

# Addressing the Enigma of the Stable Boundary Layer

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# REPORT DOCUMENTATION PAGE

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| <b>14. ABSTRACT</b><br><p>This report summarizes major achievements from the research conducted under the 6.1 NRL base project entitled “Addressing the Enigma of the Stable Boundary Layer” by researchers from NRL code 7530. The stable atmospheric boundary layer (SABL), forming over a cold land, water, ice or snow surface, is characterized by intermittent turbulence, transient wavelike motions, and strong response to surface properties, and remains the least understood element of the atmospheric boundary layer. The overarching objective of this project is to gain new insight into small-scale turbulent processes that regulate the SABL over land and waters. Specifically, we have explored the fundamental dynamics associated with i) turbulence intermittency generation and turbulence-wave-shear instability interactions in the SABL, ii) stable marine boundary layer over ocean waves, and iii) stable internal boundary layers over a heterogeneous surface. Significant findings from this research have been presented in workshops and scientific meetings and/or published in top-tier peer-reviewed journals. Some new findings have high transition potentials. For example, the SABL over waves study and the new scaling theory over coastal waters imply that the air-coupling parameterization in Navy’s NWP models and EM duct prediction tools need to be improved over ocean waves and over coastal waters.</p> |                         |                                                |                                                      |                                                                           |                                                     |
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## EXECUTIVE SUMMARY

This report presents major achievements from research conducted under the 6.1 NRL base project entitled “Addressing the Enigma of the Stable Boundary Layer” by a group of researchers from NRL code 7530. The stable atmospheric boundary layer (SABL) forms over a cold land, water, ice or snow surface associated with radiative cooling or warm air advection. The SABL, characterized by intermittent turbulence, transient wavelike motions, and strong response to surface properties, remains the least understood element of the atmospheric boundary layer. The overarching objective of this project is to identify yet-to-be discovered key fine-scale processes in the SABL, and address science issues such as shear instability near a solid surface, intermittent turbulence and internal wave generation, turbulent adjustment to a wavy or heterogeneous surface, and their roles in regulating the SABL. These issues are of fundamental scientific importance and great significance to Navy battlespace environment forecast capability.

These scientific challenges have been tackled through a combination of extra-high resolution large eddy simulations (LESs) using a state-of-the-science LES model, analysis of a wealth of new observations from field campaigns such as Coupled Air Sea Processes and Electromagnetic Ducting Research (CASPER), air-sea coupled mesoscale model simulations and theoretical analysis. Recent advances in computing power have made it possible to simulate the intermittent turbulence and internal wave-turbulence interactions in a SABL using LES models with model grid spacings down to a few meters to explicitly resolve internal waves and shear instability processes, energy-containing turbulent eddies and interactions among them. New observations from CASPER-East and CASPER-West have presented a unique opportunity for us to investigate fundamental processes in the SABL over coastal waters, and to validate our LES simulations. Major achievements under this project can be summarized as follows:

- Kelvin-Helmholtz instability (KHI) in and above the SABL has been investigated using large eddy simulations which explicitly resolve KH billows, secondary shear instability, large turbulent eddies and the interaction among them. It has been demonstrated that KHI generates internal waves and intermittent turbulence and significantly modify the SABL characteristics through vertical transport of momentum, heat and moisture. Some results from this effort have been reported in a peer-reviewed journal paper and a second manuscript based on this research has been submitted.
- Interaction between the Arctic SABL and snow or ice surface has been examined utilizing a new snow physics package recently has been put into COAMPS. Substantial improvements have been found in variables such as 2-m dew-point temperature and relative humidity, sea-level pressure, and 10-m wind speed and wind direction, which have significant impact on the SABL characteristics. A manuscript based on this research has been submitted to a peer-reviewed journal and is currently under revision.

- Mesoscale variability of the stable marine boundary layer (SMBL) can be significantly disturbed by coastal terrain and (or) islands. SMBL events documented offshore of southern California have been investigated based on two-way air-sea coupled COAMPS simulations. Main findings from this research have been reported in a peer-reviewed paper published on the Journal of Geophysical Research: Atmosphere.
- In order to address science issues associated with the SABL over a heterogeneous surface, we have developed and tested two different modeling approaches, namely a) the Lagrangian modeling strategy with a novel differential advection correction factor b) a fringe method to provide a turbulent upstream boundary condition. This new capability allows us to examine the boundary layer and EM duct variation over SST fronts and coastal waters.
- Stable internal boundary layer (SIBL) over coastal waters have been studied using field observations, large eddy simulations, and COAMPS simulations. The classical similarity theory has been found problematic in the littoral zone and a new scaling theory has been proposed, which has implications on EM duct prediction and air-sea coupling parameterizations in Navy's NWP models over coastal waters.
- Impact of ocean waves on SMBL structure and evaporation ducts has been explored utilizing idealized LES simulations that explicitly resolve both ocean waves and turbulence in the SMBL. We find that the vertical distribution of scalars in a SMBL can be dramatically changed by ocean waves and the current MOST-based air-sea coupling parameterizations and EM duct prediction algorithms are highly questionable over ocean waves.
- We have contributed to the development, implementation and testing of the COAMPS optical turbulence (Cn2) module, which provides Navy a new optical turbulence forecast capability in the SMBL and beyond.

In summary, this research has advanced our understanding of a range of fundamental science issues in the SABL associated with shear instability, intermittent turbulence and internal wave generation mechanisms, SMBL over ocean waves, and SMBL over coastal waters. In addition, some results are highly Navy relevant and have the potential of improving Navy's battlespace environment forecast capability, especially over Polar Regions and over coastal waters. For example, the new scaling theory will lead to more accurate characterization of EM ducts under stable conditions over littoral zones, which are crucial for modeling EM propagation loss, and ultimately, for Navy operations involving communications, surveillance, and electronic warfare. The new insights of SMBL over waves and over coastal waters provide necessary guidance for future air-sea coupling parameterizations for next generation Navy numerical models such as Navy Environmental Prediction sysTem Using the NUMA (NEPTUNE).

## ADDRESSING THE ENIGMA OF THE STABLE BOUNDARY LAYER

### INTRODUCTION

The stable atmospheric boundary layer (SABL) forms over a cold land, water, ice or snow surface associated with radiative cooling, or warm air advection. At any given time of the day, the SABL covers approximately fifty percent of the Earth's surface. While our understanding of neutral and convective boundary layers have advanced significantly, the SABL, characterized by intermittent turbulence, wavelike motions, and micro-fronts with abrupt changes in temperature and winds, remains the least understood element of the earth's atmospheric boundary layer. The poor understanding of the SABL is partly due to the relative dearth of systematic observations and challenges in modeling small-scale processes in the SABL using large eddy simulation (LES) models. The classical Monin-Obukhov similarity theory (MOST) is often found to be unreliable for the SABL. Observations from the Coupled Air Sea Processes and Electromagnetic Ducting Research (CASPER, 2015 and 2017) presented an unprecedented opportunity for documenting the fundamental physical processes in the SABL and for validating our LES simulations. CASPER includes two-phased field observations, taking place offshore of North Carolina and offshore of California, respectively. In some SABL cases, CASPER documented significant deviations of the observed wind, temperature and moisture profiles from those predicted by MOST, which led to dramatically different electromagnetic (EM) duct characteristics, highlighting the necessity and urgency for better understanding of the SABL dynamics and developing new theory for the surface layer scaling.

LES models have been proved to be powerful tools for investigating physical processes in a neutral or convective boundary layer. However, LES has yet to be fully exploited for the SABL largely due to the computational expenses that are only now becoming tractable. It requires extremely high resolutions ( $< 5$  m) to resolve micro-scale structure and intermittent turbulence in a SABL. We are entering an exciting new era of opportunity in which high-fidelity LES simulations for the first time can significantly advance our understanding of the SABL. NRL now has a state-of-the-science LES model (NCAR/NRL-LES) uniquely suited for performing high-fidelity simulations of micro-scale processes in a SABL.

The SABL is not only of great scientific importance, but also has significant implications on Navy operations. It tends to suppress vertical mixing of moisture, enhance vertical temperature and moisture gradients, and therefore favor the formation of EM ducts. This research potentially leads to more accurate characterization of the SABL and improved understanding of EM ducts under stable conditions, which are crucial for modeling EM propagation path loss, and ultimately, Navy operations. The improved understanding of the SABL has the potential to improve surface flux and boundary layer parameterizations for the Navy mesoscale and global models. In addition, better forecasts of fog, low level clouds, and visibility for Navy operations also demand improved understanding of the SABL dynamics. In summary, the proposed research has the potential to advance our understanding of fundamental dynamics in the SABL and address multiple Naval S&T Focus Areas.

The research reported here includes two broad categories, namely

- shear instability, intermittent turbulence and internal gravity wave generation, and SABL-snow coupling, and
- internal boundary layer (IBL) physics and simulations associated with the SABL over a heterogeneous surface with emphasis on IBL over coastal waters.

For the results that have been reported in peer-reviewed journal articles, we only include a brief description of the research along with abstracts and paper references. More details can be found from the corresponding references. The remainder of this report is organized as follows:

Section 1: LES simulation-based study of shear instability, Kelvin-Helmholtz billows, and their interactions with the SABL. Part I of this study (1.1) has been reported in a peer-reviewed paper (1.1) and a manuscript based on Part II (1.2) has been recently submitted to a peer-reviewed journal.

Section 2: Interaction between the SABL and a snow or ice surface. A manuscript has been submitted to a peer-reviewed journal and is currently in revision.

Section 3: Study of SABL mesoscale variability over coastal areas (published)

Section 4: This section covers development and testing of new LES modeling capabilities (4.1), study of the convective internal boundary layer (4.2, published), research on stable internal boundary layer (4.3, published) and a new scaling theory for the stable internal boundary layer (4.4, manuscript in preparation).

Section 5: Influence of ocean waves on the marine boundary layer and EM ducts (published).

Section 6. Optical turbulence ( $C_n^2$ ) in the SABL (published).

Section 7. Summary and concluding remarks.

## 1. KH BILLOW AND SABL INTERACTION

The occurrence of Kelvin-Helmholtz billows (KHBs) and their turbulent breakdown associated with two different wind shears have been investigated. In part I of this study, development of KHBs in layer with a preexisting elevated wind shear located above a stable atmospheric boundary layer (SABL) associated with either synoptic scale (e.g., frontal systems) or mesoscale (e.g., land-sea circulations) and their interaction with the underlying SABL have been examined using large eddy simulations. Findings from Part I has been reported in a peer-reviewed article (see abstract and reference in 1.1). In part II, KH instability (KHI) takes place in an Ekman type SABL associated with an inflection point inside the SABL. Results from part II are presented in section 2.2 and are also included in a manuscript recent submitted to a peer-reviewed journal for consideration of publication.

### 1.1 Impact of Elevated Kelvin-Helmholtz Billows on the SABL

Findings from this part of the research have been reported in a peer-reviewed journal article:

Jiang, Q., 2021: Impact of elevated Kelvin-Helmholtz billows on a stable boundary layer. *J. Atmos. Sci.*, 78, 3983-3999.

#### ABSTRACT

The impact of Kelvin-Helmholtz billows (KHBs) in an elevated shear layer (ESL) on the underlying atmospheric boundary layer (BL) is examined utilizing a group of large-eddy simulations. In these simulations, KHBs develop in the ESL and experience exponential growth, saturation, and exponential decay stages. In response, strong wavy motion occurs in the BL, inducing rotor circulations near the surface when the BL is stable. During the saturation stage, secondary instability develops in the ESL and the wavy BL almost simultaneously, followed by the breakdown of the quasi-two-dimensional KH billows and BL waves into three-dimensional turbulence. Consequently, during and after a KH event, the underlying BL becomes more turbulent with its depth increased and stratification weakened substantially, suggestive of significant lasting impact of elevated KH billows on the atmospheric BL. The eventual impact of KHBs on the BL is found to be sensitive to both the ESL and BL characteristics.

