, and $w$ averaged over 1-4 h and the temporal variability of the total variances between 1-4 h. Here, it should be noted that the vertical profiles from the homogeneous ABL (BA000) are well matched with the well-known typical convective ABL shapes shown in previous research (e.g., Lenschow et al., 1980; Nieuwstadt et al., 1993). Whereas the variances of $\theta$ from the ABLs with lower $A_{w\theta}$ (BA025 and BA050) show approximately 2-5 times those of BA000, the variances from the ABLs with higher $A_{w\theta}$ (BA150, BA200, and BA250) demonstrate approximately 20 times those of BA000. This significant temporal variability implies that the quasi-steady state cannot be easily assumed for $\theta$, even in ABLs with lower $A_{w\theta}$. 
In Fig. 3.5b, the $v$ variances from the ABLs with higher $A_{w\theta}$ were significantly higher than those with lower $A_{w\theta}$ especially at lower and upper levels, which is a typical characteristic of the simulated ABL with a heterogeneous surface (e.g., Avissar and Schmidt, 1998; Patton et al., 2005). This feature is associated with the generated mesoscale horizontal flows. The temporal oscillations of these horizontal flows in the ABLs with higher $A_{w\theta}$ result in the large temporal standard deviation, which goes up to 60% of the averaged $v$ variance magnitude.

In Fig. 3.5c, the vertical profiles of $w$ variance from the ABLs with lower $A_{w\theta}$ (BA025 and BA050) are well matched with the profile from BA000, all having less than 5% temporal variability. ABLs with higher $A_{w\theta}$ (BA150, BA200, and BA250) have maximum values of $w$ variance that are 0.6-0.8 times the value of $w$ variance in BA000. These smaller variances are associated with the increased static stability that resulted from potential temperature advection, as shown in Fig. 3.3. The temporal oscillations of this temperature advection intensity cause the temporal variability of 5-24% in Fig. 3.5 cii.

3.5.2 (Co)spectral analysis

One-dimensional spectra of $\theta$, $v$, and $w$ are obtained by integrating two-dimensional spectral density ($\psi \kappa$) along the $x_2$ axis

$$F_\kappa (\kappa) = \int_{-\infty}^{\infty} \psi\kappa (\kappa_1, \kappa_2) d\kappa_2$$  \hspace{1cm} (3.28)
where \( c \) is \( \theta, v, \) or \( w, \) and \( \kappa_i (i=1,2) \) is defined as the wave number in the \( x_i \) direction.

The energy spectra of \( \theta, v, \) and \( w, \) presented in Figs. 3.6, 3.7, and 3.8 respectively are determined by computing one spectrum every 100 sec and then temporally averaging over 1-4 h.

In the spectra of \( \theta \) at 0.4 and 0.8 \( z_i, \) mesoscale surface differential forcing appears to have given rise to microscale variances, as seen by the difference in spectral density between BA000 and other cases at approximately 1 km wavelength in Fig. 3.6. This result implies that, without considering this nonlinear transfer of variance from the mesoscale, the microscale variance would be underestimated especially at the middle- and upper- level ABL. In fact, Kang et al. (2006) reveal that, in order to fit onto the mixed-layer similarity curve, a relatively coarse spatial filter is needed for aircraft-observed \( \theta \) variances when compared with vertical velocity and water vapor mixing ratio variances. In addition, the failure of the spectra normalized by mixed layer scaling to collapse into a single curve in observational studies (Kainal et al., 1976; Young, 1987) is likely associated with this enhanced \( \theta \) variance at the microscale (about 1 km) in response to mesoscale forcing.

In Fig. 3.7, the spectra of \( v \) from BA000 show peaks only at the wavelength of about 1.5 \( z_i. \) This result matches the observations of Kaimal et al. (1976) and Young (1987). The spectra of \( v \) from the heterogeneous ABL exhibit two peaks, one at the scale of 1.5 \( z_i \) and the other at the prescribed heterogeneity scale, here 32 km. This result is consistent with the wavelet spectrum results of Baidya Roy and Avissar (2000). The two
peaks are most obvious at 0.2 and 0.8 \( z_i \). The weaker peak at the heterogeneity scale at 0.4 \( z_i \) is likely associated with the absence of mesoscale flows in the middle levels.

In contrast to the \( v \) spectra, the \( w \) spectra from the heterogeneous ABL show no spectral peak at the scale of surface heterogeneity (Fig. 3.8). From the zero-divergence continuity equation, the mesoscale vertical velocity can be estimated as

\[
W \approx \frac{2z_i}{\lambda} V
\]

(3.29)

In Fig. 3.9, the \( w \) spectra from EA250 demonstrate relatively significant mesoscale peaks whereas the \( w \) spectra from BA250 reveal no obvious mesoscale peaks. Our scaling argument Eq. (3.29) implies that shorter wavelength heterogeneity should produce, all other factors held equal, stronger mesoscale vertical velocity at least comparable to turbulent vertical velocity. Note also that the \( w \) spectra in Fig. 3.9 were computed using outputs averaged over 6-9 h, instead of over 1-4 h (Fig. 3.6, 3.7, and 3.8). During this period, the time-averaged ABL depth is 1600 m, almost two times the time-averaged ABL depth of 850 m over 1-4 h. Following Eq. (3.29), this higher ABL depth also favors a stronger mesoscale vertical velocity. In summary, our vertical velocity spectra appear to be consistent with those of Letezel and Rassch (2003) showing distinct peaks on a larger scale than the turbulence scale in the \( w \) spectra computed using outputs from the run with the surface heat flux variation with a wavelength of \( \lambda = 16 \text{km} \), over 6-9 h.

In Figs. 3.7, 3.8, and 3.9, the inertial-range slopes are much steeper than the well-known Kolmogorov slope of -2/3. These steeper slopes are likely associated with the dissipation mechanism inherent in the integration scheme, working at scales smaller
than the effective resolution of about $6\Delta$ (Bryan et al., 2003; Skamarock, 2004). Thus the steeper slope can be overcome by using higher resolution. However, we are limited in this respect by computational resources and the need to simulate large spatial scales. It does not necessarily follow that our simulation of the dominant ABL eddies is flawed: these spatial scales should be well-resolved, as discussed previously. Note that the inertial-range slope of the $v$ spectra in Fig. 3.7 is steeper than that of the $w$ spectra in Fig. 3.8. This steeper slope of the $v$ spectra is associated with the use of the fifth-order advection scheme which has an inherent sixth-order filter that is proportional to wind speed (Bryan et al., 2003; Wicker and Skamarock, 2002).

Similar to spectra, one-dimensional cospectra are obtained by integrating two-dimensional cospectral density along the $x_1$ axis. In Fig. 3.10, the one-dimensional cospectra of $w$ and $\theta$ from the heterogeneous ABLs exhibit two peaks. However, unlike the spectra, these cospectra show the mesoscale peaks at the scale of 10-20 km, instead of at the prescribed heterogeneity scale. It should be also noticed that for the ABLs with higher $A_{w\theta}$ (BA150, BA200, and BA250), these mesoscale peaks become larger in magnitude than the microscale peaks, especially higher in the ABL. We will discuss the nature of this mesoscale flux in more detail in the coming sections.

3.5.3 Mesoscale temperature gradient and horizontal flows

Fig. 3.11 shows that initially the temporal evolution of the mixed-layer potential temperature gradient between the warmer and cooler regions follows Eq. (3.20) which assumes negligible temperature advection. The deviation from the estimated temperature
gradient from a certain time implies that the temperature gradient is reduced by the temperature advection. The magnitude of the temperature advection by the generated mesoscale flow is determined by the amplitude \(A_{w\theta}\) at a given wavelength \(\lambda\) of surface heat flux variation. The higher the amplitude is, the more significant the reduction of the temperature gradient. In addition the higher amplitude case, which produces a larger \(V\), starts to deviate from the estimation of Eq. (3.20) at an earlier time, which is expected from Eq. (3.23).

Fig. 3.12 shows that the maximum velocity of the generated mesoscale horizontal flow \(V_0\) and the time at which the horizontal velocity arrives at its maximum \(t_0\) are both functions of the amplitude \(A_{w\theta}\) and the scale of heterogeneity, \(\lambda\). These \(V_0\) and \(t_0\) values over the middle region at 0.2 \(z_i\) shown in Fig. 3.12 are summarized in Table 3.3. By using Eq. (3.27), one can estimate that the value of \(V_c\) is \(2.2\, ms^{-1}\) for \(\lambda = 32\, km\) and \(1.6\, ms^{-1}\) for \(\lambda = 16\, km\) when \(w_z = 1.8\, ms^{-1}\) and \(z_i = 860\, m\) (Table 3.2). In Fig. 3.12, one can observe that for cases where \(V_0/V_c \geq 1\), the horizontal velocity starts to temporally oscillate.

3.5.4 Variance decomposition

Next we wish to describe the simulations in terms of mesoscale versus microscale components, in addition to interaction between these scales. First, however, we need to identify a scale with which we can segregate phenomena into microscale and mesoscale components. To this end we have performed the variance decomposition Eq. (3.11) of \(\theta\),
v, and w, with four different cutoff wavelengths of $\lambda/16$, $\lambda/8$, $\lambda/4$, and $\lambda/2$. As shown in Fig. 3.13, the contribution of the interscale component $2\langle \phi \phi' \rangle$ to total variance $\langle \phi'^2 \rangle$ is more significant for $\theta$ than for $v$ and $w$. This result also implies a relatively large nonlinear transfer of mesoscale $\theta$ variance to a smaller scale variance, as discussed in section 3.5.2. However, when the cutoff wavelength is $\lambda/16$, the contribution of the interscale component to total variance is less than 5% for all three variables. We use this criterion, minimizing the interscale contribution to the variance, as a guideline for performing the remaining analyses with a cutoff wavelength of $\lambda/16$.

Fig. 3.14 shows the time-height cross sections of normalized mesoscale $\theta$, $v$, and $w$ variances for the two extreme cases in this study: BA025 and BA250. In BA025, whereas mesoscale $\theta$ variance is significantly larger than turbulent $\theta$ variance (Fig. 3.14ai), especially in the mid-ABL, mesoscale $v$ (and $w$) variances are much smaller than turbulent $v$ (and $w$) variances (Fig. 3.14bi and ci). In BA250, the temporally-oscillating mesoscale $\theta$ (and $v$) variances become dominant at the maxima of the mesoscale potential temperature gradient (mesoscale horizontal velocity) (Fig. 3.14aii and bii). However, even in BA250, mesoscale $w$ variance is less significant than turbulent $w$ variance. At the maxima of the mesoscale horizontal velocity, the contribution of mesoscale $w$ variance peaks at 20% of the total vertical velocity variance.

Fig. 3.15 shows that the time-height cross sections of normalized microscale $\theta$, $v$, and $w$ variances. These microscale variances are each normalized by the appropriate scaling parameter. The microscale variances from BA025 demonstrate a quasi-steady state ABL, whereas those from BA250 reveal an ABL that is not in a quasi-steady state.
Fig. 3.16 shows the vertical profiles of turbulent variances averaged over 1-4 h and the temporal variability of these turbulent variances between 1-4 h. The vertical profiles of turbulent $\theta$ variances from the ABLs with lower $A_{w,\theta}$ (BA025 and BA050) are nearly coincident with those from BA000. The temporal variability, seen in Fig. 3.16a(ii), indicates that BA025 and BA050 may satisfy the quasi-steady state condition. However, for the ABLs with higher $A_{w,\theta}$ (BA150, BA200, and BA250), the vertical profiles of the microscale $\theta$ variances deviate somewhat from BA000. In addition, the temporal variability is much larger than that in BA000, especially at middle and upper levels. Although the microscale $v$ variances from the ABLs with lower $A_{w,\theta}$ are somewhat larger than those from BA000, the shapes of the vertical profiles are similar to those from BA000. However, the microscale $v$ variances from the ABLs with higher $A_{w,\theta}$ demonstrate values two times those from BA000 at 0.4-0.6 $z_i$. This seems to be associated with the larger transfer of mesoscale variance to the microscale at this middle level as compared to that at the lower and upper levels (Fig. 3.7). In Fig. 3.16c, the profiles of microscale $w$ variance and their temporal variability are almost the same as those of total variances (Fig. 3.5c) due to the insignificant contributions of mesoscale variance and the interscale component to total variance (Figs. 3.13, and 3.14).

3.5.5 Decomposition of vertical heat flux

Because of the negligible contribution to total heat flux, the interscale term involving $w^M$ and $\theta^r$ in Eq. (3.12) is disregarded. For cases BA025 and BA250, the
time-height cross sections of the other three terms in Eq. (3.12) are plotted in Fig. 3.17. In BA025, the turbulent vertical heat flux appears to be in a quasi-steady state (Fig. 3.17bi). In contrast, in BA250, the turbulent vertical heat flux is not in a quasi-steady state. Approximately 30 min. after each time the interscale term \( \langle w^* \theta^M \rangle \) and mesoscale term \( \langle w^M \theta^H \rangle \) reach their maxima, the microscale vertical heat flux is reduced in the middle- and upper-level ABL. Although, at their maxima, this interscale term becomes comparable to the microscale heat flux, the microscale heat flux is still the most significant contributor to total heat flux. Fig. 3.18a shows that the microscale heat fluxes from the ABLs with higher \( A_{w \theta} \) are half of those from BA000, especially at the middle-levels of the ABL. The similar linear shapes of the vertical profiles of total heat flux as compared to that from BA000 in Fig. 3.4a result from the increased magnitude of the interscale component \( \langle w^* \theta^M \rangle \), when the microscale heat fluxes are decreased. In Fig. 3.18b, the significant temporal variability of the microscale vertical heat flux in the ABLs with higher \( A_{w \theta} \) suggests that the microscale flux is substantially modulated as a function of time in the ABLs experiencing strong oscillations in the mesoscale flows.

3.6 Conclusions

A recently developed, compressible, nonhydrostatic numerical model has been used as an LES to simulate the fair-weather midday ABL (11-14 LST) with a heterogeneous surface thermal forcing. This surface heterogeneity is prescribed by surface heat flux variation that is sinusoidal as a function of mean value, wavelength, and
amplitude. The mean value is fixed at 0.21 Kms$^{-1}$, the wavelengths are 16 and 32 km (20 and 40 $z_i$ respectively), and the amplitude of surface heat flux variation has seven prescribed values ranging from 0 to 0.2 Kms$^{-1}$.