### 1.2 GROWTH AND BREAKDOWN OF KELVIN-HELMHOLTZ BILLOWS IN THE SABL

#### 1.2.1 Background and motivation

Kelvin-Helmholtz instability (KHI) occurs frequently in stably stratified flows in the presence of strong vertical wind shear (Helmholtz 1868, Kelvin 1871). The necessary conditions for KHI to occur in a continuously stratified flow include a) the presence of an inflection point (i.e., the level where  $d^2U(z)/dz^2 = 0$ , where  $U(z)$  is the wind speed at level  $z$ ), and b) the minimum Richardson number in the shear layer is less than 0.25. A KHI process in the atmosphere lasts from a few minutes to a few dozens of minutes (e.g., Newsom and Banta 2003), including an initial growth phase with roll-up of the shear layer (or vortex sheet) into wave-like undulations, the formation of mature KH billows phase, and turbulent breakdown of KHBs phase associated with overturning of density surfaces or other types of secondary instability (SI). In addition to many laboratory experiments, the secondary instability in two or three dimensional laminar flows had been extensively investigated by numerical studies for low-to-moderate Reynolds number fluids using direct numerical simulation codes (DNS). Several SI types have been

identified from numerical and laboratory studies, namely localized core vortex instability, secondary core deformation instability, secondary convective instability, stagnation point instability, secondary shear instability, secondary vorticity bands instability, subharmonic vortex pairing, knots and tubes (see Thorpe 2012 for a review). More recently, the knots and tubes dynamics was numerically investigated in a pair of papers by Fritts et al. (2022a, b).

KHI has been frequently cited as an important source for internal gravity waves (IGWs) and intermittent turbulence in the stable atmospheric boundary layer (SABL, see Sun et al. 2015 for a review). Besides known as “KH waves” or “KH billows”, KHI-induced wave-like undulations are also frequently referred to as “vorticity waves”, as they are generated by roll-up of vortex sheets, to distinguish them from buoyancy-driven IGWs in a stably stratified atmosphere. IGWs may trigger KHI in the atmosphere through locally enhancing wind shear and reducing local Richardson number (e.g., Jiang et al. 2010), and on the other hand, KHI may generate IGWs (Dong et al. 2023). Recent remote-sensing observations suggested that KHI frequently occurs in the SABL over Polar Regions. For example, in the winter months, KH billows and braids were documented in 40% of the time by a high-resolution sodar at an Antarctic site (Petenko et al. 2016, 2019). Zaitseva et al. (2022) evaluated the impact of wave-like motions on turbulence characteristics using long-term sodar measurements and found that the passage of 30% of KH billow trains was accompanied with increase of turbulent kinetic energy and heat and momentum fluxes. The role KHI plays in vertical mixing of momentum and scalars has been the subject of numerous studies, mostly for low to moderate Reynolds number fluids with a few exceptions. Through analysis of a KHI event documented in the SABL during the Cooperative Atmospheric Surface Exchange Study 1999 field campaign (CASE99), Newsom and Banta (2003) found that, associated with the onset of KH waves, a sudden increase in downward wave-momentum flux was accompanied by a sharp reduction in shear near the critical level. The numerical study by Smyth (2003) suggested that KH billows may be important for vertical mixing in stratified fluids. More recently van der Linden et al. (2020) noted, in a very stable boundary layer, KHI-induced bursting of momentum flux occurs that erodes wind shear near the ground surface.

Majority of the KHI studies focused on KH billows developing in an initially sheared laminar flow far away from any boundaries with a few exceptions. Development of shear instability in a stably-stratified shear layer in the presence of turbulence has been numerically examined by Kaminski and Smyth (2019) and they found that KH billow structure can be suppressed by the pre-existing turbulence and be eventually eliminated if the initial turbulence is strong enough. The impact of a lower boundary on KHI and turbulent breakdown of KH billows was recently studied by Liu et al. (2023) using ensembles of direct numerical simulations (DNS). They noticed that the transition into turbulence and the subsequent mixing were modified by the nearby lower boundary, where a no-slip boundary condition was applied. The objective of this study is to shed light on the KHI development, KH billow characteristics, secondary instability, turbulent breakdown of KH billows and their impact on vertical mixing in a stable atmosphere boundary layer through analyzing large eddy simulations.

### 1.2.2 LES code and model configuration

The LES code used in this study solves a set of spatially-filtered three-dimensional Boussinesq equations in a rectangular domain with periodic boundary conditions applied along the side walls (e.g., Moeng 1984; Sullivan et al. 1994). The subgrid scale (SGS) stress and fluxes are parameterized following the two-part SGS scheme illustrated in Sullivan et al. (1994). The pair of simulations presented in this study are selected from two dozens of experimental runs with different horizontal domain aspect ratios, domain sizes, and grid spacings.

For the control simulation (i.e., A), the horizontal dimensions of the rectangular domain are  $L_x = 1536$  m and  $L_y = 384$  m with a grid spacing of  $\Delta x = \Delta y = 3$  m. There are 220 vertical levels with the first model level at  $z_l = 2$  m and the model top at  $D = 1200$  m, where a radiation boundary condition (Klemp and Durran 1983)

is applied. The vertical grid spacing increases with height with a constant stretching ratio,  $\Delta z_{i+1} / \Delta z_i = 1.007$ , and the resulted average grid spacing of 2.38 m for the lowest 100 m. For simulation B, the first model level is at  $z_l = 1$  m and the model top at  $D = 1200$  m with a stretching ratio of 1.015. The average vertical grid spacing is 1.54 m in the lowest 100 m. Some control and derived parameters can be found in Table 1.1.

Table 1.1. Control and derived parameters for the two LES runs (A or control and B). From columns 5 to 11, the parameters are first model level ( $z_l$ ), identified inflection point level ( $z_{IP}$ ), KH wavelength ( $\lambda$ ), phase angle ( $\alpha$ ), inflection point level and wavelength ratio, maximum kinetic energy ( $KE_m$ ) and the ratio between mature KH billow core height and wavelength ( $R$ ).

| Exp | Grids    | $\Delta x$ (m) | $U_g$ (m/s) | $z_l$ (m) | $z_{IP}$ (m) | $\lambda$ (m) | $\alpha$ ( $^\circ$ ) | $z_{IP}/\lambda$ | $KE_m$ (m <sup>2</sup> /s <sup>2</sup> ) | $R$  |
|-----|----------|----------------|-------------|-----------|--------------|---------------|-----------------------|------------------|------------------------------------------|------|
| A   | 512x128  | 3.0            | 15          | 2         | 30           | 370           | 14.4                  | 6.2              | 5.1                                      | 0.16 |
| B   | 1024x256 | 2.5            | 9           | 1         | 20           | 247           | 14.3                  | 6.2              | 1.7                                      | 0.12 |

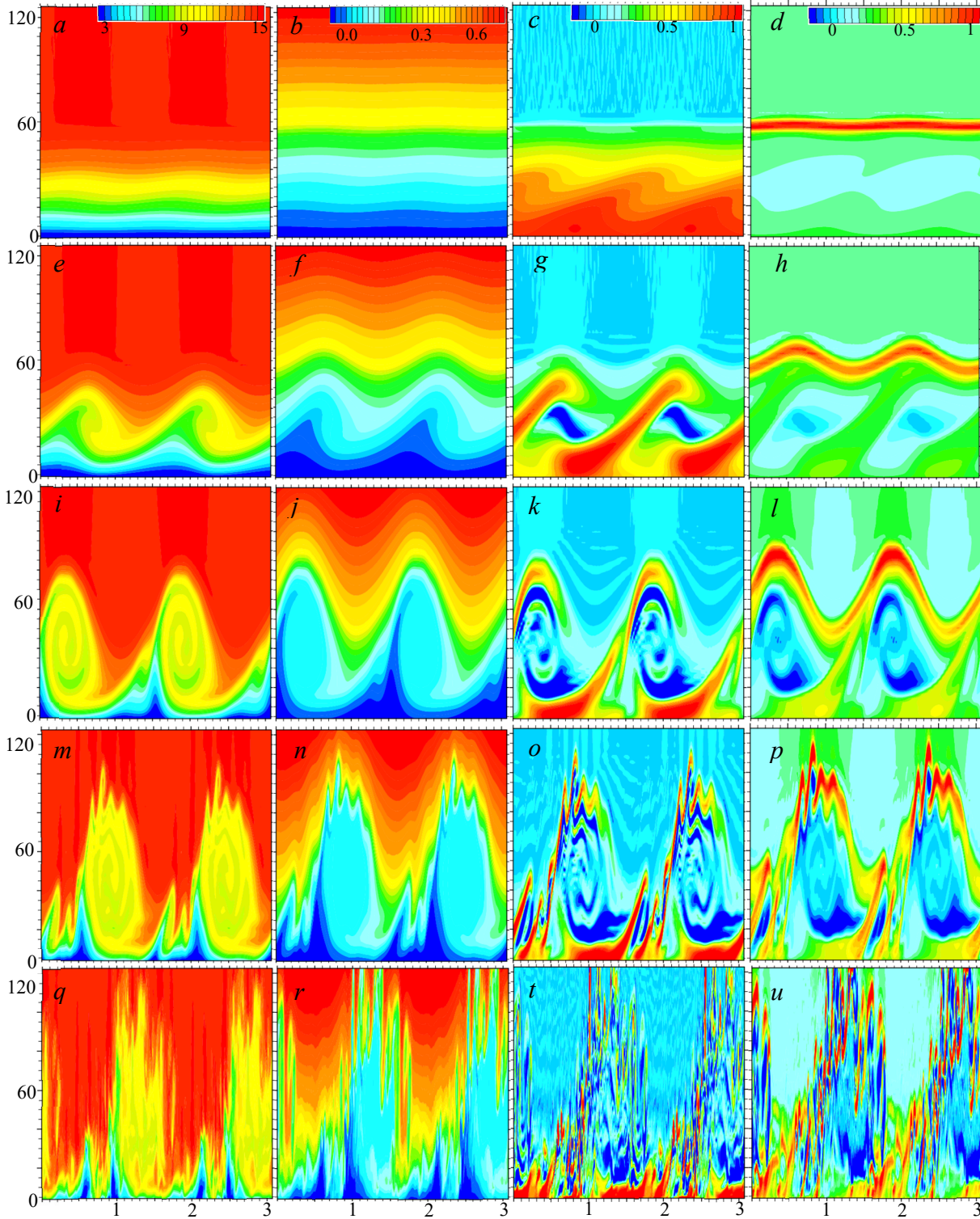


Figure 1.1. Vertical cross-sections of  $u$  (leftmost, color incr. = 0.5 m/s), potential temperature (col 2,  $\theta - \theta_0$ , color incr. = 0.025K), normalized span-wise vorticity (col 3, incr. = 0.05), and normalized buoyancy frequency squared (rightmost, incr. = 0.05) valid at (top to bottom rows)  $t_{KH} = 5, 11, 16, 17.5$  and  $18.5$  min, respectively. The vorticity is normalized by 0.35 (top row), 0.49 (row 2), and  $0.78 \text{ s}^{-1}$  (rows 3-5), respectively and  $N^2$  is normalized by  $0.001 \text{ s}^{-2}$ .

The initial horizontal winds are unidirectional and vertically uniform,  $\vec{V}_0(z) = (U_g, 0)$ , where  $U_g$  is the wind speed in the X-direction. The atmosphere is initially in geostrophic balance with a constant Coriolis parameter  $f = 1.2 \times 10^{-4} s^{-1}$ . For the control simulation, the potential temperature is given by  $\theta(z) = \theta_0 + \Gamma z$ , where  $\theta_0 = 265$  K is the potential temperature at the surface and  $\Gamma = 6 K / km$  is a constant lapse rate. In addition, small random perturbations in winds and potential temperature are introduced at the beginning of each simulation to accelerate the spin-up of the stable boundary layer.