The principal mechanism governing the surface-heterogeneity-induced mesoscale horizontal flows is understood by using $[u_t]$-momentum and $[\theta]$ equations, which are theoretically obtained. By scaling these two equations, we estimate the magnitude of the generated mesoscale horizontal flow that will be guaranteed to initiate temporal oscillations. For the wavelengths of 32 and 16 km, the minimum magnitude of the mesoscale horizontal flow needed is computed to be 2.1 and 1.5 ms$^{-1}$, respectively. Mesoscale horizontal winds greater than these flows are generated by amplitudes of surface heat flux variation greater than 0.12 Kms$^{-1}$ for the wavelength of 32 km and greater than 0.10 Kms$^{-1}$ for the wavelength of 16 km (Table 3.2 and 3.3).

The ABL structure is significantly modified by the mesoscale horizontal flows that are generated by surface heat flux variations with amplitudes of 0.10 Kms$^{-1}$ or higher. In this heterogeneous ABL, the assumption of a quasi-steady state cannot be applied due to the temporal oscillation of the generated mesoscale horizontal flow. In addition, the temperature advection (cold advection at lower levels and warm advection at upper levels) by this mesoscale flow reduces buoyancy throughout the ABL.

These temporally oscillating ABLs can be further categorized into two types: with and without significant mesoscale vertical velocities. The sea-breeze-like mesoscale circulation, which is accompanied by substantial mesoscale peaks in the vertical velocity spectra, is generated only when the amplitude is the highest used in this experiment (0.2 Kms$^{-1}$), the wavelength is the shorter of the two utilized (16 km), and the ABL depth
grows to twice the initial depth (1600 m) (Fig. 3.9b). The relation between the mesoscale vertical velocity and ABL characteristics is consistent with a simple scale analysis of the zero-divergence continuity equation.

For heterogeneous ABLs, with a low surface flux variation amplitude, 0.02-0.04 Kms\(^{-1}\), similar to those observed during IHOP_2002 (Table 3.1; Kang et al., 2006), the microscale variances and vertical heat fluxes are nearly identical to the values from a homogeneous and stationary ABL when filtered with an appropriate wave-cutoff filter (Fig. 3.16). However, even in this weakly heterogeneous ABL, the significance of the nonlinear transfer of variance from the imposed heterogeneity scale, \(\lambda\), to a smaller scale is important for \(\theta\), and somewhat important for \(v\). Vertical velocity variances are perturbed the least from the homogenous case.

The ABL with surface heat flux variation amplitude of 0.08 Kms\(^{-1}\) is denoted as an intermediate amplitude case. Compared with the high amplitude cases, the temporally-oscillatory pattern is not clearly seen for this case (Fig. 3.12). The microscale variances and vertical heat fluxes obtained with the cutoff wavelength of \(\lambda/16\), however, are not nearly identical to the values from a homogeneous and stationary ABL.

For every simulated ABL in this numerical experiment, microscale heat flux is the most significant contributor to total heat flux. For the ABLs with the highest amplitude surface heterogeneity, the contribution of the interscale component of the heat flux \(\langle w^* \theta^M \rangle\), becomes comparable to that of the microscale heat flux at the times when the periodic mesoscale heat flux, which is one order smaller than the turbulent flux due to the negligible mesoscale vertical velocities, reaches its maxima. This significant interscale
component also explains why the mesoscale peak in the cospectra of \( w \) and \( \theta \) (Fig. 3.10) is located at a smaller scale than the imposed heterogeneity scale.

Similar to mesoscale model studies, previous LES research has primarily focused on the sea-breeze-like mesoscale circulations and their effects on the ABL structure. However, the amplitudes of the surface heat flux variation required to generate these temporally oscillatory mesoscale flows appear to be extremely high compared to the observed amplitudes (Table 3.1). In order to make LES results more relevant to larger-scale models, ABLs with low and intermediate amplitudes, which likely occur most frequently, require additional investigation. While these ABLs do not exhibit dramatic, oscillating mesoscale flows, ABL characteristics still depart from the similarity theory commonly employed in ABL parameterization. Generalization of such modification of ABL structure over heterogeneous surfaces should lead to better representation of the ABL in larger-scale models.

3.7 Appendix: Proofs of \([f(x)][f(x)]''=0\) and \([[f(x)][f(x)]] = [f(x)][f(x)]\) used in section 4a

First, we represent a function \( f(x) \) as a complex Fourier series,

\[
f(x) = \sum_{n=1}^{N} \hat{f}(\kappa_n) e^{i\kappa_n x},
\]

where \( \hat{f} \) are coefficients which are complex numbers. This function can be decomposed into low-pass filtered components,

\[
[f(x)] = \sum_{n=1}^{c} \hat{f}(\kappa_n) e^{i\kappa_n x}
\]

(where \( c \) is the cutoff wavelength) and high-pass filtered components,
\((f(x))'' = \sum_{n=c+1}^{N} \hat{f}(\kappa_n) e^{i\kappa_n x}\), by the wave-cutoff filter\([\ ]\). The product of these low- and high-passed parts can be written as

\[
[f(x)][f(x)]'' = \hat{f}(\kappa_1) \hat{f}(\kappa_{c+1}) e^{i(\kappa_1 + \kappa_{c+1}) x} + \cdots + \hat{f}(\kappa_c) \hat{f}(\kappa_N) e^{i(\kappa_c + \kappa_N) x}.
\]

Applying the wave-cutoff filter to this product yields

\[
[[f(x)][f(x)]'] = \left[ \hat{f}(\kappa_1) \hat{f}(\kappa_{c+1}) e^{i(\kappa_1 + \kappa_{c+1}) x} + \cdots + \hat{f}(\kappa_c) \hat{f}(\kappa_N) e^{i(\kappa_c + \kappa_N) x} \right] \tag{3.30}
\]

In (A.1), if \(\kappa_1 + \kappa_{c+1} > \kappa_c\) then \([f(x)][f(x)]'' = 0\). The condition \(\kappa_1 + \kappa_{c+1} > \kappa_c\) is always true because \(\kappa_{c+1} > \kappa_c\) and \(\kappa_1 > 0\).

Second, the product of two low-pass filtered components of the function \(f(x)\)

\[
[f(x)][f(x)] = \hat{f}(\kappa_1) \hat{f}(\kappa_{c+1}) e^{i(\kappa_1 + \kappa_{c+1}) x} + \cdots + \hat{f}(\kappa_c) \hat{f}(\kappa_N) e^{i(\kappa_c + \kappa_N) x}.
\]

Applying the wave-cutoff filter to this product yields

\[
[[f(x)][f(x)]] = \left[ \hat{f}(\kappa_1) \hat{f}(\kappa_{c+1}) e^{i(\kappa_1 + \kappa_{c+1}) x} + \cdots + \hat{f}(\kappa_c) \hat{f}(\kappa_N) e^{i(\kappa_c + \kappa_N) x} \right] \tag{3.31}
\]
Fig. 3.1: Schematic illustration of the mesoscale flows in the ABL with the surface heat flux variation prescribed as a simple sinusoidal function (1). The thick circular arrows represent the mesoscale flows and the thin dash-dot line the surface heat flux variation. Based on the characteristics of the mesoscale potential temperature and the mesoscale wind velocity distributions expected from the sinusoidal surface heat flux variation, the ABL is divided into two pairs of regions: 1) warmer region \((0 < x_1 < \frac{\lambda}{2})\) and cooler region \((\frac{\lambda}{2} < x_1 < \lambda)\), 2) middle region \((\frac{\lambda}{4} < x_1 < \frac{3\lambda}{4})\) and edge region \((\frac{3\lambda}{4} < x_1 < \lambda \text{ and } 0 < x_1 < \frac{\lambda}{4})\).
Fig. 3.2: The filter response functions of the wave-cutoff filters used for this study. Here $\kappa_i$ is the wavenumber in the $x_i$ direction and $\kappa_f$ is a wave-cutoff filter.
Fig. 3.3: Vertical profiles of domain-averaged potential temperature at each hour between 0-240 min for the cases with $\lambda=32$ km.
Fig. 3.4: Vertical profiles of (a) total vertical heat flux calculated at each model output time (every 100 seconds) and averaged over 60-240 min and (b) temporal variability, defined as the standard deviation of total vertical heat flux over 60-240 min. This temporal variability is normalized by the magnitude of the total vertical heat flux averaged over 60-240 min. In (a), the thin lines that remain approximately zero above 0.1 \(< z/< z_i >\) indicate the contributions of subfilter-scale heat flux \(f_{w\theta}\). Here < > indicates domain average and \{\} time average over 60-240 min.
Fig. 3.5: Vertical profiles of (i) total variances of (a) potential temperature $\theta$, (b) horizontal streamwise velocity $v$, and (c) vertical velocity $w$, and (ii) their temporal variability. The temporal variability is defined as the standard deviation of total variance over 60-240 min. This temporal variability is normalized by the magnitude of the total variance averaged over 60-240 min. The thin lines in (bi) and (ci) indicate the contributions from the subfilter-scale parameterization ($2/3e$, where $e$ is a subfilter-scale TKE). Here, $<>$ represents domain average and $\{\}$ time average over 60-240 min.
Fig. 3.6: One-dimensional spectra of potential temperature at (a) 0.8, (b) 0.4, and (c) 0.2 $z_i$. These spectra were computed using output every 100 s for 180 min (60-240 min).
Fig. 3.7: One-dimensional spectra of the horizontal streamwise velocity at (a) 0.8, (b) 0.4, and (c) 0.2 $z_i$. These spectra were computed using output every 100 s for 180 min (60-240 min). A short thin line represents a -2/3 spectrum.
Fig. 3.8: One-dimensional spectra of the vertical velocity at (a) 0.8, (b) 0.4, and (c) 0.2 $z_i$. These spectra were computed using output every 100 s for 180 min (60-240 min). A short thin line represents a -2/3 spectrum.
Fig. 3.9: One-dimensional spectra of the vertical velocity at the four normalized heights for cases (a) EA250 and (b) BA250. These spectra were computed using output every 100 s for 180 min (360-540 min). A short thin line represents a -2/3 spectrum.
Fig. 3.10: One-dimensional cospectra of the vertical velocity and potential temperature at (a) 0.6, (b) 0.4, and (c) 0.2 $z_i$. A short thin line represents a $\kappa^{-7/3}$ cospectrum (Wyngaard and Cote, 1972).
Fig. 3.11: The temporal evolutions of the mixed-layer potential temperature gradient between the warmer and cooler regions (Fig.3.1). Herein, the potential temperature is horizontally averaged over each region and then vertically averaged over 0.2-0.8 $z_i$, where $z_i$ is the ABL depth. The thin lines indicate the estimation of (3.20) which is based on the assumption of negligible mesoscale flows.
Fig. 3.12: The time evolutions of the magnitudes of horizontal streamwise wind (v) averaged over the middle region at 0.2 \( z_i \). These results come from simulations with (a) \( \lambda = 32 \text{ km} \) and (b) \( \lambda = 16 \text{ km} \).
Fig. 3.13: Contribution of the interscale component in the variance decomposition of (a) potential temperature ($\theta$), (b) horizontal streamwise velocity ($\nu$), and (c) vertical velocity ($w$). Here, the contribution, $2\langle \phi^M \phi^* \rangle$, is normalized by the total variance, $\langle \phi^2 \rangle$, which results in $2\langle \phi^M \phi^* \rangle/\langle \phi^2 \rangle$, where $\phi$ can be $\theta$, $\nu$, or $w$. The error bar indicates one standard deviation over 0.2-0.8 $z_i$ and 60-240 min.
Fig. 3.14: Time-height cross sections of mesoscale variances of (a) $\theta$, (b) $v$, and (c) $w$ for cases (i) BA025 and (ii) BA250. Here the cutoff wavelength of $\lambda/16$ is used to define mesoscale variances.
Fig. 3.15: Time-height cross sections of normalized (a) $\theta$, (b) $v$, and (c) $w$ turbulence variances for cases (i) BA025 and (ii) BA250. Here the cutoff wavelength of $\lambda/16$ is used to define turbulence variances. Regarding normalization, the turbulence variance of $\theta$ is normalized by $\theta^2$ and the variances of $v$ and $w$ are normalized by $w^2$. Here $\theta_*$ is the mixed-layer temperature scale and $w_*$ the free-convection scaling velocity (Stull, 1988).
Fig. 3.16: Vertical profiles of (i) turbulent variances and (ii) temporal variability of (a) potential temperature, (b) horizontal streamwise velocity, and (c) vertical velocity. Here { } represents the time average over 60-240 min. The thin line represents the subfilter-scale parameterization contribution for each experiment. The temporal variability is defined as the temporal standard deviation of turbulence variance normalized by the average over 60-240 min.
Fig. 3.17: Time-height cross sections of the three vertical heat flux components in (3.12): (a) $\langle w^M \theta^M \rangle$, (b) $\langle w^M \theta^W \rangle$, and (c) $\langle w^M \theta^R \rangle$ for cases (i) BA025, and (ii) BA250.
Fig. 3.18: Vertical profiles of (a) turbulent vertical heat flux and (b) temporal standard deviation of the vertical heat flux normalized by the absolute value of the turbulent vertical heat flux. The thin line represents the subfilter-scale parameterization contribution for each experiment.
Table 3.1: The mean and maximum values of the sensible heat flux difference between flux tower site 1 and site 2 for four of the five days\(^1\) studied by Kang et al. (2006), which presented the monotonically varying surface heat flux variation over the 16-km transect between site 1 and site 2 during the five days. These surface heat fluxes are the averaged values over 1130-1430 LST. The number in parentheses represents the value for all the fair-weather days\(^2\) during IHOP.

<table>
<thead>
<tr>
<th>Cases</th>
<th>Number of days</th>
<th>Mean ((Wm^{-2}))</th>
<th>Maximum ((Wm^{-2}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>(H_2 &gt; H_1)</td>
<td>3(12)</td>
<td>58(44)</td>
<td>105(131)</td>
</tr>
<tr>
<td>(H_1 &gt; H_2)</td>
<td>1(10)</td>
<td>26(36)</td>
<td>26(68)</td>
</tr>
</tbody>
</table>

\(^1\)Herein, one of the five observation days, specifically 25 May, cannot be considered due to missing data.

\(^2\)A fair-weather day is defined as having a mean downward solar radiation of 950 Wm\(^{-2}\) or higher during 1130-1430 UTC both at site 1 and site 2.
Table 3.2: Experimental design parameters for the prescribed surface heat flux variation and model output mixed-layer scaling parameters. The domain-averaged value of surface heat flux is set at $\langle F_{w\theta} \rangle(z=0) = 0.21 \text{Kms}^{-1}$.

<table>
<thead>
<tr>
<th>Case</th>
<th>$\lambda$ (km)</th>
<th>$A_{w\theta}$ ($\text{Kms}^{-1}$)</th>
<th>$\langle z_i \rangle (m)$</th>
<th>$w_s (\text{ms}^{-1})$</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td>t=60</td>
<td>t=240</td>
</tr>
<tr>
<td>BA000</td>
<td>32</td>
<td>0.00</td>
<td>760</td>
<td>958</td>
</tr>
<tr>
<td>BA025</td>
<td>32</td>
<td>0.02</td>
<td>760</td>
<td>953</td>
</tr>
<tr>
<td>BA050</td>
<td>32</td>
<td>0.04</td>
<td>760</td>
<td>939</td>
</tr>
<tr>
<td>BA100</td>
<td>32</td>
<td>0.08</td>
<td>759</td>
<td>897</td>
</tr>
<tr>
<td>BA150</td>
<td>32</td>
<td>0.12</td>
<td>758</td>
<td>883</td>
</tr>
<tr>
<td>BA200</td>
<td>32</td>
<td>0.16</td>
<td>757</td>
<td>919</td>
</tr>
<tr>
<td>BA250</td>
<td>32</td>
<td>0.20</td>
<td>761</td>
<td>934</td>
</tr>
<tr>
<td>EA000</td>
<td>16</td>
<td>0.00</td>
<td>762</td>
<td>957</td>
</tr>
<tr>
<td>EA025</td>
<td>16</td>
<td>0.02</td>
<td>762</td>
<td>945</td>
</tr>
<tr>
<td>EA050</td>
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<td>0.04</td>
<td>761</td>
<td>926</td>
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<tr>
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<td>0.08</td>
<td>760</td>
<td>925</td>
</tr>
<tr>
<td>EA150</td>
<td>16</td>
<td>0.12</td>
<td>760</td>
<td>923</td>
</tr>
<tr>
<td>EA200</td>
<td>16</td>
<td>0.16</td>
<td>761</td>
<td>924</td>
</tr>
<tr>
<td>EA250</td>
<td>16</td>
<td>0.20</td>
<td>764</td>
<td>944</td>
</tr>
</tbody>
</table>
Table 3.3: Maximum values of the horizontal streamwise velocity $V_0$ in Fig. 12 and the time at which the horizontal velocity reaches its maximum $t_0$. Here $V_c$ is estimated as 2.1 $ms^{-1}$ for $\lambda=32$ km and 1.5 $ms^{-1}$ for $\lambda=16$ km by using (22).

<table>
<thead>
<tr>
<th></th>
<th>$t_0$ (min)</th>
<th>$V_0$ (ms$^{-1}$)</th>
<th>$V_0/V_c$</th>
</tr>
</thead>
<tbody>
<tr>
<td>BA025</td>
<td>505</td>
<td>0.5</td>
<td>0.2</td>
</tr>
<tr>
<td>BA050</td>
<td>288</td>
<td>0.7</td>
<td>0.3</td>
</tr>
<tr>
<td>BA100</td>
<td>255</td>
<td>1.4</td>
<td>0.7</td>
</tr>
<tr>
<td>BA150</td>
<td>198</td>
<td>1.9</td>
<td>0.9</td>
</tr>
<tr>
<td>BA200</td>
<td>163</td>
<td>2.4</td>
<td>1.1</td>
</tr>
<tr>
<td>BA250</td>
<td>142</td>
<td>2.7</td>
<td>1.3</td>
</tr>
<tr>
<td>EA025</td>
<td>313</td>
<td>0.5</td>
<td>0.3</td>
</tr>
<tr>
<td>EA050</td>
<td>535</td>
<td>0.9</td>
<td>0.6</td>
</tr>
<tr>
<td>EA100</td>
<td>152</td>
<td>1.1</td>
<td>0.7</td>
</tr>
<tr>
<td>EA150</td>
<td>107</td>
<td>1.6</td>
<td>1.1</td>
</tr>
<tr>
<td>EA200</td>
<td>100</td>
<td>1.9</td>
<td>1.3</td>
</tr>
<tr>
<td>EA250</td>
<td>83</td>
<td>2.2</td>
<td>1.5</td>
</tr>
</tbody>
</table>
Chapter 4

Moist ABL Simulation

4.1 Chapter Introduction

Accurately describing and forecasting moist convection is one of the most demanding and challenging issues in the field of atmospheric sciences. Moist convection has a variety of forms, from stratocumulus to deep precipitating cumulus, which are associated with the diversity of physical processes air parcels would experience (Stevens, 2005). Thus, various forms of moist convection are sensitive to the properties of the atmospheric boundary layer (ABL). For example, Weckwerth et al. (1996), analyzing data collected by radar, aircraft, and sounding during the Convective and Precipitation/Electrification (CaPE) field experiment, concluded that the horizontal variability of moisture in the convective ABL is closely associated with the potential for deep, moist convection. Numerous studies have suggested that the ABL is a key component of the processes related to cloud development and precipitation in both global circulation models (GCMs) and in mesoscale models (MMs) (e.g., Wisse and Arellano, 2004; Byun and Hong, 2004; Martin et al., 2000; Basu et al., 2002).

Many descriptions of the ABL assume spatial homogeneity (Arya, 2001; Stull, 1988). The ABL by definition is directly influenced by the earth’s surface which is
always heterogeneous. Thus, many studies have examined the effect of surface heterogeneity on the ABL (for a literature review, see Pielke, 2001) with the aim of a more realistic description of the relationship between the earth’s surface and moist convection. Much of this research has utilized MMs and has focused on the influence of sea-breeze-like mesoscale circulations on the initiation and development of moist convection (e.g., Chen and Avissar, 1994b; Weaver and Avissar, 2001). Chen and Avissar (1994b), using surface flux conditions representing an extreme contrast at the boundary of irrigated/non-irrigated land surfaces, suggested that heterogeneous surface conditions generate sea-breeze-like mesoscale circulations and significantly affect the timing of cloud formation and the intensity and distribution of precipitation. Weaver and Avissar (2001) suggested that negligence of mesoscale vertical fluxes, the product of mesoscale fluctuations of vertical velocity and a scalar (Chen and Avissar, 1994a), may cause a serious failure in GCMs used to predict climate change associated with human modification of the earth’s surface.

While many studies using MMs have suggested a significant relationship between surface heterogeneity and moist convection, other studies disagreed (e.g., Zhong and Doran, 1997). Using a MM with observation-based surface flux distributions instead of idealized extreme surface flux contrasts, Zhong and Doran (1997) have argued that the spatial variability of the ambient meteorology is more significant than that of surface fluxes in determining cloud amount and favorable places for cloud formation. Zhong and Doran (1998) additionally found that profiles of potential temperature and water vapor mixing ratio, averaged over a model domain of 300 km × 350 km, which is similar to a typical grid cell of a GCM, showed little difference whether simulated with spatially
varying fluxes or uniform fluxes. Thus, they concluded that the failure to include the
effects of mesoscale circulations in a coarse-resolution model such as a GCM is less
consequential than other scientists (e.g., Chen and Avissar, 1994b; Weaver and Avissar,
2001) have suggested. However, on the length scale of tens of kilometers, similar to a
typical grid cell of a MM, instead of hundreds of kilometers, Zhong and Doran (1997,
1998) also noted a significant response of the ABL to surface heterogeneity.

MM, however, may not be appropriate to investigate the response of the ABL to
land surface heterogeneity at a scale of 10 km. Considering an effective resolution of 6Δ
(where Δ represents a grid spacing; Bryan et al., 2003), the grid spacing to explicitly
resolve the flows on the order of 10 km should be at a scale on the order of 1 km. Such a
1-km grid spacing may approach the l/Δ~1 range (where l is the scale of the energy-
containing turbulent eddies and typically 1.5 times the ABL depth), which is outside of
the range of scale for which MMs have been designed (Wyngaard, 2004). Large eddy
simulation (LES) is one modeling approach that can reconcile this problem. Since LES
can explicitly resolve energy-containing eddies at least in the convective ABL, LES can
be an appropriate tool to study the effects of surface heterogeneity at a scale of 10 km.

Although some studies (e.g., Avissar and Schmidt, 1998; Patton et al., 2005) have
used LES to simulate the ABL over a heterogeneous surface at these smaller scales, they
have mainly focused on determining the scale of surface heterogeneity that is optimal for
generating sea-breeze-like circulations, not on investigating the modification of ABL
properties by the surface heterogeneity. Kang and Davis (2006) used LES to investigate
the modification of dry ABL structure by 10-km scale surface heterogeneity. Thus, they
did not explore the modification of moisture-related properties of the ABL by surface heterogeneity and its potential influence on moist convection.

In this study, we simulate moist ABLs over heterogeneous surfaces using the LES (Bryan and Fritsch, 2002) used in Kang and Davis (2006). We investigate two aspects of these simulated fair-weather moist ABLs (1130-1430 LST): altered properties of the moist ABL over a heterogeneous surface, and the horizontal transport of heat and moisture by surface-heterogeneity–induced mesoscale flows.

The following section describes the LES used here and its setup. Section 4.3 presents the spectral and cospectral analysis results. In Section 4.4, a decomposition method is used to separate fluctuations of a variable into mesoscale and miscro-scale (turbulent) fluctuations. In Section 4.5, the joint frequency distribution (JFD) technique is applied to analyze both the mesoscale and miscro-scale fluctuations. In Section 4.6, the significance of the horizontal transport of heat and moisture by surface-heterogeneity-induced mesoscale flows is examined. Section 4.7 summarizes the results and presents future plans.

**4.2 Numerical Experiment**

4.2.1 Model description and setup

In this study, the compressible nonhydrostatic numerical model of Bryan and Fritsch (2002) is used as an LES. The model, which was originally developed as a cloud-
resolving model using LES techniques, has been applied to investigate the appropriate spatial resolution for the simulation of deep moist convection from an LES perspective (Bryan et al., 2003). Kang and Davis (2006) have demonstrated that the Bryan-Fritsch model can reproduce turbulence statistics of the ABL over a homogeneous land surface, suggesting that is can be used as an LES for convective ABL study.

Our model domain is 32 km long in the $x_1$ (here, north-south and streamwise) direction, and 5 km long in the $x_2$ (here, west-east and crosswind) direction. The grid spacing in the horizontal is 100 m. The vertical extent of the domain is 3.5 km. The grid spacing in the vertical is 10 m up to the height of 100 m, linearly increases from 10 m to 40 m between the height of 100 m and 1900 m, and then remains constant at 40 m up to the model top at 3500 m. Thus, there are (320, 50, 122) grid points in the $(x_1, x_2, z)$ directions. In both horizontal directions the lateral boundary conditions are periodic. The upper boundary is a flat, rigid wall with a Rayleigh damping layer (Durran and Klemp, 1983) occupying 1 km beneath the model top. The lower boundary is also a flat, rigid surface. The Bryan-Fritsch model does not include atmospheric radiation, which indicates that the effect of clouds on the surface fluxes is not considered. In this study surface heat and moisture fluxes are fixed during the integration period as prescribed in the next section. Unlike the prescribed surface heat and moisture flux, surface momentum flux is derived from a simple surface drag parameterization (Stull, 1988). The surface heat and moisture fluxes are activated upon initiation of the simulation. Also, at the start of the simulation, random perturbations of 0.1 K are superimposed on the potential temperature at the lowest atmospheric level, which allows three-dimensional turbulent flows to
develop quickly. The microphysics selected for this study is the simple bulk microphysics scheme of Kessler (1969), which considers only water. Thus the Kessler scheme produces three mixing ratios: water vapor, cloud water, and rainwater. Considering that a model spin-up time of approximately 30-60 min is expected and this is a midday ABL simulation (1130-1430 UTC), the model is run for 4 h. Thus, the discussion is based on results from the model run between 60 and 240 min. For all cases the integration time step is set to 1 sec and the model output was saved every 100 seconds.

4.2.2 Experimental Design

We shall prescribe a surface heat flux variation that is sinusoidal with mean value \( \langle F_{w\theta} \rangle \), amplitude \( A_{w\theta} \), and wavelength \( \lambda \)

\[
F_{w\theta}(x_1) = \langle F_{w\theta} \rangle + A_{w\theta} \sin\left(\frac{2\pi}{\lambda} x_1\right) \tag{4.1}
\]

Surface moisture flux variation is prescribed with a phase lag of \( \lambda/2 \) relative to that of the surface heat flux

\[
F_{wr}(x_1) = \langle F_{wr} \rangle - A_{wr} \sin\left(\frac{2\pi}{\lambda} x_1\right) \tag{4.2}
\]

The simulations performed with various values of \( \langle F_{w\theta} \rangle \), \( \langle F_{wr} \rangle \), \( A_{w\theta} \), and \( A_{wr} \) are summarized in Table 4.1. The wavelength \( \lambda \) in Eqs. (4.1) and (4.2) is chosen to be 32
km to investigate the heterogeneous ABL with surface forcing at a scale one order of magnitude greater than the ABL depth. \( F_{w\theta} \) is set at 0.19 \( Kms^{-1} \) and \( F_{wr} \) is set at 0.033 \( gkg^{-1}ms^{-1} \), which produce a Bowen ratio of 2.3. By controlling \( F_{wr} \), the Bowen ratio is varied from 0.3 to 3.3.