The surface stress ( $\tau$ ) and dynamic heat flux ( $F_\theta$ ) are computed using bulk formula (e.g., Fairall et al. 1996), namely,

$$\tau = \rho_a C_D (u_1^2 + v_1^2) \quad (1.1a)$$

$$F_\theta = -C_\theta \sqrt{u_1^2 + v_1^2} (\theta_1 - \theta_s) \quad (1.1b)$$

where the subscripts  $s$  and  $1$  denote variables at the surface and the 1<sup>st</sup> model level, respectively. The exchange coefficients,  $C_D$  and  $C_\theta$ , are computed following the Monin-Obukhov similarity theory using a constant surface roughness length  $z_0 = 0.1$  m for both the surface stress and temperature flux. The ground temperature linearly decreases with time, i.e.,  $\theta_s(T) = \theta_0 + \alpha t$ , where  $\theta_0 = 265$  K is the initial ground temperature,  $t$  is the integration time, and  $\alpha = -0.5 K / h$  is the ground surface cooling rate. Therefore, the surface heat flux is always negative and the boundary layer is stable. The model integration time interval ( $\Delta t$ ) is assessed at every time step by setting the maximum Courant-Fredrichs-Lewy number to 0.3 [i.e.,  $CFL_{\max} = \Delta t \max(u / \Delta x, v / \Delta y, w / \Delta z) = 0.3$ ] to maintain computational stability. For the control simulation, the time interval varies between 0.055 and 0.062 seconds with an average of 0.059 seconds. It is worth noting that, in majority of previous KHI studies, the model was initialized with an idealized shear layer usually without directional shear. Accordingly, the fastest growth unstable mode can be predicted by solving Taylor-Golden equation, and the horizontal KH phase-line is normal to the flow direction in the shear layer. In this study, the LES model is initialized with a vertically uniform unidirectional geostrophic wind (i.e.,  $U_g$  in the X or  $u$ -wind direction). Vertical wind shears in both X and Y directions develop naturally in an Ekman type of boundary layer until KHI occurs associated with an inflection point. Therefore, the fastest growth mode and its phase line orientation are not known a priori. Especially in a horizontal domain, the development of a KHI mode is dependent on the domain size and its aspect ratio. Some of these issues are discussed in Section 4.

### 1.2.3 KH billows and breakdown in the SABL

In the control simulation (A), a shallow stable Ekman boundary layer develops rapidly. As will be shown, KHI develops associated with an inflection point in the  $v$  profile followed by secondary instability. We start by diagnosing the control simulation with emphasis on the characteristics and evolution of the simulated KH billows as well as their impact on vertical turbulent mixing.

#### a) KH billows and turbulent breakdown

The evolution of KH waves and billows are revealed in the vertical and horizontal cross-sections in Figs. 1.1 and 1.2. Figure 1.1 shows  $u$ -wind, potential temperature ( $\theta - \theta_0$ ), spanwise vorticity (i.e.,

$$\zeta = \frac{\partial u}{\partial z} - \frac{\partial w}{\partial x}), \text{ and the buoyancy frequency squared (i.e., } N^2 = \frac{g}{\theta_0} \frac{\partial \bar{\theta}}{\partial z}) \text{ in a vertical cross-section oriented}$$

along the geostrophic wind (i.e., X) direction. KHI starts around  $t_0 = 40$  min and we use  $t_{KH} = t - t_0$  in the following discussion of the KH billow evolution. At  $t_{KH} = 5$  min, the KH waves become evident, manifested

as vertical undulations of the isotachs, isentropic surfaces, spanwise vorticity patterns, and the boundary layer top inversion (Figs. 1.1a-d). At this early stage, the KH waves are most noticeable in the vorticity cross-section (Fig. 1.1c). As expected, the spanwise vorticity is positive in the boundary layer with a maximum immediately above the surface and decreases with height toward zero at the SABL top level, presumably dominated by the contribution from the vertical wind shear term. The largest vertical undulation of the vertical isosurfaces occurs between 20-40 m, suggesting that the KH layer is centered approximately at  $z = 30$  m.

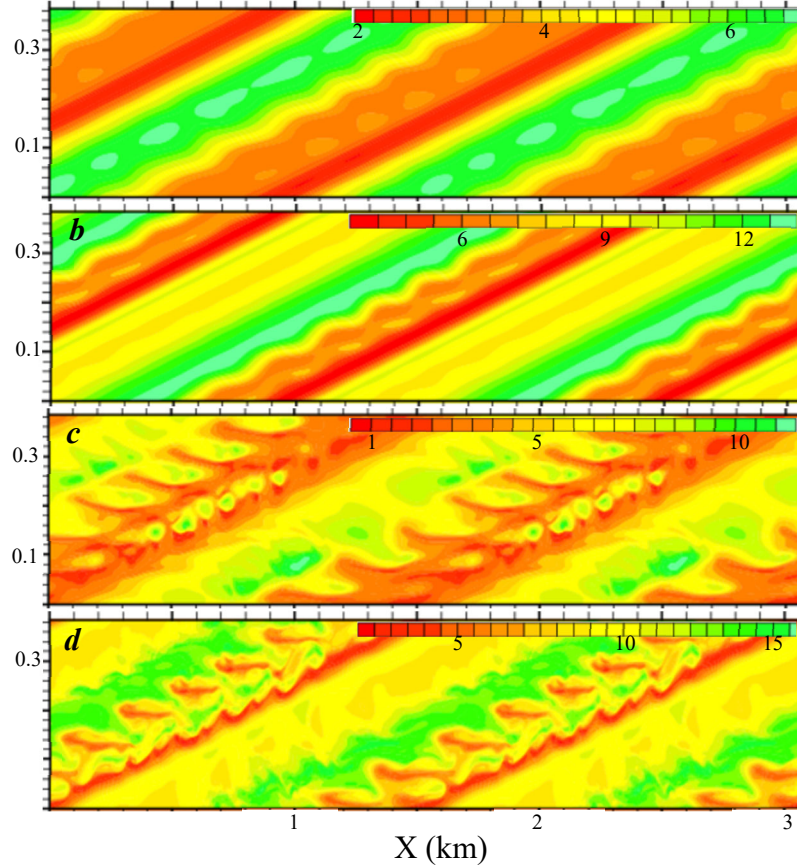


Figure 1.2. Horizontal (X-Y) cross-sections of  $u$  at levels  $z=6$  (a and c) and 21 m (b and d) valid at  $t_{KH} = 15$  and 17 min, respectively. The color shading increments are a) 0.2 m/s, b) 0.6 m/s, c) 0.5 m/s, and d) 0.5 m/s, respectively. The graphics are stretched in y direction by a factor of two.

These unstable KH waves grow in amplitude rapidly with time, and the KH billow cores start forming underneath wave crests in 5 min (i.e.,  $t_{KH} = 11$  min), as evidenced by the overturning of isotach and isentropic surfaces (Figs. 1.1e and f). In Figs. 1.1g and h, the KH billow cores are characterized by reduced vorticity (slightly negative in places) and weakened stratification encompassed by S-shaped braids where both the vorticity and stratification are enhanced. Approximately by  $t_{KH} = 15$  min, the KH billows are fully developed with their cores between 20-80 m (Figs. 1.1i-l). KHI induced vertical undulation extends from the surface up to 120 m and beyond (Fig. 1.1j). The mature billow cores contain multiple rings of alternating wind and temperature anomalies, associated with entrainment of the warmer and faster airflow from above. Similarly, negative vorticity filaments are evident inside the cores along with much enhanced positive vorticity in the lowest 20 m (Fig. 1.1k). According to Fritts et al. (1996), the negative vorticity in KH billow cores is generated by baroclinicity, which also contributes to the enhancement of the positive

vorticity in the braids. There are negative  $N^2$  filaments inside the cores as well, suggesting that the KH cores are convectively unstable (Fig. 1.11).

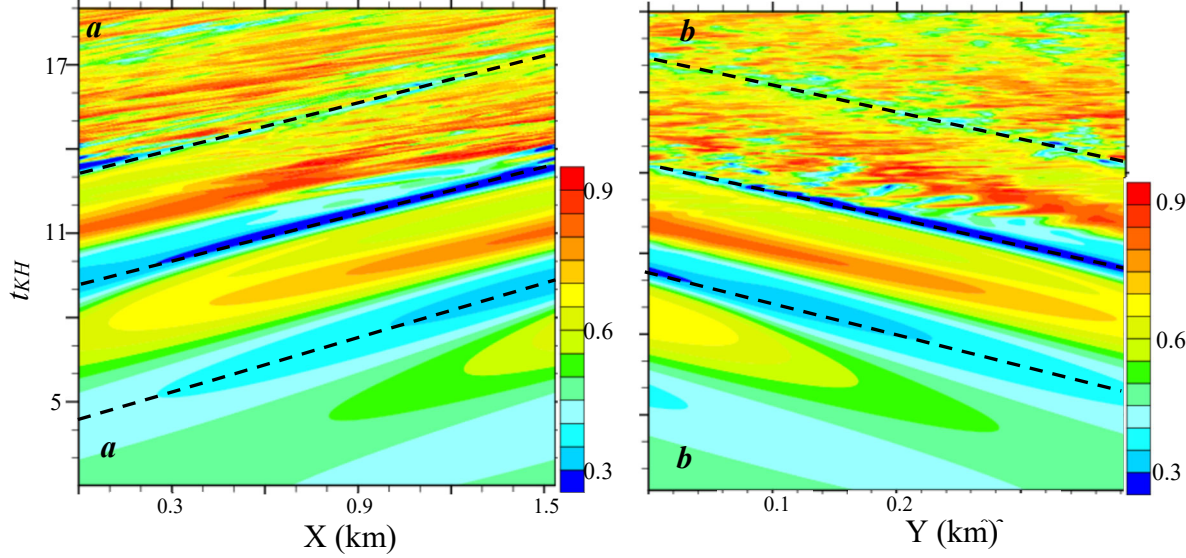


Figure 1.3. Distance-time sections, namely a)  $X-t_{KH}$  and b)  $Y-t_{KH}$ , of  $u$  at  $z = 18.6$  m normalized by  $U_g$ . The dashed lines are used to estimate the KH wave propagation speed.

At  $t_{KH} = 16$  min, small kinks first appear along the S-shaped braids that wrap around the KH billow cores (not shown), implying the onset of the secondary shear instability (SSI). By 16.5 min, several spikes appear along the braids and the wave crests with a horizontal length scale of 100-150 m, suggesting fast growth of SSI (Fig. 1.1e). Negative vorticity and negative  $N^2$  appear in the spikes as well (Figs. 1.1o-p), implying that the secondary KH billows become convectively unstable as well. By  $t_{KH} = 17.5$  min, while the vertical undulations of the isotachs and isentropic surfaces associated with the primary KH waves are still evident, KH cores are virtually destroyed by SSI and become turbulent (Figs. 1.1q-t).

Shown in Fig. 1.2 are the horizontal sections of the  $u$  wind at two different levels during the SSI. Prior to the SSI and KHB breakdown, the KH waves are two-dimensional with the orientation angle between the phase lines and the geostrophic wind (i.e.,  $X$ ) direction  $\alpha \approx 14.4^\circ$  (Fig. 1.2a). The horizontal wavelength (i.e., the distance between two adjacent phase lines) is  $\lambda_H \approx 372$  m. Therefore, the ratio of the KH core depth and wavelength is around 0.16, which is sizably smaller than typical KH billows further away from a solid surface (e.g., 0.2 in Jiang 2021 and Fritts et al. 2014). The smaller KH core depth is likely due to the negative impact of the ground surface, as suggested by Liu et al. (2023).

KH wave induced wind perturbations are substantially larger near the KHB cores than at the surface. Secondary shear instability first develops below each wave crest, where the  $u$ -wind is minimum (Fig. 1.2a), and finer-scale perturbations are noticeably stronger at the  $z = 21$  m level than at the surface (Figs. 1.2a and b). In 2 min, the amplitude of the SSI induced  $u$ -wind perturbations become comparable to that of the primary KH waves (Figs. 1.2c and d). The KH waves propagate towards lower-right while growing in amplitude. At  $t_{KH} = 15$  min, finer-scale features appear along the upwind edges of wave crests (Fig. 1.2b), corresponding to the spikes along braids in the vertical cross-sections (Fig. 1.1d). At  $t_{KH} = 17$  min, the primary KH waves are blurred by finger-like patterns (Fig. 1.2c). The fast-growing fine-scale perturbations along the KH braids, where the wind shear is the strongest (Figs. 1.1 and 1.2), imply that the secondary shear instability is three-dimensional in nature and plays a dominant role in the initial turbulent breakdown of the simulated KH billows, although convective instability likely contributes to the breakdown as well.