The ratio (\( R \)) of horizontal variation (magnitude) of latent heat (\( LE \)) to sensible heat (\( H \)) is changed from 1 to 2 by varying \( A_{wr} \). LeMone et al. (2006) suggested that the slope of the relative amplitudes of \( H \) and \( LE \) horizontal variability (\( R \equiv \Delta LE/\Delta H \), where \( \Delta \) represents horizontal variability) is steeper under wetter soil conditions. In other words, the slope is steepest just after rainfall and then becomes shallower with time (LeMone et al., 2006). Thus, we set two different surface conditions to represent short and long times after rainfall by prescribing \( R \) to be 2 and 1, respectively.

Fig. 4.1 shows the initial profiles of potential temperature and water vapor mixing ratio for the cases summarized in Tables 4.1 and 4.2. This sounding is based on the rawinsonde released over the Homestead site (36.55 °N, 100.6 °N) at about 1130 LST on 25 May during the International H2O Project (IHOP_2002; Weckwerth et al., 2004). Since the sounding shows no convective available potential energy (CAPE), conditions are not favorable for the initiation and development of deep, moist convection. Despite initial conditions unfavorable for deep convection, the cases with \( B=0.3 \) develop clouds and precipitation. In Tables 4.1 and 4.2, the maximum and total values of cloud water (and rain water, if there is precipitation) are noted. The effect of the clouds on ABL structure is ignored because the cloud is located above the ABL. Also radiative process,
which may influence the surface energy budget in the cloudy ABL, is absent in this study.

4.3 Spectral analysis

The effect of the imposed mesoscale surface heterogeneity on ABL structure is investigated using spectral analysis. The spectra of potential temperature ($\theta$), water vapor mixing ratio ($r$), and vertical velocity ($w$) are computed to measure the relative significance of the surface heterogeneity effect at each scale for each variable. Similarly, the cospectra are computed to estimate the relative significance of the correlations between $\theta$ and $r$ ($\theta - r$), $\theta$ and $w$ ($\theta - w$), and $r$ and $w$ ($r - w$) at each scale and to identify the physical processes associated with the correlations at each scale.

4.3.1 Why MR (co) spectra?

The MR (co)spectrum is an alternative method to Fourier (co)spectrum. Unlike the Fourier (co)spectrum using a sinusoidal basis set, the MR (co)spectrum is directly linked to Reynolds averaging because it uses a wavelet basis set with a constant basis function (Vickers and Mahrt, 2003). Howell and Mahrt (1997) said that the MR spectrum more precisely decomposes the turbulent variances and fluxes as a function of scale than a Fourier spectrum does. By applying this MR spectrum to the low-level aircraft data collected over a heterogeneous surface during the International H2O Project
(IHOP_2002; Weckwerth et al., 2004), Kang et al. (2006) identified the existence of surface-heterogeneity-induced mesoscale flows in the ABL over a heterogeneous surface. This study uses the MR spectrum to compare LES results with these observational findings.

The MR (co)spectra are obtained by computing one spectrum every 100 s and then temporally averaging over 60-240 min. The (co)spectra are normalized by the total absolute values of (co)spectral density at each segment to see the relative significance of the contribution at each scale without changing the signs of the (co)spectra density. The streamwise \((x_1;\text{north-south})\) dimension of 320 \(= 2^8 + 64\) grid points is extended into 512 \(= 2^9\) points by adding 96 grid points at each end from the opposite end. This extension complies with the computational requirement of precisely \(2^M\) data elements (where \(M\) is the largest integer such that \(2^M < N\); \(N\) is the length of the data samples). This extension is feasible, given the periodic lateral boundary conditions used in these numerical simulations.

4.3.2 MR spectral analysis

In Fig. 4.2, the vertical velocity \((w)\) spectra demonstrate that spectral density peaks exist only at the microscale (or at the turbulence scale), namely about 1.5 \(z_i\) (the ABL depth), both in the ABLs over homogeneous surfaces and over heterogeneous surfaces. Even for case MB200W2, which used the highest amplitude of surface heat flux variation in this numerical study, the spectral density peaks are absent at the imposed
surface heterogeneity scale (32 km). The absence of mesoscale peaks in the $w$ spectra is consistent with both the MR spectra from the observed ABLs over a heterogeneous surface during IHOP_2002 (Kang et al., 2006), and the Fourier spectra from dry ABLs simulated with a similar numerical scheme (Kang and Davis, 2006). The negligible mesoscale vertical velocities result in insignificant mesoscale vertical fluxes, the product of mesoscale fluctuations of $w$ and a scalar (Chen and Avissar, 1994a).

Unlike the $w$ spectra, the spectra of water vapor mixing ratio ($r$), potential temperature ($\theta$), and horizontal velocity ($v$) from the ABLs over heterogeneous surfaces exhibit spectral density peaks at the imposed surface heterogeneity scale, namely 32 km. In Fig. 4.3, the $r$ spectra from the ABL over a heterogeneous surface present spectral density peaks at 32 km, but the $r$ spectra from the homogeneous ABL (MB000) do not. In the $r$ spectra at $0.8 \langle z_i \rangle$ (where $\langle z_i \rangle$ represents a domain-averaged value of the ABL depths), for cases MB025 and MB050, the spectral density peaks at the mesoscale are less than those at the microscale. However, for cases MB100 and MB200W, the spectral peaks at the microscale are greater than those at the mesoscale. These $r$ spectra imply that the contribution of turbulence to entrainment from the free atmosphere is likely more significant than that of mesoscale circulation in ABLs with a low amplitude of surface heat flux variation, namely 25 or 50 $Wm^{-2}$. However, when the amplitude is high, greater than or equal to 100 $Wm^{-2}$, the contribution of mesoscale circulation to the entrainment may be more significant than that of turbulence (Fig. 4.3). Thus, this perspective of the amplitude of surface heat flux variation likely explains two seemingly contradictory results of the effect of surface heterogeneity on entrainment, from Avissar and Schmidt.
(1998) and Patton et al. (2005). Avissar and Schmidt (1998), prescribing only high amplitudes of surface heat flux variation (greater than 100 $Wm^{-2}$) in their numerical experiments, assert that mesoscale circulations significantly contribute to entrainment in the convective ABL. However, Patton et al. (2005), indirectly prescribing a low amplitude of surface heat flux variation (about 40 $Wm^{-2}$), contend that turbulence dominates the entrainment process even over a heterogeneous surface.

Compared with the $r$ spectra, the $\theta$ spectra in Fig. 4.4 exhibit more sensitive response to the imposed surface heterogeneity at the mesoscale, namely 32 km, through all the ABL. This very sensitive response of $\theta$ to mesoscale surface forcing has been demonstrated both in the MR spectra from observed ABLs in Kang et al., (2006) and the Fourier spectra from simulated ABLs (Kang and Davis, 2006). For $\theta$, removing variances from mesoscale fluctuations and/or removing nonlinear transfer of variance from the meoscale fluctuations to higher wavenumbers likely causes underestimation of microscale (turbulence) variances. In other words, the very sensitive response of $\theta$ variance to the mesoscale surface heterogeneity may be associated with the decrease of microscale (turbulence) variances. To correct this sensitive response, Kang et al. (2006) used a longer spatial high-pass filter for $\theta$ than for $r$ to fit the observations to the vertical profiles predicted by mixed-layer similarity.

For case MB050, the effects of different surface conditions on the spectra are investigated. In Fig. 4.5, the $r$ spectra from the ABL simulated with $B=0.3$ and $R=1$ (MB050) are compared with those from the ABLs with $B=2.3$ and/or $R=2$ (MB050W2, MB050D2R2, and MB050W2R2). The ABLs over a dry surface ($B=0.3$; MB050, MB050D2R2) exhibit somewhat larger mesoscale variances than the ABLs over a moist
surface ($B=2.3$; MB050W, MB050W2R2). The ABLs representing a time short after rainfall ($R=2$; MB050D2R2, MB050W2R2) demonstrate larger mesoscale variances than the ABLs at a time long after rainfall ($R=1$; MB050, MB050W2). The $\theta$ spectra do not show any noticeable difference between the ABLs with different surface conditions given in this study (compare MB100 with MB100W2R2 in Fig. 4.4).

4.3.3 MR cospectral analysis

Fig. 4.6 shows the MR cospectra of $\theta-r$, $\theta-w$, and $r-w$ at 0.2, 0.4, 0.6 and 0.8 $\langle z_i \rangle$ (where $\langle z_i \rangle$ represents a domain-averaged value of the ABL depths) from MB000. The combinations of the positive cospectral densities of $\theta-r$, $\theta-w$, and $r-w$ obviously infer the heating and moistening from the surface (Mode I in Table 3 of Kang et al. (2006)), which is the dominant physical process in the lower part of the ABL, namely 0.2 and 0.4 $\langle z_i \rangle$. In the upper part of the ABL, namely 0.6 and 0.8 $\langle z_i \rangle$, entrainment from the free atmosphere is involved in the physical processes. While the entrainment of dry and warm air from the free atmosphere (Mode II in Table 3 of Kang et al. (2006)) seems to be the principal process at 0.8 $\langle z_i \rangle$, air coming from the surface (warm and moist air updraft) and from the entrainment (warm and dry air downdraft) are likely mixed at 0.6 $\langle z_i \rangle$ (Mode IV in Table 3 of Kang et al. (2006)). The different physical processes active between at the lower and upper parts of the ABL are associated with the transition of $\theta-r$ correlation coefficient ($\rho_{\theta_r}$) from positive to negative values.
at about 0.45 $z_i$ which is a well-known characteristic of the homogeneous ABL (e.g. Berg and Stull, 2004; Mahrt, 1991; Wyngaard et al., 1978).

Also in Fig. 4.6, in the middle part of the ABL, namely 0.4 and 0.6 $\langle z_i \rangle$, the cospectral density peak of moisture flux is located at a larger scale than that of heat flux. For case MB000, this extension of moisture flux to larger scale is likely associated with the spectral density peak of $r$ (Fig. 4.3) at a larger scale compared to those of $w$ (Fig. 4.2) and $\theta$ (Fig. 4.4). Jonker et al. (1999), using an LES for investigating the convective ABL in absence of large-scale forcings, concluded that a passive scalar can acquire larger-scale fluctuations, while temperature and vertical velocity fluctuate on a horizontal scale of the order of the ABL depth. Considering that $r$ can be assumed as a passive scalar for unsaturated air, one can expect the cospectral peak of moisture flux at a larger scale than that of heat flux even in the homogeneous ABL.

The cospectra from case MB050 in Fig. 4.7 demonstrate that the $\theta-r$ correlation is most negative in the lower part of the ABL, namely 0.2 and 0.4 $\langle z_i \rangle$. This negative correlation is mainly due to the large negative cospectral density of $\theta-r$ imposed at the surface heterogeneity scale, namely 32 km. Given the prescribed condition of warm, dry surface and cool, moist surface based on the surface energy balance, the large negative cospectral density of $\theta-r$ is not a surprising result. In other words, the negative $\theta-r$ correlation is basically caused by the imposed half-wave phase lag between the surface heat and moisture flux distributions, (4.1) and (4.2). However, both the large negative cospectral density of $\theta-r$ at about 10 km and the negative $\theta-r$ correlation at the low-level ABL are also observed. Kang et al. (2006) presented a large negative cospectral
density at about 10 km in the $\theta - r$ cospectra (Fig. 13 of Kang et al. (2006)) from the aircraft observations over the 60-km track on 25 May during IHOP_2002. In fact, this numerical experiment is set up based on the background weather (specifically, the sounding selected as the initial condition) and surface (specifically, surface heat and moisture flux) conditions from Kang et al. (2006). In addition Mahrt (1991), analyzing data collected by aircraft over a 120-km track at 150 m above ground level, concluded that, on the scale of larger than 10 km, the $\theta - r$ correlation becomes negative near surface, which is associated with constraints imposed by the surface energy balance.

Compared with the cospectra from case MB050 in Fig. 4.7, the signature of the mesoscale circulation, warm dry air updraft and cool moist air downdraft (Mode III in Table 3 of Kang et al. (2006)), becomes more obvious in the cospectra from case MB100 in Fig. 4.8. Although this noticeable exhibition of Mode III at the imposed surface heterogeneity scale (namely 32 km) looks like a result of enhanced mesoscale heat and moisture fluxes, it is in fact caused by the increases of mesoscale fluctuations of $\theta$ and $r$. In other words, due to the insignificant mesoscale vertical velocities (Fig. 4.2), Mode III is the consequence of the microscale (turbulent) $w$ fluctuations transporting enhanced mesoscale $\theta$ and $r$ fluctuations instead of the consequence of increased mesoscale fluxes (the product of mesoscale $w$ fluctuations and mesoscale fluctuations of $\theta$ or $r$). For example, at 0.8 $\langle z_i \rangle$ in case MB100, the spectral density peak at the imposed surface heterogeneity scale in the $\theta$ spectra (Fig. 4.4) is a bit greater than that in the $r$ spectra (Fig. 4.3). Also at 0.8 $\langle z_i \rangle$, the cospectral density peak at the heterogeneity scale in the cospectra of $\theta - w$ is somewhat greater than that in the cospectra of $r - w$. 

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Fig. 4.9 presents the cospectra from the ABL simulated with a moist surface, namely $B=0.3$ (MB100W2). Compared with the cospectra from the ABL with dry surface (MB100) in Fig. 4.8, the $r-w$ cospectra exhibit less significant vertical moisture flux at the imposed surface heterogeneity scale. This less significant vertical moisture flux is associated with less mesoscale variance in the $r$ spectra. Fig. 4.5 shows that the spectral peak at the mesoscale for case MB050W is less than the spectral mesoscale peak for case MB050. Similarly the greater mesoscale variance in the $r$ spectra from the ABL at a time shortly after rainfall ($R=2$) than at a time long after rainfall ($R=1$) results in greater vertical moisture flux at the mesoscale in the $r-w$ cospectra from the ABL simulated with $R=2$, which is not shown here. Thus, at the imposed mesoscale surface heterogeneity scale, the vertical moisture flux obtained from the ABL over a dry surface at a short time after rainfall responds more sensitively to the mesoscale surface forcing than that from the ABL over a moist surface at a long time after rainfall. However, as can be expected from the $\theta$ spectra (Fig. 4.4), the changed surface conditions do not affect the cospectra of $\theta-w$, as shown in Fig. 4.9.

The cospectra from case MB200W2, shown in Fig. 4.10, are appreciably different from those from cases MB050 (Fig. 4.7) and MB100 (Fig. 4.8). First, at the surface heterogeneity scale, 32 km, mixed signatures of air coming from free atmosphere and from the surface (Mode IV in Table 3) are exhibited instead of the signatures of mesoscale circulation (Mode III in Table 3 of Kang et al. (2006)). In fact, the sign change of the $r-w$ cospectral density from negative to positive causes the transform from Mode III to Mode IV. The positive $r-w$ cospectral density at the mesoscale through all the
ABL can be interpreted as deep entrainment \((w < 0 \text{ and } r < 0)\) or strong vertical transport of moisture from surface to the ABL top \((w > 0 \text{ and } r > 0)\). As shown in Fig. 4.2, mesoscale \(w\) variance is absent even in case MB200W2. Thus, the significant mesoscale peaks in the \(r\) spectra (Fig. 4.3) imply that the positive \(r - w\) cospectral density is likely caused by the combination of microscale \(w\) and mesoscale \(r\) fluctuations. Second, the large negative cospectral density of \(\theta - r\) is apparent at about 10 km, which is somewhat smaller than the imposed surface heterogeneity scale, 32 km. This result seems to be associated with the blunt mesoscale peaks of the \(r\) spectra for case MB200W2 in Fig. 4.3. The blunt mesoscale peak implies more considerable transfer of mesoscale variances to higher wavenumbers compared with the sharp mesoscale peaks for the other cases. Thus for case MB200W2, the relocated scale of the large negative cospectral density of \(\theta - r\) results from more active transfer of mesoscale \(r\) variance to higher wavenumbers. However, detailed investigation on the ABL with a high amplitude (higher than 100 \(Wm^{-2}\)) of surface flux variation is beyond the scope of this study and thus will be left for future work.

### 4.4 Decomposition

Although the surface heterogeneity is prescribed at the mesoscale, namely 32 km, the variances and fluxes in the ABLs are influenced across the overall scale range. Thus, to investigate the ABL from the two perspectives of mesoscale and microscale fluctuations, a scalar \(\phi\), which can be potential temperature \((\theta)\), water vapor mixing
ratio \( r \), or one of the three velocity components \((u,v,w)\), is decomposed into mesoscale and microscale fluctuations.

First the scalar \( \phi \) is separated into domain average \( \langle \phi \rangle \) and its deviation \( \phi' \).

\[
\phi(x_i; z, t) = \langle \phi \rangle(z, t) + \phi'(x_i; z, t)
\]  \hspace{1cm} (4.3)

where \( i = 1, 2 \). The deviation in (4.3) is referred to as total fluctuation. The microscale fluctuation is defined as the deviation from the mesoscale mean \[ \phi \], which is obtained by using a one-dimensional, low-pass wave-cutoff filter along the streamwise \( (x_i) \) direction. Here, the cutoff wavelength \( (C_f) \) of 2 km is selected. Two factors are considered for choosing \( C_f \): one is the typical size of energy-containing eddies, which is 1.5 \( z_i \) (where \( z_i \) is the ABL depth) and here about 1000-1500 m, and the other is the effective resolution of \( 6\Delta \) (where \( \Delta \) is a horizontal grid spacing), which is 600 m here. (For details on the wave-cutoff filter used here, refer to Chapter 3.)

\[
\phi''(x_i; z, t) = \phi(x_i; z, t) - \langle \phi \rangle(x_i; z, t)
\]  \hspace{1cm} (4.4)

Meanwhile, the difference of the mesoscale mean from the domain average is defined as mesoscale fluctuation.

\[
\phi^M(x_i; z, t) = \langle \phi \rangle(x_i; z, t) - \langle \phi \rangle(z, t)
\]  \hspace{1cm} (4.5)

For \( \theta \) at the height of \( z = 487 \text{ m} \) and the time of \( t = 240 \text{ min} \), the decomposition is illustrated in Fig. 4.11.
Fig. 4.12 and 4.13 exhibit $[w]$ and $[u_i]$, and $[r]$ and $[\theta]$, respectively for case MB050. These mesoscale means are temporally averaged over 60-240 min. Although the magnitudes are somewhat different, the organized pattern of the temporally averaged mesoscale mean fields is similar for other cases (not shown). For this investigation, the ABL over the mesoscale surface heterogeneity is divided into three regions: (1) warmer region ($-\lambda/2 \leq x_i < 0$) (2) cooler region ($0 \leq x_i < \lambda/2$), and (3) middle region ($-\lambda/4 \leq x_i < \lambda/4$). Here it should be noted that the temporally averaged $[w]$ field (Fig. 4.12) demonstrates that stronger upward (or downward) thermals persist in the warmer (or cooler) region. However, the absence of the spectral density peak at the mesoscale in the $w$ spectra (Fig. 4.2) indicates that the persistent thermal patterns are not associated with mesoscale vertical velocities. Rather, they are associated with microscale vertical motions.

4.5 Joint Frequency Distributions

To examine physical processes in the ABL, many scientists (e.g., Berg and Stull, 2004; Mahrt and Paumier, 1984; Holland, 1973) utilized the joint frequency distribution (JFD) technique. Particularly, Berg and Stull (2004) developed a parameterization of JFDs of potential temperature ($\theta$) versus water vapor mixing ratio ($r$) to predict the tilt of the JFD from positive to negative with height, based on the assumption that the only source regions of air mixture are the surface and the entrainment zone. We apply the JFD to investigate ABL structure over the mesoscale surface heterogeneity from two
perspectives: mesoscale and microscale (turbulent) fluctuations. Especially, using microscale fluctuations of $\theta$ and $r$, the applicability of the parameterization of $\theta - r$ JFD is tested in the ABL over the mesoscale surface heterogeneity.

The total fluctuations ($\phi'$) and microscale fluctuations ($\phi''$) of potential temperature ($\theta$), water vapor mixing ratio ($r$), and vertical velocity ($w$) are obtained at every 100 seconds between 60 and 240 min. At each time, the $\phi'$ and $\phi''$ are normalized by an appropriate scaling parameter. For $w$, the convective velocity scale (Deardorff, 1970a, b) computed as

$$w_s = \left( \frac{g}{\theta_{ML}} \langle w' \theta'_{sfc} \rangle \langle z_i \rangle \right)^{1/3}$$

(4.6)

is used for the normalization. In (4.6), $g$ is the gravitational acceleration, $\theta_{ML}$ the potential temperature of the mixed-layer (0.2-0.8 $z_i$), $\langle w' \theta'_{sfc} \rangle$ the domain-averaged heat flux at the surface, and $\langle z_i \rangle$ the domain-averaged ABL depth. For $\theta$, we use the mixed-layer temperature scale (Deardorff, 1970b) computed as

$$\theta_s = \frac{\langle w' \theta'_{sfc} \rangle}{w_s}$$

(4.7)

For $r$, we use the mixed-layer humidity scale (Stull, 1988) computed as

$$r_s = \frac{\langle w' r'_{sfc} \rangle}{w_s}$$

(4.8)

In (4.8), $\langle w' r'_{sfc} \rangle$ is the domain-averaged moisture flux at the surface at each time.
In Fig. 4.14, the JFDs of total fluctuations of potential temperature ($\theta'$) and water vapor mixing ratio ($r'$) from case MB050 are compared with those from case MB000. Through the mixed layer (0.2-0.8 $z_i$), the JFDs from case MB050 extend into the quadrants of warm dry and cool moist. In the ABL over a homogenous surface (MB000), the slope of the JFD of $\theta$ and $r$ is changed from positive at the lower ABL to negative at the upper ABL. However, in the ABL over a heterogeneous surface (MB050), the slope of the JFD is negative through the mixed layer. The negative correlations between $\theta$ and $r$ through the mixed layer over a heterogeneous surface indicate that the prescribed surface fluxes based on the surface energy balance influence the whole ABL structure, as already discussed in section 4.3.

Fig. 4.15 compares the JFDs of microscale fluctuations of potential temperature ($\theta''$) and water vapor mixing ratio ($r''$) from cases MB000 and MB050. The JFDs from MB050 are similar to those from MB000. These results suggest that the parameterization (Berg and Stull, 2004) of $\theta$--$r$ JFDs, built based on the homogeneous ABL condition, could be applied even in the ABL over the mesoscale surface heterogeneity, as long as an appropriate high-pass wave cutoff filter is used to remove mesoscale fluctuations.

In the JFDs of $r'$ and $w'$ shown in Fig. 4.16, the frequency distribution of $w'$ from MB050 has a slightly narrower range than that from MB000, whereas the frequency distribution of $r'$ from MB050 is wider than that from MB000. In the JFDs of $\theta'$ and $r'$ (Fig. 4.14), the frequency distributions of both $\theta'$ and $r'$ from MB050 have a wider range compared with those from MB000. Also in the JFDs of $\theta'$ and $w'$ (not shown here), the range of $w'$ from MB050 is slightly narrower than that from MB000, whereas
the range of $\theta'$ from MB050 is wider than that from MB000. In fact, comparing the JFDs of $\dot{r}^*$ and $\dot{w}^*$ (Fig. 4.17) with those of $\dot{r}'$ and $\dot{w}'$ from MB050, the frequency distribution range of $\dot{w}^*$ remains almost unchanged, whereas the range of $\dot{r}^*$ becomes much narrower. These results support the absence of mesoscale vertical velocity in case MB050. Thus, in the ABL over a heterogeneous surface, the magnitude of the domain-averaged total variance of vertical velocity is that of the domain-averaged microscale (turbulence) variance of vertical velocity, namely $\langle \dot{w}'^2 \rangle \approx \langle \dot{w}^*^2 \rangle$. The magnitude of the domain-averaged vertical velocity variance is a function of static stability which is controlled by the surface-heterogeneity-induced mesoscale horizontal flow (Kang and Davis, 2006).

Fig. 4.18 shows the JFDs from MB000W2 and MB050W2. The range of the frequency distribution of $\dot{r}'$ in the JFDs from MB050W2 is similar to that in the JFDs from MB000W2. In contrast, in Fig. 4.14, the range of the frequency distribution of $\dot{r}'$ in the JFDs from MB050 is about two times that in the JFDs from MB000. This result is consistent with the fact that in the $\dot{r}$ spectra the spectral density peak at the imposed surface heterogeneity scale becomes much less significant from MB050 to MB050W2 (Fig. 4.5). Also consistent with that in the $\dot{r}$ spectra, the JFDs from MB050W2 and MB050W2R2 (not shown here) demonstrate larger mesoscale variances for the ABL at a short time after rainfall (MB050W2R2) than for the ABL at a long time after rainfall (MB050W2).

In Fig. 4.18, it should be noted that the range of $\dot{r}'/\dot{r}$ in MB000W2, where $\dot{r}$ is the mixed-layer humidity scale parameter in (4.8), demonstrates a significant difference
from that in MB000 (Fig. 4.14). Thus, in Fig. 4.19, we plot the vertical profiles of domain-averaged total variances of $\theta$ and $r$ from the homogeneous ABLs with different Bowen ratios (MB000D3, MB000, MB000D1, MB000W1, and MB000W2). Here the $\theta$ variances are normalized by $\theta$, and the $r$ variances are normalized by $r$. Fig. 4.19 demonstrates that the mixed-layer humidity scale parameter $r$ fails to collapse the variances from the homogeneous ABLs with various Bowen ratios to a universal curve, whereas the mixed-layer temperature scale parameter $\theta$ meets the universal curve except at the upper-level ABL of MB000W2, which is likely associated with the cloud development for this moistest case. Some scientists (Bern and Stull, 2004; Wyngaard, 1998) have suggested that the $r$ may not be an appropriate scaling parameter due to the negligence of the entrainment from the free atmosphere. However, here, we suggest that the $r$ should be revised to be universally used for the data from the ABLs with a different Bowen ratio.

In Fig. 4.20, the JFDs of $\theta''$ and $r''$ from cases MB200W2 are compared with those from MB000W2. Unlike for case MB050 in Fig. 4.15 and 4.17, the JFDs from the ABL over a heterogeneous surface (MB200W2) do not match those from the ABL over a homogeneous surface (MB000W2). For the ABL with high amplitude of surface heat flux variation, the microscale (turbulence) fluctuations, which are obtained after applying the high-pass wave-cutoff filter, also do not fit into the mixed-layer similarity theory. This deviation of microscale (turbulence) fluctuations in case MB200W2 from the mixed-layer similarity can be expected from the cospectra in Fig. 4.10. In fact, in case
MB200W2, the ABL is not quasi-stationary due to temporally oscillating mesoscale horizontal flows, which are explained in detail in Kang and Davis (2006).

4.6 Horizontal transport of heat and moisture

This section examines the significance of the horizontal transport of heat and moisture by the surface-heterogeneity-induced mesoscale flows. In addition, the effect of the horizontal transport of moisture is discussed from a perspective of the creation of a favorable environment for moist convection.

4.6.1 Surface-heterogeneity-induced horizontal flows

Fig. 4.21 shows the time evolution of $V$, the magnitude of $\left[ u_1 \right]$ averaged over the middle region ($-\lambda/4 \leq x \leq \lambda/4$), at 0.2 $z$. Fig. 4.21 also exhibits the time evolution of the gradient of $\theta$ (or $r$) averaged between the warmer region and the cooler region at 0.2 $z$, which will be denoted as $\delta[\theta]$ and $\delta[r]$, respectively. The $V$ steadily increases with time, which implies that, in the range of the amplitude of surface heat flux variation used here (100 Wm$^{-2}$ or lower), the temporally oscillating mesoscale horizontal flows (Kang and Davis, 2006; Letzel and Raasch, 2003) are not generated. However, for case MB100, the $\delta[\theta]$ starts to decrease from 190 min, whereas for other cases the $\delta[\theta]$ steadily increases with time. The reduction of the $\delta[\theta]$ likely results from horizontal transport of temperature by the generated mesoscale horizontal flows. (The detailed
processes are explained in Kang and Davis (2006)). The $\delta[r]$ starts to attenuate from 180 min and 210 min. for MB100 and MB075 respectively. This earlier start of the $\delta[r]$ reduction implies that, comparing $[r]$ with $[\theta]$, the effect of the horizontal transport of $[r]$ by the generated mesoscale flows is likely more significant.