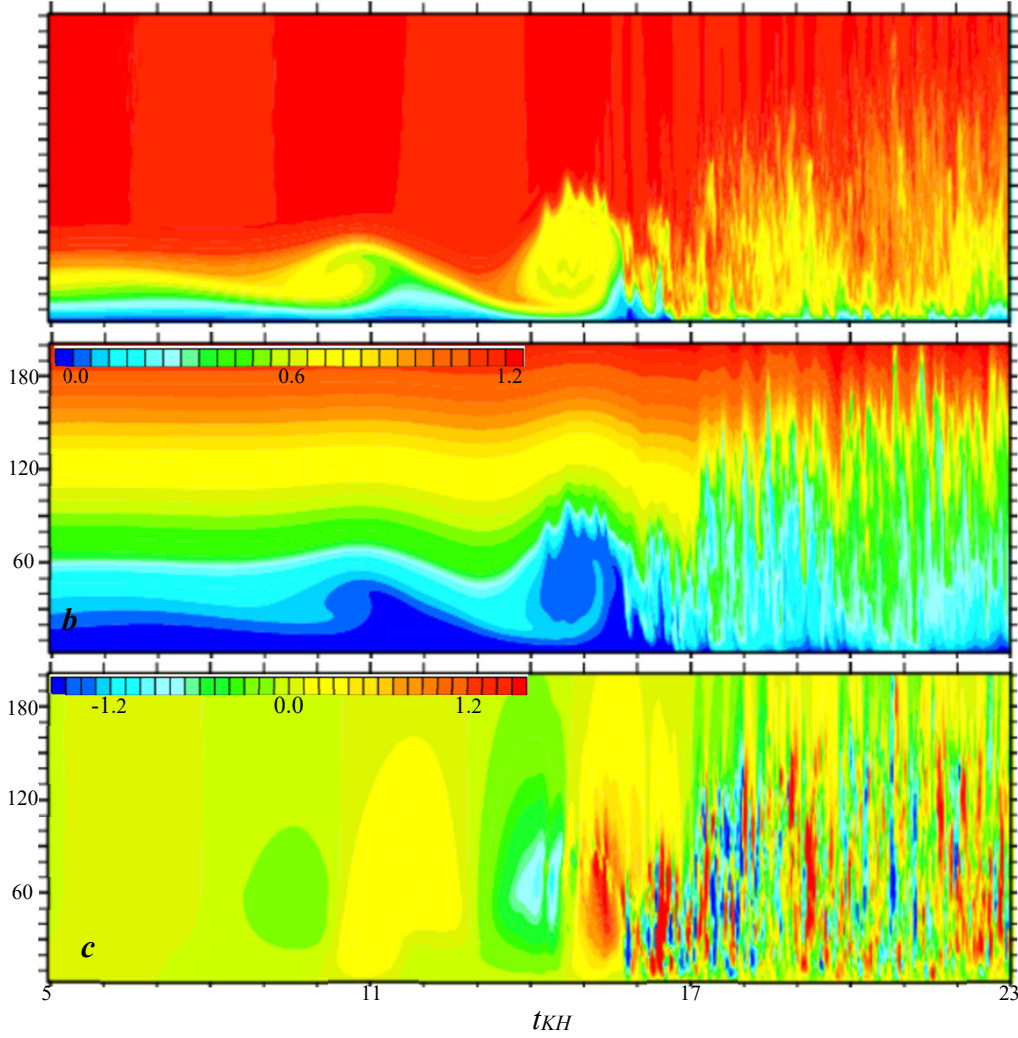


Figure 1.4. Time-height ( $t_{KH}$ - $z$ ) sections of a) u-wind ( $u$ , m/s), b) potential temperature ( $\theta - \theta_0$ , K) and c) vertical velocity ( $w$ , m/s) sampled near the center of the domain. Only the lowest 200 m portion and for  $t_{KH}$  between 5-23 minutes is shown.

Shown in Fig. 1.3 are the distance-time sections of  $u$  at  $z = 18.6$  m. In the X direction, the KH waves propagate along the wind direction while growing in amplitude and subsequently experiencing SSI and turbulent breakdown (Fig. 1.3a). The wave speed (i.e., the slope of patterns indicated by dashed lines in Fig. 1.3a) is around 5.10 m/s. In the Y-direction, the KH wave speed is around 1.35 m/s, implying that the KH waves tend to propagate in the direction perpendicular to the horizontal phase lines (Fig. 1.3b). The growth of KH billows and their breakdown “observed” near the center of the LES domain are revealed in Fig. 1.4. KHI first manifests itself as undulations of the cold sheared layer above the surface (Fig. 1.4a, b), which rolls up in the second cycle (between 8-13 min) with the onset of overturning of the isentropic surfaces under the wave crest. In the third cycle (13-17 min), a mature KH billow is evident with signs of early development of the secondary shear instability (fine-scale spikes along the top edge of the KH billow in Fig. 1.4a and b). Associated with the abrupt breakdown of the KH billow ( $t_{KH} > 17$  min), the turbulent layer thickens substantially with relatively large eddies reaching beyond 200 m above the surface.

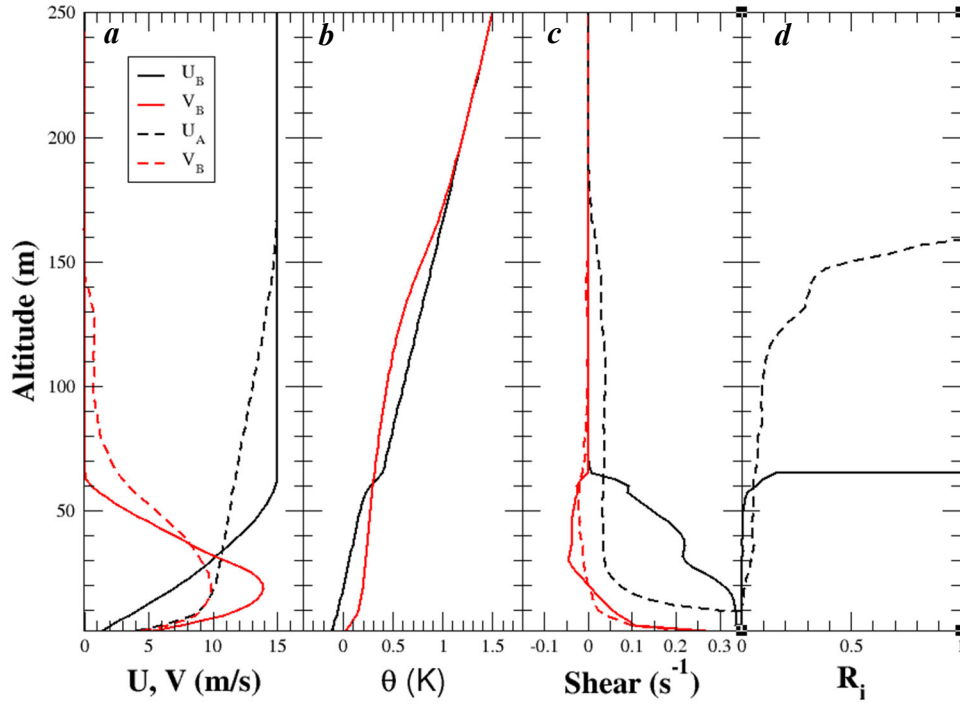


Figure 1.5. Profiles of a)  $U$  and  $V$ , b)  $\theta - \theta_0$ , c) shear (i.e.,  $\partial U / \partial z$  and  $\partial V / \partial z$ ), and d) Richardson number before KH billows fully developed (i.e.,  $t_{KH} = 8$  min, denoted with subscript “B”) and after their breakdown (i.e.,  $t_{KH} = 20$  min, denoted with subscript “A”). The  $V$ -wind is multiplied by 10. An inflection point is located at  $z \sim 30$  m, above the  $V$  maximum before KH.

The domain-averaged profiles of the horizontal winds ( $U$ ,  $V$ ), potential temperature ( $\bar{\theta} - \theta_0$ ), vertical wind shear ( $dU / dz$  and  $dV / dz$ ), and Richardson number ( $R_i = \frac{g}{\theta} \frac{\partial \bar{\theta}}{\partial z} \left( \left( \frac{\partial U}{\partial z} \right)^2 + \left( \frac{\partial V}{\partial z} \right)^2 \right)^{-1}$ ) before and after the KH event are shown in Fig. 1.5. Prior to the KH event, the stable boundary layer is about 60 m deep capped by a thin more stable layer. While the  $u$ -wind and its vertical shear are much larger than its  $v$ -wind counterparts, the vertical shear in  $u$ -wind is positive throughout the boundary layer. Although the vertical shear of the  $v$ -wind is weak, a well-defined inflection point (i.e., where  $d^2v / dz^2 = 0$ ) is present at  $z_{IP} \sim 30$  m. In general, the gradient Richardson number is small ( $Ri < 0.1$ ) in the boundary layer. Therefore, the development of the KH instability is likely associated with the inflection point in the  $v$ -wind. For the  $u$  wind, the depth of the shear layer is the comparable to the SABL depth (i.e.,  $\sim 60$  m). A low-level jet is evident in the  $v$  wind and the depth of the shear layer above the jet level is around 40 m. Accordingly, the KH wavelength and shear layer depth is 6.2 for  $u$  wind and 9.3 for  $v$  wind, respectively, which fall into the range for fastest growing KH mode (i.e., 6-13, e.g., Miura and Pritchett 1982). After the onset of SSI (i.e.,  $t_{KH} > 17$  min), the boundary layer depth extends to 150 m, above which the Richardson number  $> 0.25$ , the wind shear reduces to zero, and a more stable layer is evident between 150-180 m.

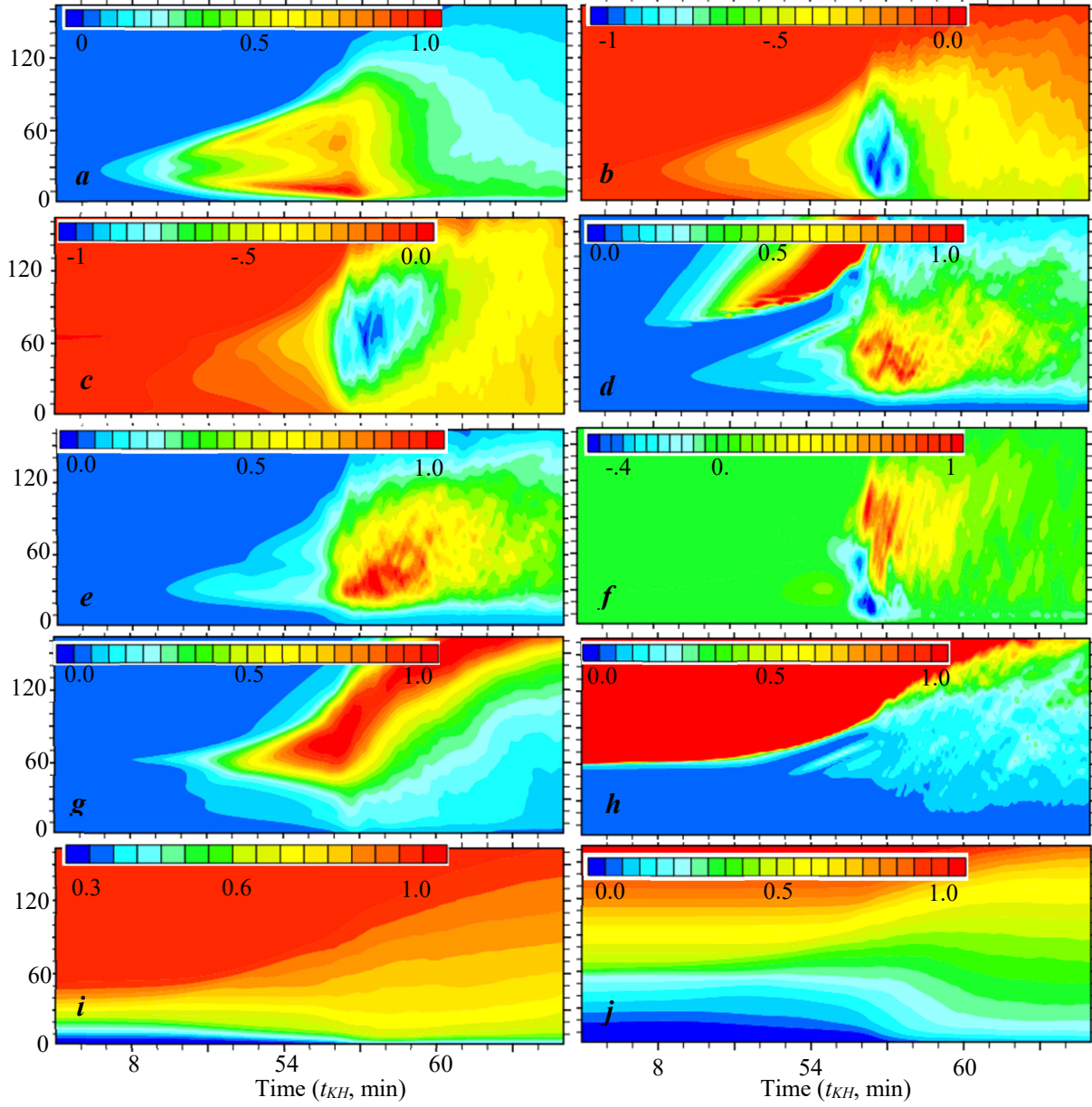


Figure 1.6. Time-height sections of a) kinetic energy (KE, normalized by  $5 \text{ m}^2/\text{s}^2$ ), b) momentum flux in the geostrophic wind direction ( $\overline{u'w'}$ , normalized by  $0.8 \text{ m}^2/\text{s}^2$ ), c) dynamic heat flux ( $\overline{w'\theta'}$  normalized by  $0.064 \text{ K m/s}$ ), d)  $K_m$  (normalized by  $20 \text{ m}^2/\text{s}$ ), e)  $K_h$  (normalized by  $20 \text{ m}^2/\text{s}$ ), f)  $\overline{w'^3}$  (normalized by  $0.6 \text{ m}^3/\text{s}^3$ ), g)  $\overline{\theta'^2}$  (normalized by  $0.044 \text{ K}^2$ ), h) Richardson number, i)  $U$  (normalized by  $U_g$ ), and j)  $\theta - \theta_0$  (K). Only the lowest 160 m is shown.

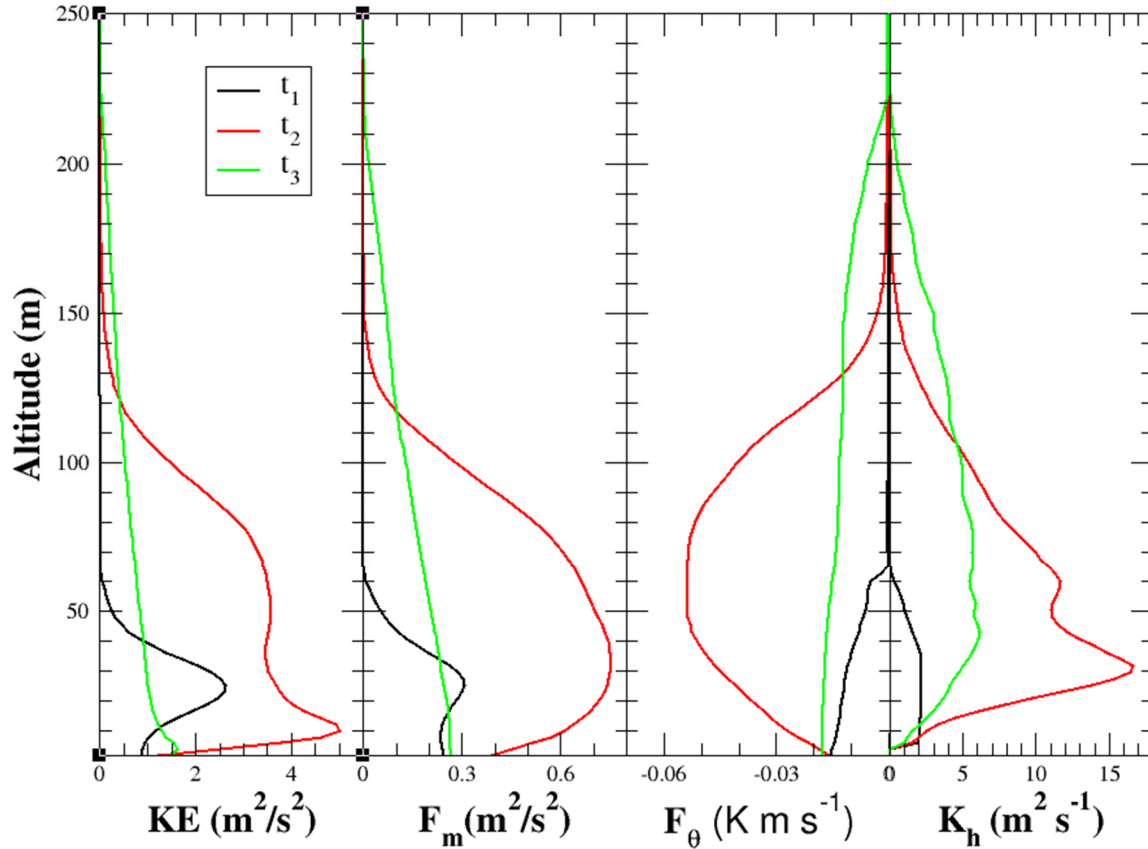


Figure 1.7. Profiles of the domain-averaged a) kinetic energy (KE,  $\text{m}^2/\text{s}^2$ ), b) momentum flux ( $F_m = \sqrt{u'w'^2 + v'w'^2}$ ,  $\text{m}^2/\text{s}^2$ ), c) dynamic heat flux ( $F_\theta = \overline{w'\theta'}$ ,  $\text{K m/s}$ ), and d) eddy diffusivity of potential temperature ( $K_h$ ,  $\text{m}^2/\text{s}$ ) at three different times,  $t_1=8$ ,  $t_2=11$ , and  $t_3=78$  min. Those at  $t_1=8$  min are multiplied by 10 for the sake of comparison.

#### b) Impact of KH billows on vertical mixing

The development of KHI and its impact on the mean boundary layer structure is further revealed in Fig. 1.6, which shows the time-height sections of several domain-averaged variables along with some second and third order moments. A kinetic energy (KE, including both KH induced perturbations and turbulent kinetic energy or TKE) maximum first appears approximately at  $z \sim 30$  m and  $t_{KH} \sim 6$  min (Fig. 1.6a). The KE maximum grows rapidly with time in terms of its magnitude and vertical extension, and then breaks into two separate maxima centered at approximately 10-15 m and 40-50 m, respectively, with a relative minimum at  $z = 30$  m. The KE distribution, namely two KE maxima separated by an inflection point level, is in qualitative agreement with the Doppler lidar observations of an KH event by Newsom and Banta (2003, their Fig. 1.6a). Further analysis indicates that these KE maxima are dominated by contributions from horizontal wind perturbations, which, according to the linear wave theory, are minimum near the inflection point level (e.g., Fig. 1.8a in Newsom and Banta 2003). Since the onset of the SSI ( $t_{KH} \sim 17$  min), both KE maxima start decreasing rapidly with time, and for  $t_{KH} > 20$  min, KE is characterized by a single maximum near the surface. In the meantime, the KE maximum envelope experiences a sharp increase in its vertical extension during the SSI and turbulent breakdown period and extends up to  $\sim 160$  m above the ground.

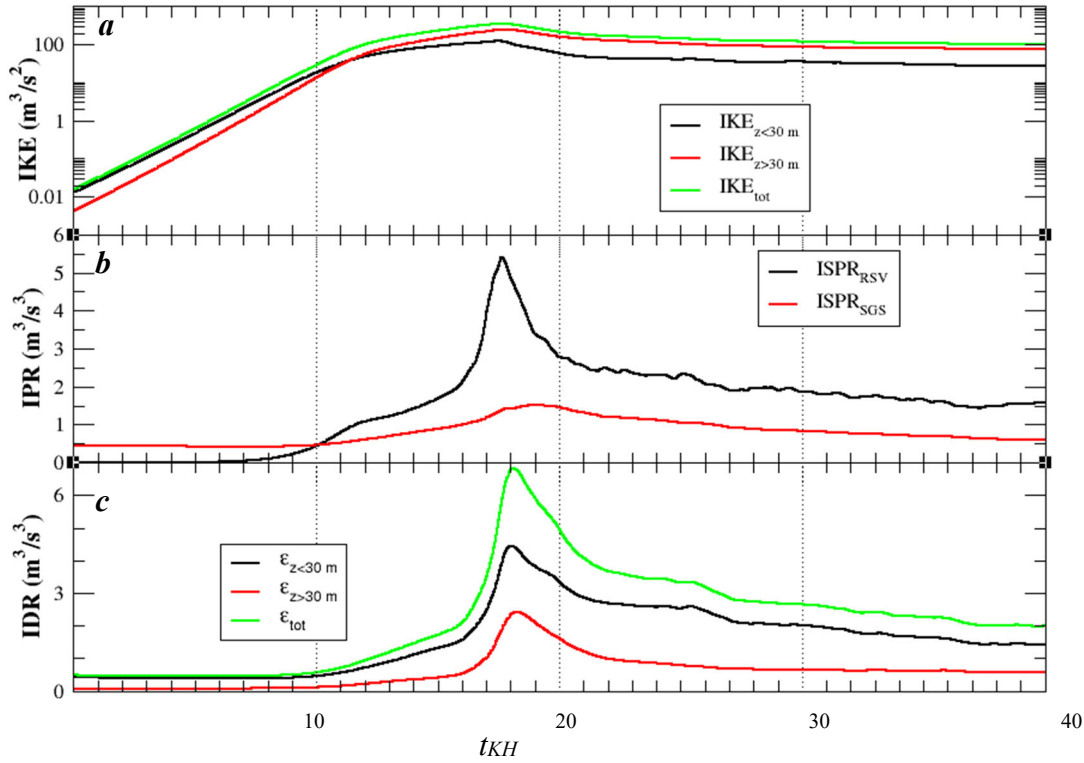


Figure 1.8. Evolution of the a) vertically integrated kinetic energy (IKE,  $\text{m}^3/\text{s}^2$ , top) in log-scale, b) vertically integrated shear production rate of KE by wind shear (ISPR,  $\text{m}^3/\text{s}^3$ , middle), and c) integrated dissipation rate (IDR,  $\epsilon$ ,  $\text{m}^3/\text{s}^3$ , bottom).

Shown in Fig. 1.6b is the  $u$ -wind flux,  $\overline{u'w'}$ , which is 1-2 orders of magnitude larger than its counterpart in the Y direction (i.e.,  $\overline{v'w'}$  not shown). The  $u$ -wind momentum flux is characterized with a negative maximum centered approximately at  $z \sim 25$  m level. The negative maximum grows in magnitude before the onset of SSI in pace with the KE growth, implying the role of KH waves in downward, and therefore down-gradient, momentum transport. The momentum flux peaks between  $t_{KH} = 16$ -18 min, coincided with the occurrence of the secondary shear instability, suggesting that the SSI and turbulent breakdown of the KH billows are more effective in vertical mixing of momentum than the KH billows themselves. The dynamic heat flux (i.e.,  $\overline{w'\theta'}$ , which is proportional to the sensible heat flux) is negative as well during this KH event, suggesting generally downward transport of heat by KH billows (Fig. 1.6c). The vertical extension and evolution of the heat flux resemble the momentum flux. However, the largest momentum flux occurs noticeably lower than its counterpart in the heat flux, presumably due to the fact that the strongest mean vertical wind shear is confined in the lowest 20 m while the vertical gradient of potential temperature is the largest near the SABL top.

The KHI-induced vertical mixing effect is also evident in the eddy diffusivity cross-sections (Fig. 1.6d-e). The eddy diffusivities of momentum ( $K_m$ ) and scalars ( $K_h$ ) are evaluated from the domain-averaged fluxes, vertical wind shear and potential temperature gradient using

$$K_m = \sqrt{\overline{u'w'^2} + \overline{v'w'^2}} / \sqrt{(\partial U / \partial z)^2 + (\partial V / \partial z)^2} \quad \text{and} \quad (1.2a)$$

$$K_h = -\overline{\theta'w'} / (\partial \bar{\theta} / \partial z). \quad (1.2b)$$

Both  $K_m$  and  $K_h$  exhibit similar vertical structure and evolution as their corresponding fluxes. A sharp increase in the eddy diffusivity occurs around  $t_{KH} \sim 16$  min, coincided with the onset of SSI. For both

momentum and scalars, the eddy diffusivity maxima exceed  $20 \text{ m}^2/\text{s}$ , suggesting much stronger mixing effect than in a typical stable atmospheric boundary layer (e.g., Costa et al. 2020, Figs. 1f and 1g). The strongest mixing lasts for only a few minutes, after which the maximum eddy diffusivity decreases to around  $10 \text{ m}^2/\text{s}$  and is nearly constant with time to the end of the simulation. The momentum eddy diffusivity also shows a detached maximum above the KH layer (Fig. 1.6d), where no strong turbulence is expected. Further diagnosis finds that this spurious eddy diffusivity maximum is due to a combination of relatively small momentum flux from KHI induced internal gravity waves and nearly zero vertical wind shear aloft.