4.6.2 Horizontal transport of heat and moisture

The amounts of heat and moisture transported by the generated mesoscale horizontal flows will be computed in this section. For this estimation, the re-filtered equation, neglecting the molecular diffusion term, for a scalar $\phi$ (which is $\theta$ or $r$) with the assumption of $[u_z] \approx 0$ and $[u_z] \approx 0$ is utilized. By assuming negligible horizontal turbulent fluxes and disregarding the subfilter-scale flux, the time change in mean fluid scalar $[\phi]$ is expressed as

$$\frac{\partial[\phi]}{\partial t} = -\frac{\partial}{\partial z} [w^* \phi^*] - [u_i] \frac{\partial[\phi]}{\partial x_i}$$

(4.9)

Integrating (4.9) from the surface ($z_{Sfc}$) to the model top ($z_{Top}$) yields

$$\frac{1}{z_{Top}} \int_{z_{Sfc}}^{z_{Top}} [\phi] dz = -\int_{z_{Sfc}}^{z_{Top}} \frac{\partial}{\partial z} [w^* \phi^*] dz - \int_{z_{Sfc}}^{z_{Top}} [u_i] \frac{\partial[\phi]}{\partial x_i} dz$$

(4.10)

Integrating (4.10) from the initial time to a specified time $t$ and assuming $[w^* \phi^*]_{z_{Top}} = 0$ yields

$$\int_{0}^{t} \int_{z_{Sfc}}^{z_{Top}} [u_i] \frac{\partial[\phi]}{\partial x_i} dz \ dt = \left( [w^* \phi^*]_{Sfc} \right) t - \int_{z_{Sfc}}^{z_{Top}} ([\phi]_t - [\phi]_{t=0}) dz.$$  

(4.11)
We shall refer to the term in the left hand side as the advection term, and refer to the first term and the second term in the right hand side as the surface flux term and the temporal term, respectively.

For case MB050, the vertical profiles of $[\theta]$ and $[r]$ averaged over the warmer region and over the cooler region at 0 min. and 240 min. are presented in Fig. 4.22. Also in Fig. 4.22, the vertical profiles of the difference of $[\theta]$ (or $[r]$) between 240 min. and 0 min. are exhibited. Integrating the area of the difference of $[\theta]$ (or $[r]$) averaged over the warmer region (or over the cooler region) from the surface ($z_{sl}$) to the model top ($z_{top}$) yields the amount of $[\theta]$ (or $[r]$) absorbed ($ABS$, the second term in the right hand side of (4.11)) by the atmosphere in the warmer region (or the cooler region). Here the model top ($z_{top}$) can be replaced by the uppermost height ($h_z$) reached by upward penetrating mixed-layer turbulent thermals. As in (4.11), the amount of $[\theta]$ (or $[r]$) transported from the other region ($ADV$, the term in the left hand side of (4.11)) is computed by subtracting $ABS$ from $SUP$ (the first term in the right hand side of (4.11)), which is obtained by the temporal integration of the surface flux prescribed in (4.11) averaged over each region, namely over the warmer region or over the cooler region.

Fig. 4.23 shows the ratios of $ADV/SUP$ for various cases with different amplitudes of surface flux variation. In the warmer region $ADV/SUP > 0$ for potential temperature ($\theta$) and $ADV/SUP < 0$ for water vapor mixing ratio ($r$). In other words, in the warmer region the amount of heat supplied from the surface is more than that absorbed by the atmosphere whereas the amount of moisture supplied from the surface is
less than that absorbed by the atmosphere. For potential temperature, the atmosphere in
the cooler region absorbs 130% of the amount of heat supplied from the surface in the
cooler region for case MB100. That is to say, 30% of the amount of heat absorbed by the
atmosphere in the cooler region is transported from the warmer region. For potential
temperature, a 30% gain in the cooler region from the warmer region is the most
significant case in Fig. 4.23. Compared with heat, the horizontal transport of moisture by
the generated mesoscale flows is much more substantial. In Fig. 4.23, for case MB100 the
amount of moisture transported from the cooler region to the warmer region reaches up to
360% of the amount supplied from the surface in the warmer region. Even for case
MB050, the quantity of moisture transported from the cooler region to the warmer region
becomes up to 90% of the amount supplied from the surface in the warmer region by 240
min, as shown in Fig. 4.23.

We also investigate the influence of the Bowen ratio, which is here controlled by
the surface moisture flux, on the horizontal transport of moisture. As shown in Fig. 4.24,
as the Bowen ratio increases (from MB050W2 to MB050D3), the horizontal transport of
moisture becomes more significant. That is to say, in the ABL over a dry surface, much
more moisture relatively is transported horizontally from the other region than is supplied
from the surface. Nevertheless, the real amount of moisture transported can be more in
the ABL over a moist surface than over a dry surface due to the amount of available
moisture. Thus in Fig. 4.24, for a dry surface when there is more available moisture in the
cooler region (MB100D1R2), the amount of moisture transported is 220% of the amount
of moisture supplied from the surface in the warmer region at 240 min, which is much
larger than 105% for case MB100D1. However, for the moist surface, the ABL at a time
short after rainfall does not show any difference from the ABL at a time long after rainfall (MB100W2 .VS. MB100W2R2).

4.6.3 Relative humidity (RH) distribution near the ABL top

The RH distributions in Fig. 4.25 at 120 min. and 240 min. for case MB100W2 exhibit high values near the ABL top. Ek and Mahrt (1994), using a one-dimensional model to analyze observations from a fair-weather day in a field experiment, concluded that when the capping inversion is strong and the above-ABL air is somewhat dry, the RHs near the ABL top over a moister surface is generally greater due to slower boundary layer growth and less dry air entrainment. The initial conditions of this study as shown in Fig. 4.1 satisfy the requirements to meet this result. As Ek and Mahrt (1994) suggested, the RHs near the ABL top in the cooler region (the moist region) are greater than the values in the warmer region (the dry region) at 120 min. However, at 240 min., the RHs near the ABL top in the warmer region are greater than those in the cooler region. This transition of the greater near-the-ABL-top RH region is likely associated with the mesoscale horizontal transport of moisture from the cooler region to the warmer region. For case MB100W2, in the warmer region at 240 min., the amount of moisture transported from the cooler region is 40 % of the amount of moisture supplied from the surface, as shown in Fig. 4.24. Considering that the latent heat (LE) averaged over the warmer region is 702 \text{ Wm}^{-2}, the latent heat (LE) transported from the cooler region is 281 \text{ Wm}^{-2} in the warmer region at 240 min for case MB100W2. Although for case MB100D1R2 the amount of moisture transported is 220 % of the amount of moisture
supplied in the warmer region at 240 min (Fig. 4.24), the LE transported from the cooler region is 240 Wm$^{-2}$, considering that the LE averaged over the warmer region is only 72 Wm$^{-2}$. Thus, the real amount of moisture transported from the cooler region to the warmer region is somewhat more for case MB100W2 than for the highest $ADV/SUP$ ratio case (MB100D1R2).

Figs. 4.26 and 4.27 show the difference of the RH averaged over the warmer region and over the cooler region at 0.9 [$z_i$] for various cases. In Fig. 4.26, the near-the-ABL-top RH in the warmer region becomes greater than that in the cooler region at 165 min. for case MB075 and 135 min. for case MB100. For cases MB025 and MB050, the cooler region has a greater near-the-ABL-top RH as is expected by Ek and Mahrt (1994). Without considering the horizontal transport of moisture, Ek and Mahrt (1994) examined the dependence of boundary layer RH on surface wetness, capping inversion intensity, above-the-ABL dryness, and mesoscale vertical velocity.

In Fig. 4.27, the near-the-ABL-top RH differences from cases for the dry surface condition ($B=1.2$; MB050D1, MB100D1) are compared with those from cases for the moist surface condition ($B=0.3$; MB050W2, MB100W2). In the ABL over a moist surface, the influence of the moisture transport on the RH difference seems to be more significant. Also in Fig. 4.27 the near-the-ABL-top RH differences from cases for the ABL at a time long after rainfall ($R=1$) are compared with those from cases for the ABL at a time short after rainfall ($R=2$). In the ABL shortly after rainfall, the influence of the moisture transport on the RH difference is likely more significant. To summarize, when
there is more available moisture in the upwind region (here in the cooler region), the effect of the horizontal transport becomes more significant.

4.7 Chapter Conclusions and future plans

Fair-weather moist ABLs (about 1130-1430 LST) are simulated with a surface flux variation that is sinusoidal with mean, amplitude, and wavelength. The wavelength ($\lambda$) of the surface flux variation is fixed at 32 km, but the amplitude of the variation ($A_{w\theta}$ or $A_w$) varies from 0 to 200 Wm$^{-2}$. (However, this study is primarily focused on ABLs with amplitude variation in the range of 0-100 Wm$^{-2}$.) From all simulations, the multiresolution (MR) spectra of vertical velocity ($w$) demonstrate that the spectral density peak is absent at the imposed surface heterogeneity scale, namely 32 km (Fig. 4.2). In contrast to the absence of mesoscale $w$ variance, both potential temperature ($\theta$) and water vapor mixing ratio ($r$) obviously respond to the mesoscale surface heterogeneity (Figs. 4.3, 4.4, and 4.5). Compared with water vapor mixing ratio ($r$), potential temperature ($\theta$) responds more sensitively to the mesoscale surface forcing (Fig. 4.3 versus Fig. 4.4). In the spectra of $\theta$ and $r$, the spectral density peaks at the imposed heterogeneity scale are associated with the cospectral density peaks at the mesoscale in the cospectra of $\theta - w$ and $r - w$ (Figs. 4.7, 4.8, and 4.9). Here it should be noted that the cospectral peaks are caused by the correlations between mesoscale fluctuations of $\theta$ (or $r$) and microscale fluctuations of $w$, not mesoscale fluctuations of $w$. Thus, compared with the response of $r$, the more sensitive response of $\theta$ to the
mesoscale surface forcing results in the somewhat more significant heat flux compared with moisture flux at the mesoscale in Figs. 4.7, 4.8, and 4.9. Even in the ABL over the mesoscale surface heterogeneity, the microscale (turbulent) fluxes of heat and moisture are still substantial in particular at the lower part of the ABL due to the absence of mesoscale fluctuations of \( w \) (Figs. 4.7, 4.8, and 4.9).

The cospectra of \( \theta - \tau \), however, exhibit much more significant cospectral density peaks at the mesoscale than at the microscale. In fact, the spectral density at the microscale in the cospectra of \( \theta - \tau \) can be ignored, as shown in Figs. 4.7, 4.8, and 4.9. Comparing the cospectral density peaks at the cospectra of \( \theta - \tau \) with those at the cospectra of \( \theta - w \) and \( r - w \), one can conclude that mesoscale fluctuations of \( w \) are nominal. Thus, more mesoscale \( r \) variance in the ABL with drier surface or shortly after rainfall results in more vertical moisture flux at the imposed surface heterogeneity scale.

The joint frequency distributions (JFD) demonstrate that the correlation between \( \theta' \) (where the prime represents total fluctuation) and \( r' \) is highly negative through the ABL over a heterogeneous surface. However, the JFDs of \( \theta'' \) (where the double prime represent microscale fluctuation) and \( r'' \) from the ABL over a heterogeneous surface present similar patterns to those from the homogeneous surface. This result confirms that the negative correlations between \( \theta \) and \( r \) at the lower-level ABL over a heterogeneous surface are caused by the negative correlations between mesoscale fluctuations of \( \theta \) and \( r \). Thus, this result implies that if an appropriate high-pass wave cutoff filter is applied, a JFD parameterization of \( \theta \) and \( r \) (e.g., Berg and Stull, 2004), which is built based on the assumption of the homogenous ABL, still can be used. However, this result is valid only to the ABLs simulated with low amplitude (100 Wm\(^{-2}\) or lower) of surface heat flux.
variation. For cases where the ABL is simulated with high amplitude (greater than 100 Wm$^{-2}$) of surface flux variation (MB150W2 and MB200W2), the JFDs of $\theta''$ and $r''$ do not match those from the homogeneous ABL because the ABL does not satisfy the quasi-steady state condition (For details, see Kang and Davis (2006)).

For the ABL with low amplitude (100 Wm$^{-2}$ or lower) of surface heat flux variation, the amount of heat (or moisture) transported (ADV) by the surface-heterogeneity-induced mesoscale horizontal flows is compared with the amount of heat (or moisture) supplied (SUP) from the surface at each region, namely at the warmer region and at the cooler region. The amount of heat transported (ADV) from the warmer region to the cooler region is less than 30% of the amount of heat supplied (SUP) from the surface in the cooler region. However, the amount of moisture transported (ADV) from the cooler region to the warmer region is as much as 130% of the amount of moisture supplied (SUP) from the surface in the warmer region. From the perspective of the ratio of $ADV/SUP$, the horizontal transport of moisture becomes much more significant in the ABL over a dry surface shortly after rainfall (Fig. 4.24).

When the amplitude of surface flux variation is rather low, specifically for cases MB025 and MB050, the surface-heterogeneity-induced mesoscale horizontal flow is relatively weak and the mesoscale gradient of heat (or moisture) is also less significant (Fig. 4.21). As the advection term in (4.11) suggests, the horizontal transport of heat (or moisture) seems to grow rapidly with the increasing amplitude of surface flux variation due to the increases of both the intensity of mesoscale horizontal flow and the mesoscale gradient of heat (or moisture). In Figs. 4.23 and 4.24, with the higher amplitude of surface flux variation, specifically for cases MB075 and MB100, the exponential
temporal growth of the amount of moisture transported from the cooler region to the warmer region (ADV) is demonstrated. Thus, it is not a surprising result that the difference of the near-the-ABL-top RH between the warmer region and the cooler region grows so fast for cases MB075 and MB100 in Figs. 4.26 and 4.27. Based on this result, in the argument about the environment favorable for moist convection associated the surface heterogeneity, the effect of the moisture transport by the surface-heterogeneity-induced mesoscale flows should be included.