The evolution of the boundary layer depth is evident in Fig. 1.6f-g, which show the potential temperature variances and gradient Richardson number. Large  $\theta$  variances are evident along the upper edge of the KE and flux maximum envelopes, associated with downward entrainment of the warmer air from above. Prior to the KH event, the SABL top ( $\sim 60 \text{ m}$ ) is defined by a sharp Richardson number gradient with  $Ri < Ric$  in the boundary layer, where  $Ric = 0.25$  denotes the critical Richardson number. Across the SSI and subsequent KH billow breakdown phase, the SABL top increases abruptly, and afterwards the sharp  $Ri$  gradient along the SABL top thickens and becomes blurry. The Richardson number in the upper portion of the SABL gradually increases presumably due to entrainment of the warmer air across the SABL top. It is worth noting that perturbations generated by the KH billow breakdown exhibit high skewness (Fig. 1.6h) with  $\overline{w'^3} < 0$  near the surface and  $\overline{w'^3} > 0$  above the KH billow cores, suggestive of downward transport of the KE near the surface and upward transport above the KHB cores.

Finally, the thickening of the SABL is also evident in the  $u$ -wind and  $\theta$  sections (Fig. 1.6i-j). For the  $u$ -wind, the whole shear layer expands substantially, from  $\sim 60 \text{ m}$  before the KH event to more than  $120 \text{ m}$  in about  $15 \text{ min}$ . Near the surface the winds are noticeably enhanced while the wind speed above the  $30 \text{ m}$  level weakens. Even in the presence of surface cooling, KHI enhanced vertical mixing leads to warming of the air below  $\sim 60 \text{ m}$  and cooling aloft. The evolutions of the mean  $u$ -wind and potential temperature over a homogeneous surface are governed by the following budget equations,

$$\frac{\partial \bar{u}}{\partial t} + f\bar{v} = -\frac{\partial \overline{w'u'}}{\partial z} \quad (1.3a)$$

$$\frac{\partial \bar{\theta}}{\partial t} = -\frac{\partial \overline{w'\theta'}}{\partial z} \quad (1.3b)$$

where the right hand side terms represent the vertical divergence of momentum and heat fluxes associated with both turbulence and KH waves, respectively. As shown in Fig. 1.6c, the negative heat flux reaches its maximum at  $z \sim 60 \text{ m}$ , and accordingly, the right hand side term in Eqn. (1.3b) is positive in the lowest  $\sim 60 \text{ m}$  and negative aloft. Therefore, we see warming near the surface and cooling in the upper portion of the SABL. In Eq. (1.3a), the Coriolis term is small, because the time scale for KH process is much shorter than the geostrophic adjustment time scale (i.e.,  $1/f$ ) and  $v$ -wind is much weaker than  $u$ -wind. If neglecting the Coriolis term, the  $u$ -wind evolution is solely driven by the vertical divergence of the momentum flux. The negative momentum flux maximum is located below  $30 \text{ m}$ , which is noticeably lower than the negative heat flux maximum. Accordingly, the  $u$ -wind component is enhanced near the surface where the right hand side of Eq. (1.3a) is positive and weakened aloft.

For a more quantitative comparison, the profiles of kinetic energy, momentum and heat fluxes and eddy diffusivity at times before, during and after the KH event are shown in Fig. 1.7. At the beginning of the KH event, KE is relatively small and so are the fluxes and eddy diffusivities. When secondary instability and turbulent breakdown of KH billows are taking place, KE becomes substantially larger with two separate maxima located at  $z = 10 \text{ m}$  and  $60 \text{ m}$ , respectively. The momentum and heat fluxes are dramatically increased by KH processes and exhibit a broad elevated maximum. Turbulence production in an elevated shear layer has been frequently observed in the stable atmospheric boundary layer and such a boundary

layer is often referred to as an “upside-down” boundary layer (e.g., Mahrt 1999). Accordingly, the eddy diffusivity becomes  $\sim 30$  times larger than at the earlier time when the KH instability just starts (Fig. 1.7d). The levels where the momentum and heat fluxes reach their maxima are  $\sim 30$  and  $60$  m, respectively. It is worth noting that the eddy diffusivities for the momentum and heat are quite similar (Fig. 1.6f-g), as the two are dependent mostly on the KH eddy and turbulent characteristics in the SABL.

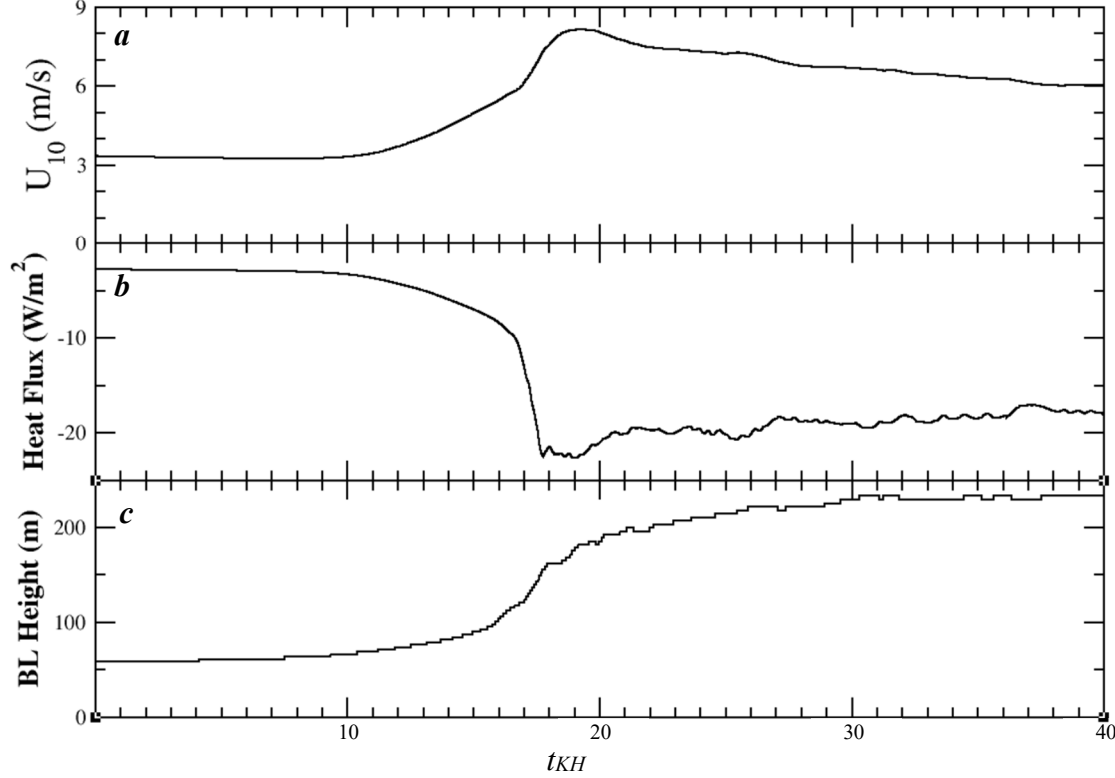


Figure 1.9. Evolution of the domain-averaged a) 10-m wind speed (m/s), b) surface sensible heat flux ( $\text{W/m}^2$ ), and c) SABL height during a 40-min period.

About twenty minutes later, the KE, fluxes and eddy diffusivities are much reduced from its peak values during the breakdown of KH billows and the elevated maxima in KE and flux profiles vanish, suggesting the weakening of vertical transport of momentum and heat by KH induced perturbations and turbulence. The momentum flux linearly decreases with height between the surface and the SABL top, which is now located at  $\sim 220$  m, suggesting that the SABL thickens more than three times over a 30 min period.

Shown in Fig. 1.8 are the vertically integrated KE, shear production rate and dissipation rate over a 40-min time period. In general, the integrated KE grows exponentially over the first 18 minutes by approximately four orders of magnitude and monotonically decays afterwards. Quantitatively, KE grows faster over the first 12 min, presumably corresponding to the growth of primary KH waves and billows. The KE growth slows down over the next  $\sim 6$  minutes associated with mature billows and onset of the secondary shear instability. If we divide the total KE into the lower portion below the inflection point level (i.e.,  $z_{IP} = 30$  m) and the upper portion above, the former grows noticeably slower than latter. For comparison, the e-folding growth times,  $T_g$ , can be estimated for the first 12 minutes using  $KE_2 \sim KE_1 \exp(\Delta t / T_g)$ , where  $KE_1$  and  $KE_2$  denote the integrated KE at two reference times 1, and 2 over time period  $\Delta t$ . The derived e-folding growth times for the lower, upper, and total KE are 104.4, 69.3, and 73.6 sec, respectively. The slower growth of KE near the surface is likely due to the influence from the ground surface and surface stress. This is in qualitative agreement with Liu et al. (2023), who demonstrated that the KHB grows slower near a solid surface.

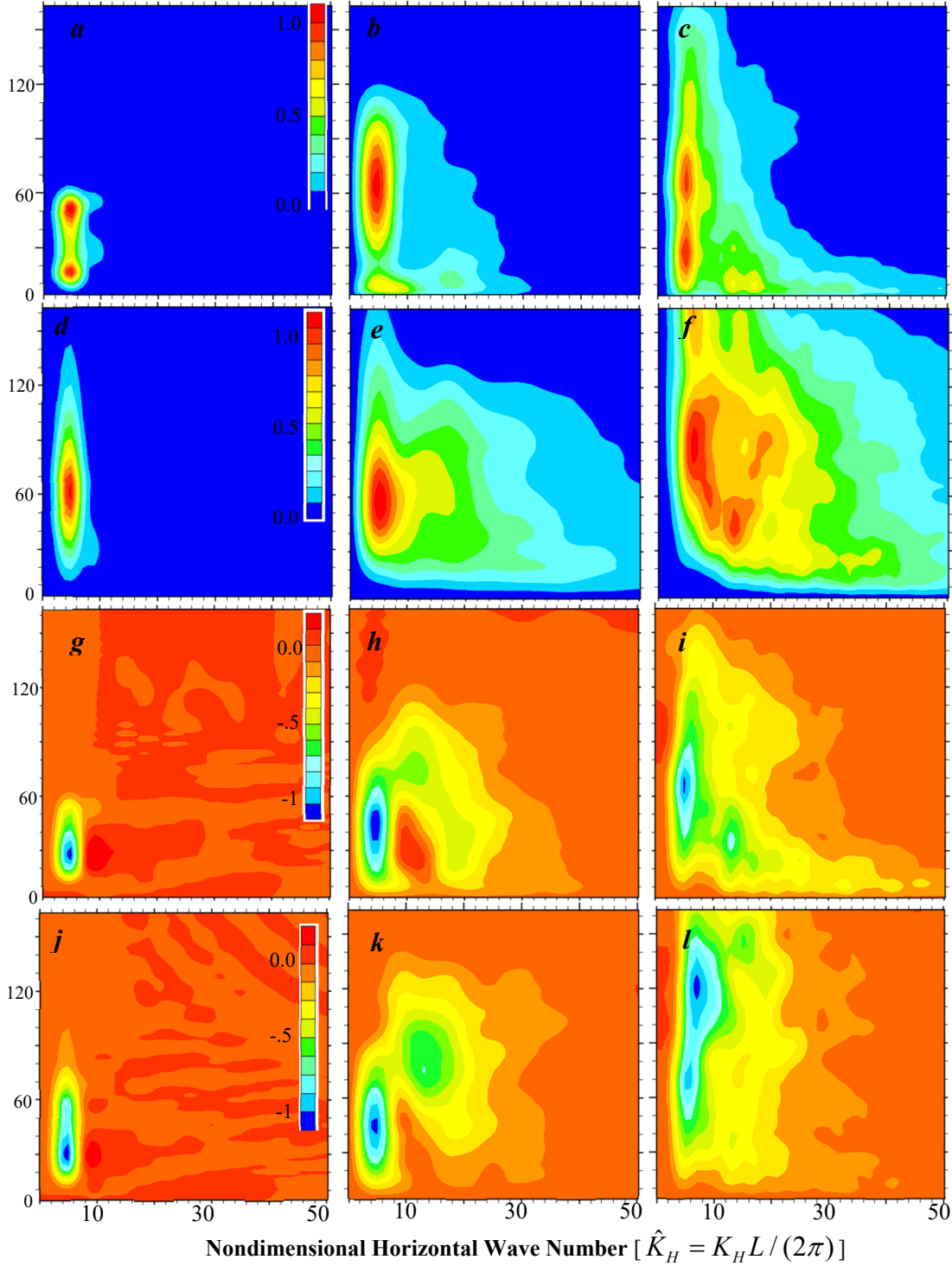


Figure 1.10. Dimensionless power spectra of a)  $u$  and b)  $w$ , and dimensionless co-spectra of c)  $u-w$  and d)  $w-\theta'$  valid at  $t_{KH}=13$  (left), 18 (middle), and 23 min (right), respectively, are shown in nondimensional wave number ( $\hat{K}_H$ ) and height sections. Only the lowest 160 m and  $\hat{K}_H$  up to 50 is shown.