In fact, the dependence of moist convection on the distribution of heat and moisture in the ABL is quite complex. From the perspective of a whole domain, negative correlation of potential temperature ($\theta$) and water vapor mixing ratio ($r$) increases the variability of RH or LCL whereas positive correlation of $\theta$ and $r$ decreases the variability of RH or LCL (Mahrt, 1991). Thus, compared with the homogeneous ABL, clouds could form at a lower ratio of $z/z_i$ (where $z$ is height and $z_i$ is the ABL height) in the ABL over the mesoscale surface heterogeneity due to the increased negative correlation of mesoscale fluctuations of $\theta$ and $r$. Between the warmer region and the cooler region, the warmer region can develop strong moist convection if the surface-heterogeneity-induced mesoscale horizontal flow is strong enough and there is enough moisture available in the upwind region. However, depending on the intensity of the capping inversion and above-the-ABL dryness, the result could be different.

In future research, we plan to simulate the ABL under different atmospheric conditions (e.g., the capping inversion intensity, the above-the-ABL dryness, and the relatively moist ABL). Depending on the atmospheric conditions, the significance of the advection of moisture in initiating and developing cloud and precipitation would be
different. Thus, simulations with the different atmospheric conditions will allow us to study in detail the link of the ABL structure over the mesoscale surface heterogeneity to cloud development and precipitation.
Fig. 4.1: Initial profiles of potential temperature and water vapor mixing ratio
Fig. 4.2: Composite of the normalized MR spectra of vertical velocity \((w)\) at 0.2, 0.4, 0.6, and 0.8 \(\langle z_i \rangle\) for cases MB000, MB050, MB100, MB150W2, MB200W2. These spectra were computed using output every 100 s between 60-240 min. The \(\langle z_i \rangle\) represent a domain-averaged value of the ABL depth at each time. The error bar indicates temporal standard deviation over 60-240 min.
Fig. 4.3: Composite of the normalized MR spectra of water vapor mixing ratio ($r$) at 0.2, 0.4, 0.6, and 0.8 $\langle z_i \rangle$ for cases MB000, MB025, MB050, MB100, MB100W2, MB200W2. These spectra were computed using output every 100 s between 60-240 min. The $\langle z_i \rangle$ represent a domain-averaged value of the ABL depth at each time. The error bar indicates temporal standard deviation over 60-240 min.
Fig. 4.4: Composite of the normalized MR spectra of potential temperature ($\theta$) at 0.2, 0.4, 0.6, and 0.8 $\langle z_i \rangle$ for cases MB000, MB025, MB050, MB100, MB100W2R2, and MB200W2. These spectra were computed using output every 100 s between 60-240 min. The $\langle z_i \rangle$ represent a domain-averaged value of the ABL depth at each time. The error bar indicates temporal standard deviation over 60-240 min.
Fig. 4.5: Composite of the normalized MR spectra of water vapor mixing ratio ($r$) at 0.2, 0.4, 0.6, and 0.8 $\langle z_i \rangle$ for cases MB050, MB050W2, MB050D2R2, and MB050W2R2. These spectra were computed using output every 100 s between 60-240 min. The $\langle z_i \rangle$ represent a domain-averaged value of the ABL depth at each time. The error bar indicates temporal standard deviation over 60-240 min.
Fig. 4.6: Composites of the normalized MR cospectra of $\theta_r$ (solid), $\theta_w$ (dotted), and $rw$ (dashed) at 0.2, 0.4, 0.6, and 0.8 $\langle z_i \rangle$ for case MB000. These cospectra were computed using output every 100 s between 60-240 min. The $\langle z_i \rangle$ represent a domain-averaged value of the ABL depth at each time. The error bar indicates temporal standard deviation over 60-240 min.
Fig. 4.7: Composites of the normalized MR cospectra of $\theta r$ (solid), $\theta w$ (dotted), and $rw$ (dashed) at 0.2, 0.4, 0.6, and 0.8 $\langle z_i \rangle$ for case MB050. These cospectra were computed using output every 100 s between 60-240 min. The $\langle z_i \rangle$ represent a domain-averaged value of the ABL depth at each time. The error bar indicates temporal standard deviation over 60-240 min.
Fig. 4.8: Composites of the normalized MR cospectra of \( \theta r \) (solid), \( \theta w \) (dotted), and \( rw \) (dashed) at 0.2, 0.4, 0.6, and 0.8 \( \langle z_i \rangle \) for case MB100. These cospectra were computed using output every 100 s between 60-240 min. The \( \langle z_i \rangle \) represent a domain-averaged value of the ABL depth at each time. The error bar indicates temporal standard deviation over 60-240 min.
Fig. 4.9: Composites of the normalized MR cospectra of $\theta_r$ (solid), $\theta_w$ (dotted), and $rw$ (dashed) at 0.2, 0.4, 0.6, and 0.8 $\langle z_i \rangle$ for case MB100W2. These cospectra were computed using output every 100 s between 60-240 min. The $\langle z_i \rangle$ represent a domain-averaged value of the ABL depth at each time. The error bar indicates temporal standard deviation over 60-240 min.
Fig. 4.10: Composites of the normalized MR cospectra of $\theta_r$ (solid), $\theta_w$ (dotted), and $rw$ (dashed) at 0.2, 0.4, 0.6, and 0.8 $\langle z_i \rangle$ for case MB200W2. These cospectra were computed using output every 100 s between 60-240 min. The $\langle z_i \rangle$ represent a domain-averaged value of the ABL depth at each time. The error bar indicates temporal standard deviation over 60-240 min.
Fig. 4.11: Decomposition of potential temperature ($\theta$) at the height of 487 m at 240 min. for case MB050. Here, $[\theta]$ is mesoscale mean, $\langle \theta \rangle$ is domain average, $\theta'$ is total fluctuation, $\theta^M$ is mesoscale fluctuation, and $\theta^\ast$ microscale fluctuation.
Fig. 4.12: Two-dimensional cross sections ($x_1 - z$ plane) of mesoscale means of (Top) vertical velocity, and (Bottom) streamwise horizontal velocity averaged over 60-240 min. for case MB050. Here the height is normalized by the ABL depth, which is computed at each horizontal point at each time but later averaged over the $x_2$ direction.
Fig. 4.13: Same as in Fig. 4.12 but for (Top) water vapor mixing ratio and (Bottom) potential temperature. In the top figure, the white shaded area indicates the area which has moisture less than 2.4 \( gkg^{-1} \).
Fig. 4.14: Joint frequency distributions of the total fluctuations of potential temperature ($\theta'$) and water vapor mixing ratio ($r'$) at 0.2, 0.4, 0.6, and 0.8 $z_i$ for cases (shaded) MB000 and (contoured) MB050. Here, $z_i$ is the domain-averaged value of the ABL depth $\langle z_i \rangle$, which is computed at each output (100 sec) between 60-240 min. Also, for every output, the values of $\theta_e$ and $r_e$ are computed. The contour and color indicate the range of the frequency density within a grid of $\Delta \theta'/\theta_e$ and $\Delta r'/r_e$. The contour interval is 0.01.
Fig. 4.15: Joint frequency distributions of the microscale fluctuations of potential temperature ($\theta''$) and water vapor mixing ratio ($r''$) at 0.2, 0.4, 0.6, and 0.8 $z_i$ for cases (shaded) MB000 and (contoured) MB050. Here, $z_i$ is the domain-averaged value of the ABL depth $\langle z_i \rangle$, which is computed at each output (100 sec) between 60-240 min. Also, for every output, the values of $\theta_e$ and $r_e$ are computed. The contour and color indicate the range of the frequency density within a grid of $\Delta \theta''/\theta_e$ and $\Delta r''/r_e$. The contour interval is 0.01.
Fig. 4.16: Joint frequency distributions of the total fluctuations of water vapor mixing ratio \( r' \) and vertical velocity \( w' \) at 0.2, 0.4, 0.6, and 0.8 \( z_i \) for cases (shaded) MB000 and (contoured) MB050. Here, \( z_i \) is the domain-averaged value of the ABL depth \( \langle z_i \rangle \), which is computed at each output (100 sec) between 60-240 min. Also, for every output, the values of \( r_i \) and \( w_i \) are computed. The contour and color indicate the range of the frequency density within a grid of \( \Delta r'/r_i \) and \( \Delta w'/w_i \). The contour interval is 0.01.
Fig. 4.17: Joint frequency distributions of the microscale fluctuations of mixing ratio ($r''$) and vertical velocity ($w''$) at 0.2, 0.4, 0.6, and 0.8 $z_i$ for cases (shaded) MB000 (contoured) MB050. Here, $z_i$ is the domain-averaged value of the ABL depth $\langle z_i \rangle$, which is computed at each output (100 sec) between 60-240 min. Also, for every output, the values of $r_*$ and $w_*$ are computed. The contour and color indicate the range of the frequency density within a grid of $\Delta r''/r_*$ and $\Delta w''/w_*$. The contour interval is 0.01.
Fig. 4.18: Joint frequency distributions of the total fluctuations of potential temperature ($\theta'$) and water vapor mixing ratio ($r'$) at 0.2, 0.4, 0.6, and 0.8 $z_i$ for cases (shaded) MB000W2 and (contoured) MB050W2. Here, $z_i$ is the domain-averaged value of the ABL depth $\langle z_i \rangle$, which is computed at each output (100 sec) between 60-240 min. Also, for every output, the values of $\theta_i$ and $r_i$ are computed. The contour and color indicate the range of the frequency density within a grid of $\Delta \theta'/\theta_i$ and $\Delta r'/r_i$. The contour interval is 0.01.
Fig. 4.19: Vertical profiles of (left) potential temperature and (right) water vapor mixing ratio for cases MB000, MB000D1, MB000W1, MB000W2, and MB000D3
Fig. 4.20: Joint frequency distributions of the microscale fluctuations of potential temperature ($\theta''$) and vertical velocity ($w''$) at 0.2, 0.4, 0.6, and 0.8 $z_i$ for cases (shaded) MB000W2 and (contoured) MB200W2. Here, $z_i$ is the domain-averaged value of the ABL depth $\langle z_i \rangle$, which is computed at each output (100 sec) between 60-240 min. Also, for every output, the values of $r_*$ and $w_*$ are computed. The contour and color indicate the range of the frequency density within a grid of $\Delta r''/r_*$ and $\Delta w''/w_*$. The contour interval is 0.01.
Fig. 4.21: Time evolutions of (Top) the magnitude of $u_1$ averaged over the middle region, (Middle) the gradient of $\theta$, and (Bottom) $r$ averaged between over the warmer region and over the cooler region at 0.2 $z_i$ for cases MB000, MB025, MB050, MB075, and MB100. The magnitude of the $u_1$, the gradients of $\theta$ and $r$ are denoted as $V$, $\delta[\theta]$, and $\delta[r]$ respectively.
Fig. 4.22: (Left) Vertical profiles of $\theta$ and $r$ averaged over the warmer region (W) and over the cooler region (C) at 0 min. and 240 min. for case MB050. (Right) Vertical profiles of the difference of $\theta$ (or $r$) averaged over the warmer region (or over the cooler region) between at 240 min. and at 0 min.
Fig. 4.23: Time evolution of the ratio of the amount of (Top) potential temperature or (Bottom) water vapor mixing ratio transported from the other region, to the amount of the scalar supplied locally from the surface over each region, namely in the warmer region (solid line) and in the cooler region (dotted line) for cases MB000, MB025, MB050, MB075, and MB100.
Fig. 4.24: Same as in Fig.4.23 but for water vapor mixing ratio (Top) for cases MB050D3, MB050, MB050D1, MB050W1, and MB050W2, and (Bottom) for cases MB100D1, MB100D1R2, MB100W2, and MB100W2R2.
Fig. 4.25: Two-dimensional cross sections ($x_1$-$z$ plane) of relative humidity (Top) at 120 min. and (Bottom) at 240 min for case MB100W2. The solid line is the height normalized by the ABL depth. Here the relative humidity and the ABL depth are averaged over the $x_2$ direction.
Fig. 4.26: Time evolution of the difference of relative humidity averaged between over the warmer region and over the cooler region at 0.9 $z_i$ for cases MB000 (solid line), MB025 (dotted line), MB050 (dashed line), MB075 (one-dot-dashed line), and MB100 (three-dot-dashed line).
Fig. 4.27: Same as in Fig. 4.26 but for cases MB050D1, MB050D1R2, MB050W2, MB050W2R2, MB100D1, MB100D1R2, MB100W2, and MB100W2R2.
Table 4.1: Experimental design parameters for the prescribed surface heat and moisture flux variation. The units of $\langle F_{w\theta} \rangle$ and $A_{w\theta}$ are $Kms^{-1}$ and the units of $\langle F_{wr} \rangle$ and $A_{wr}$ are $gkg^{-1}ms^{-1}$. In the columns of cloud water and rain water, the number is the maximum value over 180-240 min. and the number in the parenthesis is the total value summed over 180-240 min.

<table>
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<th>Case</th>
<th>$\langle F_{w\theta} \rangle$</th>
<th>$\langle F_{wr} \rangle$</th>
<th>$A_{w\theta}$</th>
<th>$A_{wr}$</th>
<th>$B$</th>
<th>$R$</th>
<th>Cloud water (gKg$^{-1}$)</th>
<th>Rain water (gKg$^{-1}$)</th>
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<td>0.04</td>
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<td>2</td>
<td>0.4 (0.4)</td>
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<td>0.04</td>
<td>0.033</td>
<td>0.3</td>
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<td>0.0</td>
</tr>
<tr>
<td>MB100W1R2</td>
<td>0.19</td>
<td>0.095</td>
<td>0.08</td>
<td>0.066</td>
<td>0.8</td>
<td>2</td>
<td>0.1 (0.1)</td>
<td>0.0</td>
</tr>
<tr>
<td>MB100W2R2</td>
<td>0.19</td>
<td>0.253</td>
<td>0.08</td>
<td>0.066</td>
<td>0.3</td>
<td>2</td>
<td>1.6 (22)</td>
<td>0.008 (0.027)</td>
</tr>
</tbody>
</table>
Table 4.2: Experimental design parameters for the prescribed surface heat and moisture flux variation. These experiments are to test the mixed-layer humidity scale parameter with various Bowen ratio. The units of $\langle F_{w\theta} \rangle$ and $A_{w\theta}$ are $Kms^{-1}$ and the units of $\langle F_{wr} \rangle$ and $A_{wr}$ are $gkg^{-1}ms^{-1}$. In the columns of cloud water and rain water, the number is the maximum value over 180-240 min. and the number in the parenthesis is the total value summed over 180-240 min.

<table>
<thead>
<tr>
<th>Case</th>
<th>$\langle F_{w\theta} \rangle$</th>
<th>$\langle F_{wr} \rangle$</th>
<th>$A_{w\theta}$</th>
<th>$A_{wr}$</th>
<th>$B$</th>
<th>$R$</th>
<th>Cloud water (gKg$^{-1}$)</th>
<th>Rain water (gKg$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>MB000W2</td>
<td>0.19</td>
<td>0.253</td>
<td>0.00</td>
<td>0.000</td>
<td>0.3</td>
<td>-</td>
<td>2.0 (50)</td>
<td>0.012 (0.08)</td>
</tr>
<tr>
<td>MB000W1</td>
<td>0.19</td>
<td>0.095</td>
<td>0.00</td>
<td>0.000</td>
<td>0.8</td>
<td>-</td>
<td>0.0</td>
<td>0.0</td>
</tr>
<tr>
<td>MB000D1</td>
<td>0.19</td>
<td>0.066</td>
<td>0.00</td>
<td>0.000</td>
<td>1.2</td>
<td>-</td>
<td>0.0</td>
<td>0.0</td>
</tr>
<tr>
<td>MB000D3</td>
<td>0.19</td>
<td>0.023</td>
<td>0.00</td>
<td>0.000</td>
<td>3.3</td>
<td>-</td>
<td>0.0</td>
<td>0.0</td>
</tr>
</tbody>
</table>
Chapter 5

Conclusions and Future Plans

From observational and numerical studies of data collected during IHOP_2002 and using a recently-developed LES, new insight has been provided the ABL structure over a range of mesoscale surface heterogeneities. Based on these results, the key questions raised in the introduction are answered:

- In the ABL over relatively weak surface heterogeneity on the 10-km scale, the mixed-layer similarity relationship still works, even though the similarity is constructed based on the assumption of the homogeneous ABL conditions. The ABLs observed over mesoscale surface heterogeneities, using aircraft and surface flux facilities, include all scales of fluctuations. After filtering out mesoscale fluctuations from the total fluctuations in the observed ABLs with high-pass spatial filters, the vertical profiles of dimensionless variances of velocity components ($u$, $v$, $w$), potential temperature ($\theta$), and water vapor mixing ratio ($r$) are fit to the mixed-layer similarity curves. For simulated ABLs with a low surface flux variation amplitude, 0.02-0.04 Kms$^{-1}$ (50 Wm$^{-2}$ or lower), similar to those observed during IHOP_2002, the microscale (turbulent) variances and vertical heat fluxes are identical to values from a homogeneous and quasi-stationary ABL when filtered with an appropriate wave-cutoff filter. Also for the
ABLs simulated with a low amplitude of surface heat flux variation, if an appropriate high-pass wave cutoff filter is applied, a JFD parameterization of $\theta$ and $r$ (e.g., Berg and Stull, 2004), which is built based on the assumption of a homogeneous ABL, still can be used. Therefore, one concludes that in the ABL with an amplitude of surface heat flux variation in the observed range, mixed-layer similarity still works.

- **In the ABL over mesoscale surface heterogeneity, the mesoscale vertical flux, the product of mesoscale fluctuations of vertical velocity ($w$) and a scalar ($\theta$ and $r$) (Chen and Avissar, 1994a), is found to be much less significant than the turbulent flux due to the absence of mesoscale $w$ fluctuation.** For all the simulated ABLs with a surface flux variation amplitude of $0.0-0.2 \text{ Kms}^{-1}$ ($0-250 \text{ Wm}^{-2}$), the spectral density peak at the imposed surface heterogeneity scale is absent in the $w$ spectra. In contrast to $w$, both $\theta$ and $r$ obviously respond to the imposed mesoscale surface heterogeneity. Thus, instead of the mesoscale flux, the interscale component of the vertical flux (which is the product of microscale (turbulent) $w$ and mesoscale fluctuation of a scalar ($\theta$ or $r$)) becomes significant as the amplitude of surface flux variation increases. However, in general, the microscale (turbulent) vertical flux remains the most significant contributor to total flux, and vertical transport due to mesoscale $w$ is always negligible in the simulated ABLs with a surface flux variation amplitude of $0.0-0.2 \text{ Kms}^{-1}$ ($0-250 \text{ Wm}^{-2}$). This more significant microscale (turbulent) vertical flux is demonstrated also in the observational study. In the spectra of $w$ from the observation during IHOP_2002, the mesoscale peak is absent, although there exist significant mesoscale peaks in
the spectra of $\theta$ and $r$. Consistently, in the cospectra of $w-\theta$, and $w-r$, which are measured by repeated aircraft passes at 65 m agl on the day, when the surface-heterogeneity-induced mesoscale circulations were likely present, microscale (turbulent) vertical flux accounts for more than 80% of total vertical flux.

- **The ABL over surface heterogeneity at a 10-km-order scale has a considerably different structure, depending on the heterogeneity intensity (in this study, the amplitude of surface heat flux variation).** Specifically, the ABLs over the mesoscale surface heterogeneity can be categorized into two groups as functions of the amplitude of surface heat flux variation. One group contains the ABLs with a low variation amplitude (100 Wm$^{-2}$ or lower); the other contains the ABLs with a high variation amplitude (higher than 100 Wm$^{-2}$). The observed ABLs, which have a low amplitude of surface heat flux variation, show persistent microscale (turbulent) thermals coexisting with mesoscale fluctuations of $\theta$ and $r$. Consistently, in the ABLs simulated with a low amplitude of surface heat flux variation, the nonlinear transfer of variance of the imposed heterogeneity scale to a smaller scale cannot be ignored for $\theta$ and $r$, whereas $w$ variances are perturbed the least from the homogeneous ABL. After removing the mesoscale fluctuations, however, the vertical profiles of microscale (turbulent) variances fit to the mixed-layer similarity curves. Unlike in the ABLs with a low variation amplitude, in the ABLs simulated with a high amplitude of surface heat flux variation (higher than 100 Wm$^{-2}$), the assumption of a quasi-steady state cannot be applied due to the temporal oscillation of the generated mesoscale horizontal flow. The temporally oscillating mesoscale flows result from the advection of temperature, which is
cold advection at lower levels and warm advection at upper level. Thus, in the ABL with a high variation amplitude, the domain-averaged static stability and the magnitude of $w$ variance are also controlled by the generated mesoscale horizontal flows.

In addition to the previous answers to key questions, results that answer also questions raised in the introduction, but pertinent only to each chapter are presented in the following. First, concerning questions related to the observational study in Chapter 2 the answers are as follows:

- Surface heterogeneity is a combined representation of multiple heterogeneous factors. Specifically, over the study area during IHOP_2002, mesoscale surface heterogeneity shown in the surface temperature distribution is mainly caused by the combination of soil moisture and soil thermal properties. The closure of surface energy balance is estimated as $Rn - G \approx H + LE$ (where $Rn$ is net radiation, $G$ is heat flux into the soil, $H$ is sensible heat flux, and $LE$ is latent heat flux). Assuming available energy ($Rn - G$) is similar between surfaces over moist soil and dry soil, most of the available energy over dry soil is used to produce $H$. Meanwhile, over two neighboring surfaces with a similar amount of $LE$ (which indicates a similar amount of soil moisture), the smaller $G$ is due to different soil thermal properties and results in more sensible heat flux $H$ by increasing the available energy ($Rn - G$).

- Relatively small-scale surface heterogeneities influence the whole ABL on the days when the ABL is likely dominated by buoyancy-generated turbulence, given
the smallness of the Obukhov lengths. On these days the vertical profiles of dimensionless variances of wind components \((u, v, w)\), potential temperature \((\theta)\), and mixing ratio \((r)\) from aircraft measurements exhibit significant deviations from the similarity curves due to the effect of mesoscale circulations generated by the surface heterogeneity. In addition in the low-level ABL on these days, the tilt of the joint frequency distribution (JFD) of \(\theta-r\) changes with height from slightly positive to negative, which also implies the generation of mesoscale circulations (warm, dry updraft and cool, dry downdraft).

Second, concerning questions related to the simulated fair-weather dry ABLs (about 1130-1430 LST) over the mesoscale surface heterogeneity in Chapter 3, the answers are as follows:

- The ABL over the mesoscale surface heterogeneity generates mesoscale horizontal velocity, the magnitude of which is proportional to the intensity of the surface heterogeneity (the amplitude of surface heat flux variation). ABLs with a higher amplitude of surface heat flux variation at a mesoscale have a more significant temperature advection due to the generated, stronger mesoscale horizontal flow. The temperature advection in the lower ABL is cold advection. Thus, temperature advection reduces the mesoscale temperature gradient between the warmer region (the region over the surface having heat flux above the domain average) and the cooler region (the region over the surface having heat flux below the domain average). Once the temperature gradient is reduced by temperature advection, the generated mesoscale horizontal flows are weakened. The weakened
mesoscale flows will then reduce the temperature advection and create an environment for the temperature gradient to be reintensified by the mesoscale variation of surface heat flux. If the amplitude of surface heat flux variation is high enough to generate a horizontal velocity greater than the critical velocity \((V_c)\), an oscillatory interaction between the generated mesoscale horizontal flows and an temperature advection is initiated. In other words, the ABL, simulated with an amplitude produces variation large enough to generate a mesoscale horizontal flow greater than \(V_c\), contains temporally oscillating mesoscale horizontal flows and thus does not satisfy the quasi-steady state condition of similarity theory. However, in an ABL simulated with a low amplitude (100 Wm\(^{-2}\) or lower) of surface heat flux variation, the assumption of a quasi-steady ABL still can be applied.

Finally, concerning the question associated with the moist ABLs simulated with a low amplitude (100 Wm\(^{-2}\) or lower) of surface heat flux variation in Chapter 4, the answer is as follow:

- Horizontal transport of moisture by the generated mesoscale horizontal flows can be significant from a perspective of the environment for moist convection. Even for an ABL with a low amplitude (here 100 Wm\(^{-2}\) or lower) of surface flux variation, the amount of moisture transported from the cooler region (the region over the surface having heat flux below the domain average and moisture flux above the domain average) to the warmer region (the region over the surface having heat flux above the domain average and moisture flux below the domain
average) is as much as 130% of the amount of moisture supplied from the surface in the warmer region. Thus, negligence of the horizontal transport of moisture causes a serious error in conclusions about the link between the ABL structure over a mesoscale surface heterogeneity and the development of cloud and precipitation.

Future work will focus on extending the present study to results that are of practical use for improving the simulation of ABL processes in a larger scale model such as a GCM or an MM, especially in the context of surface-ABL-moist convection interactions. The nonlinear processes involved in the interaction would be studied mainly by investigating ABLs simulated with the LES, but with more realistic surface and atmospheric conditions, as in the following examples:

- The study, initiated here on the contribution of surface heterogeneity to moist convection would be extended to include different overlying atmospheric conditions, such as capping inversion intensity and above-the-ABL dryness. Thus, in the argument about the link of the ABL structure over the surface heterogeneity to cloud initiation and development, the effect of the horizontal transport of moisture would be included under different upper atmospheric conditions as well as under different surface conditions (e.g., the Bowen ratio, the ratio of horizontal variation of $LE$ to $H$).

- The results from the previous studies would be tested for ABLs simulated with an experimental setup which considers currently ignored background weather conditions. First, the effects of background wind would be included. The results
would be summarized as a function of the direction and magnitude of background wind. Second, the effects of real-case large-scale flows (larger than the LES domain) would be included by using the flows from an MM as LBCs instead of using cyclic LBCs. Through this numerical experiment, one can investigate the coupled nonlinear processes of the surface heterogeneity and large-scale ambient meteorology in the ABL.

- The effects of a variable terrain elevation, one of the mechanical factors that create heterogeneous surface conditions, would be included in the ABL simulations. In other words, the surface conditions would be prescribed using both mechanical and thermal factors. Here, the combined effects of the two different surface heterogeneity factors (mechanical and thermal-hydrological factors) on the ABL structure can be investigated.

- In this dissertation study, one-dimensional surface heterogeneity was prescribed to make the simulation results easier to interpret. Eventually, the results from one-dimensional surface heterogeneity would be extended by using two-dimensional surface heterogeneity.

- Finally, the conclusions obtained from the LES experiments would be compared to results from a larger scale model (an MM or a GCM) specifically from a perspective of the ABL parameterization in a larger scale model. Then, the effect of improved ABL parameterization using the new findings obtained from this research on the results from a larger scale model would be evaluated.
This dissertation work investigated the detailed structure of the ABL over surface heterogeneity at a 10-km-order scale from a perspective of the intensity of the mesoscale surface heterogeneity as a first step to the development of an appropriate treatment of surface heterogeneity effects in a larger scale model (an MM or a GCM). For the results to be practically applied to the ABL processes in a larger scale model, this study should be extended to include more realistic surface and atmospheric conditions, as suggested previously. Further, in the context of surface-ABL-moist convection interactions, the processes of microphysics and radiation also should be involved.
Bibliography


VITA

Song-Lak Kang

Education:
- Ph.D. (2001-expected in 2007), Department of Meteorology, Pennsylvania State University, University Park, PA
- M.Sc. (1996-1998), Department of Atmospheric Sciences, Yonsei University, Seoul, Korea
- B.Sc. (1988-1992), Department of Atmospheric Sciences, Yonsei University, Seoul, Korea

Employment and Professional Experiences:
- 2003-present Research Assistant, Department of Meteorology, Pennsylvania State University
- 1996-2001 Research Staff, Numerical Weather Prediction Division, Korea Meteorological Administration (KMA)

Awards:
- Korean Government Fellowship, 2001-2003

Publications (including Preprints):