For the mature billow and SSI phase, the KE in the lowest 30 m grows slower than its counterpart aloft as well. Similarly, the KE decay portion can be divided into two phases, the breakdown phase (between  $t_{KH} = 18-21$  min) during which KE decreases exponentially, predominantly due to turbulent breakdown of the

primary and secondary KH billows, and the dissipation phase ( $t_{KH} > 21$  min), which features slower decrease of KE dominated by turbulent dissipation. The shear production rate from resolved eddies (i.e.,  $-U \frac{\partial \overline{u'w'}}{\partial z} - V \frac{\partial \overline{v'w'}}{\partial z}$ ) shows a pronounced maximum at  $t_{KH} = 18$  min, coincided with the KE maximum

(Fig. 1.8b). The shear production rate related to the parameterized turbulence (i.e.,  $-U \frac{\partial \overline{\tau_{xz}}}{\partial z} - V \frac{\partial \overline{\tau_{yz}}}{\partial z}$ , where  $\overline{\tau_{xz}}$  and  $\overline{\tau_{yz}}$  denote SGS shear stress) increases with the growth of KH billows as well, suggesting that KH billows tend to enhance small-scale turbulence. The vertically-integrated dissipation rate exhibits a less pronounced maximum that lags behind the SPR peak by approximately 1.5 min. In addition, stronger dissipation takes place in the lowest 30 m than in the layer above (Fig. 1.8c).

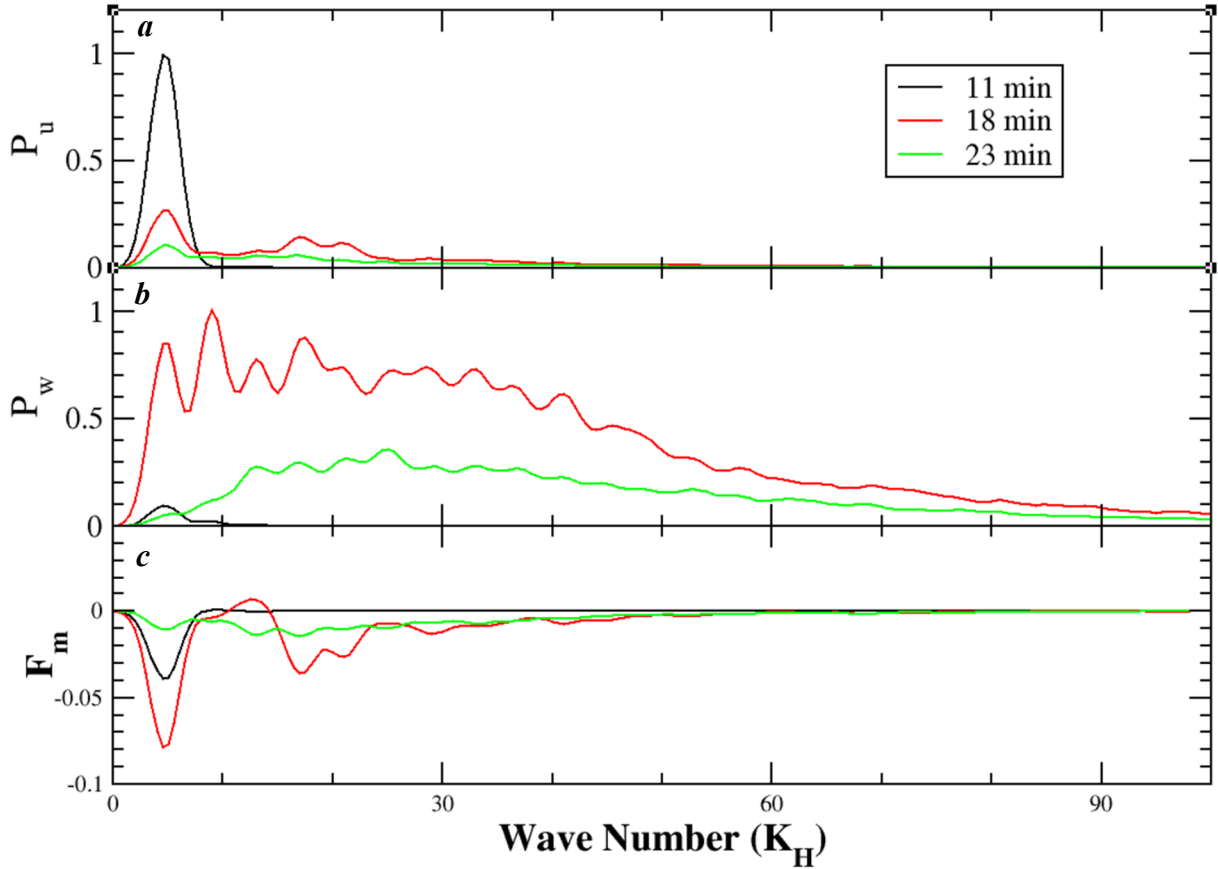


Figure 1.11. Dimensionless power spectra of  $u$  and  $w$  and  $u$ - $w$  co-spectrum at  $z = 21$  m valid at  $t_{KH} = 11, 18$ , and  $23$  min, respectively.

The evolution of the domain-averaged 10-m wind, surface heat flux, and SABL depth over the same time period are shown in Fig. 1.9. The 10-m wind increases with time during the KE growth phase and reaches its maximum approximately at  $t_{KH} = 19$  min, apparently associated with the downward transport of momentum by KH billows. Afterwards, the 10-m wind speed gradually decreases with time. In accordance to the sudden increase in the 10-m wind, the surface stress increases significantly, approximately in proportion to the square of the 10-m wind speed (not shown). The magnitude of the surface heat flux increases by several folds during this KH event as well. It is worth noting that the increase in 10-m wind during the KH event also implies substantial stronger wind shear near the surface. The enhanced shear layer or surface vortex sheet is lifted off the surface (Fig. 1.1c) and wrapped into KH braids during

the mature phase of the KH billows and therefore, plays a role in the development of secondary shear instability. Finally, the SABL depth increases from  $\sim 60$  m prior to the KH event to more than 200 m in less than 30 min, suggesting the dramatic impact of the KH event on the mean stable boundary layer characteristics.

c) *Power spectra and co-spectra*

The above analysis suggests that KHI may significantly enhance vertical transport of momentum and scalars, substantially deepen the SABL, generate residual waves and intermittent turbulence, modulate mean boundary layer structure and enhance interaction with the surface. Questions remain regarding contributions from KH waves, secondary instability, and smaller scale turbulence induced or enhanced by the KH waves. To address these questions, we have computed the power spectra of the three wind components and a pair of co-spectra using the volume data. Shown in Fig. 1.10 are the power spectra of  $u$  and  $w$  and co-spectra of  $w-u$  (i.e., momentum flux in the wave number space) and  $w-\theta'$  (i.e., dynamic heat flux) as a function of the dimensionless horizontal wave number,  $\hat{K}_H = L_x \sqrt{k^2 + l^2} / (2\pi)$ , for the lowest 160 m. Here  $k$  and  $l$  denote the wave numbers in the X and Y directions, respectively, and only the portion for  $\hat{K}_H \leq 50$  is shown in Fig. 1.10. At the early stage of the KHI event, the  $u$  variance is largely confined in the lowest 60 m, and characterized by two maxima separated by the inflection point level (i.e.,  $z_{IP} \sim 30$  m) centered at the nondimensional horizontal wave number  $K_H \sim 4.1$ . This corresponds to a wavelength of  $\lambda = L_x / \hat{K}_H \sim 374$  m, which is comparable to the wavelength estimated from the plan-views in Fig. 2. The  $v$ -wind spectrum (not shown) resembles its  $u$ -wind counterpart except that it is much smaller in magnitude. The  $w$  power spectrum shows a single maximum centered at  $z = 60$  m and  $\hat{K}_H = 4.1$ , which extends up to  $\sim 140$  m. The maximum of the  $w$  spectrum is only around 2% of its  $u$ -wind counterpart. Therefore, the largest contribution to KE comes from  $u$ -wind perturbations associated with the primary KH waves. It is worth mentioning that, in an LES study of hurricane boundary layer rolls by Wang and Jiang (2017), they found that, while the shear instability is attributed to an inflection point in the radial wind profile, the strength of the rolls are positively correlated with the shear in the much stronger tangential wind. The same seems to be true for this study. The  $u-w$  co-spectrum is characterized by a single maximum at 30 m, suggesting that the KH waves are able to induce substantial vertical transport of momentum fluxes. The same is true for the  $w-\theta'$  co-spectrum, except that the latter extends up to  $\sim 100$  m, noticeably deeper than the maximum in the  $w-u$  co-spectrum.

During the SSI and breakdown phases (mid-column), a pair of pronounced maxima centered at  $\hat{K}_H = 4.1$  is still evident in the  $u$ -wind spectra, implying that the primary KH waves (or billows) induced perturbations remain significant. However, the peak value in Fig. 1.10b is only about a third of that in Fig. 1.10a (i.e., 5 min ago), implying that the primary waves are weakened by the SSI substantially and nearly two-third of its energy is cascaded into the secondary waves and large eddies. Accordingly, the  $u$  and  $w$  spectra spread toward higher wave numbers, especially near the surface (i.e.,  $z < 30$  m), where a secondary maximum in the  $u$  spectra is located at  $\hat{K}_H = 18$  corresponding to a wave length of  $\sim 85$  m. The  $w$  spectra show more substantial spreading of energy to the higher wave numbers (Fig. 1.10e). The  $u-w$  co-spectra at  $t_{KH} = 18$  min show three discernible maxima, namely the primary maximum, now centered at  $z = 45$  m, a minor maximum centered at  $\hat{K}_H \sim 12$  and  $z \sim 70$  m, and a third maximum at  $\hat{K}_H \sim 18$ , suggesting that while the primary KH waves still play a substantial role in vertical momentum transport, perturbations associated with SSI billows become equally important. For the  $w-\theta'$  co-spectrum, in addition to the primary maximum located at  $\hat{K}_H = 4.1$  and  $z = 45$  m, there is a marked maximum centered at  $\hat{K}_H = 12$  and  $z = 80$  m (Fig. 1.10k), presumably associated with downward entrainment of warmer air from above by the SSI billows.

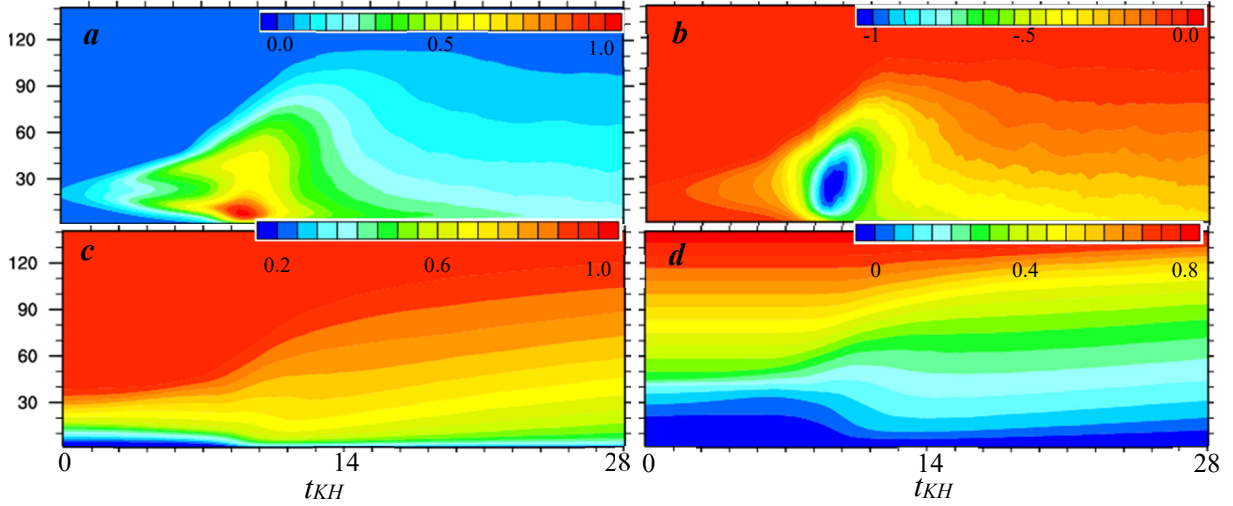


Figure 1.12. Time-height sections of a) KE (normalized by  $1.7 \text{ m}^2/\text{s}^2$ ), b) momentum flux (normalized by  $0.32 \text{ m}^2/\text{s}^2$ ), c)  $U$  (normalized by  $U_g = 9 \text{ m/s}$ ), and d) potential temperature ( $\theta - \theta_0$ , incr. =  $0.05 \text{ K}$ ).

At  $t_{KH} = 23 \text{ min}$  (right column), both  $u$  and  $w$  power spectra extend further up beyond  $160 \text{ m}$  while spreading more towards higher wave numbers. While at this time, KH billow cores are barely distinguishable from vertical cross-sections, there is still a pronounced maximum centered at the primary KH wave number, implying that the residual waves persist after the turbulent breakdown occurs. For the  $u$  spectra, a few secondary maxima appear in the lowest  $30 \text{ m}$ , suggesting a downscale energy cascade from the KH billows to the turbulence in the SABL. The residue waves still play an important role in the vertical transport of the momentum and sensible heat along with substantial contributions from higher wave numbers (Fig. 1.10i and l).

The evolution of the energetics and momentum mixing in the boundary layer (i.e., at  $z = 21 \text{ m}$ ) is further revealed by Fig. 1.11. From  $t_{KH} = 16$  to  $23 \text{ min}$ , a significant portion of the kinetic energy cascades from the primary KH waves to SSI waves and large eddies. In addition, more kinetic energy is transferred from horizontal to vertical motion by the SSI process. From  $t_{KH} = 23$  to  $28 \text{ min}$ , kinetic energy weakens at all wave numbers, presumably due to turbulent dissipation. In terms of momentum flux, the primary wave contribution to the downward flux doubles from  $t_{KH} = 16$  to  $23 \text{ min}$  with comparable contributions from secondary waves with  $\hat{K}_H$  between  $13$  and  $23$ .

#### 1.2.4 Discussion

Development of the secondary instability and the subsequent turbulent breakdown of KH billows have been the subject of numerous two-dimensional and three-dimensional numerical studies (see Thorpe 2012 for a brief review). Some proposed secondary instability may only appear in two-dimensional simulations or relatively low Reynolds number fluids. It is evident that secondary shear instability occurs and plays the dominant role in turbulent breakdown of the KH billows in the control simulation. However, the small model domain used in the control simulation may have suppressed some other types of secondary instability, such as vortex pairing instability and formation of knots and Tubes (Fritts 2021). To address this issue, sensitivity simulations have been conducted with varying geostrophic winds, domain sizes and grid spacings. This section provides a brief summary of these sensitivity simulations using simulation B as an example with emphasis on the secondary instability and domains size dependence of the simulated KH waves.

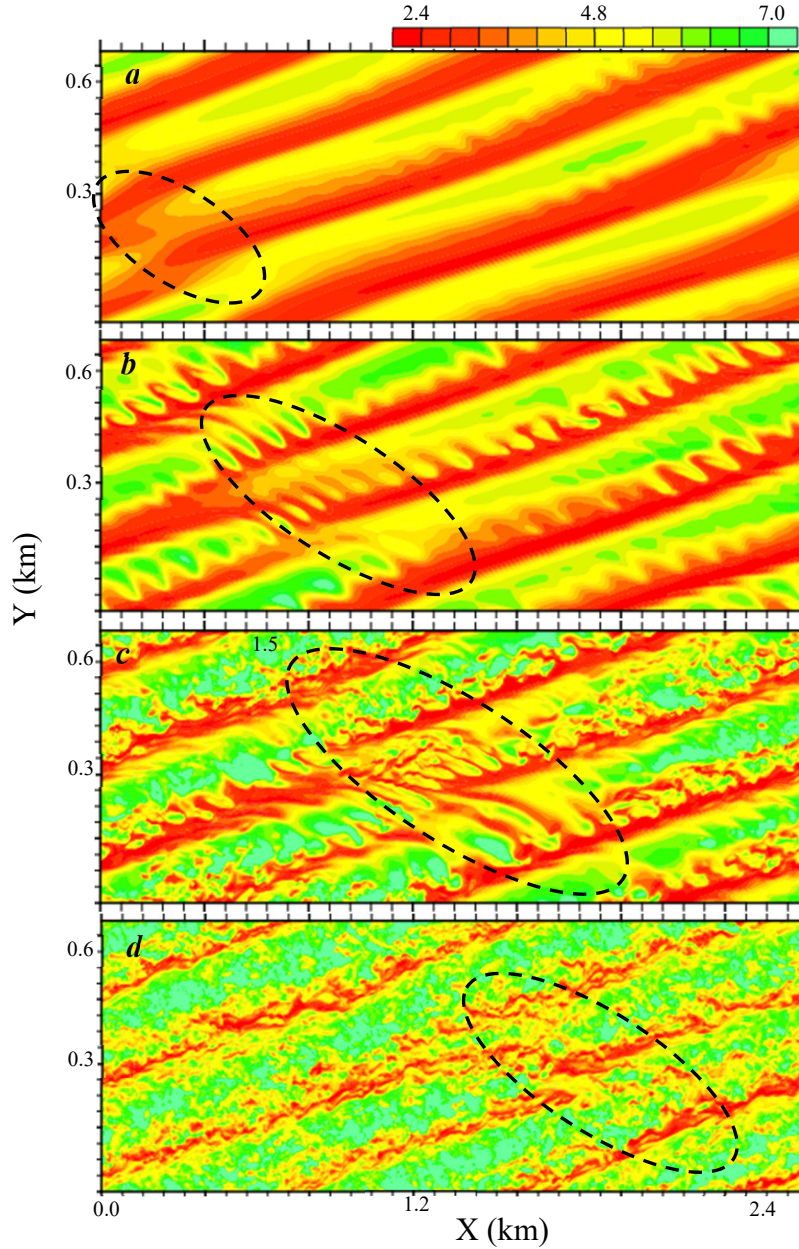


Figure 1.13. The X-Y sections of  $u$  at  $z = 10.6$  m from simulation B valid at  $t_{KH} = 5.5, 6.9, 8.3,$  and  $9.7$  min, respectively. The color increment =  $0.4$  m/s. The misalignment portions (or knots) are highlighted by dashed ovals.

The model configuration for simulations B is identical to the control simulation except that the initial geostrophic wind is  $U_g = 9$  m/s, the horizontal domain is larger (i.e.,  $2560$  m  $\times$   $640$  m) and the horizontal grid spacing is finer (i.e.,  $\Delta x = \Delta y = 2.5$  m). KHI takes place in simulation B and the resulted KE maximum resembles its counterpart in the control except that its peak value is smaller, the KE maximum is centered at  $\sim 22$  m, which is closer to the surface (Fig. 1.12), and the vertical extension of the KH maximum layer is lower. It is interesting that the ratio of KE maxima between the two simulations is around 2.8, which is approximately equal to the geostrophic wind squared,  $(15/9)^2 \sim 2.8$ , suggesting that KE tends to be proportional to the vertical shear squared in the  $u$  component, although the inflection point appears in the  $v$  profile in this pair of simulations.

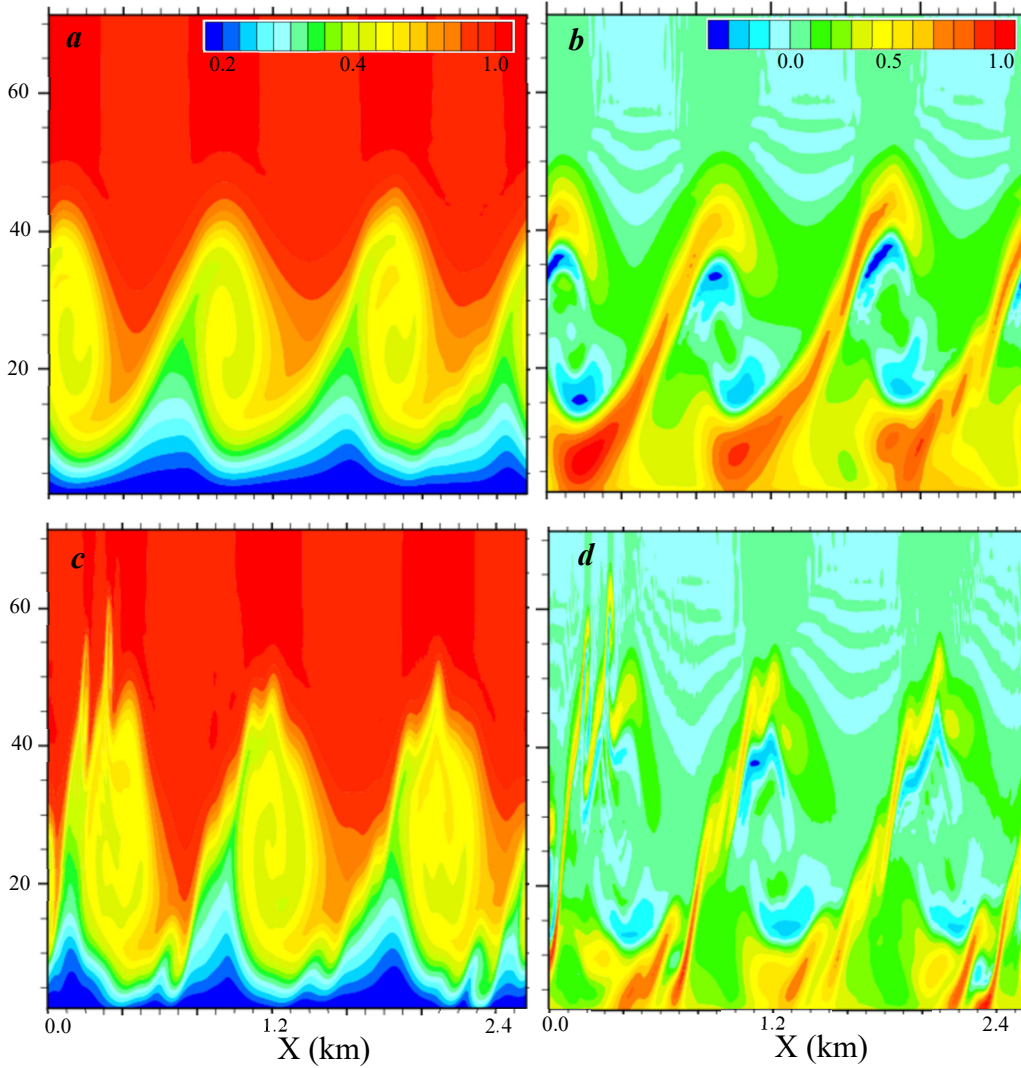


Figure 1.14. Vertical cross-sections of a, c)  $u$  (m/s, normalized by  $U_g = 9$  m/s, incr. = 0.05) and b, d) normalized span-wise vorticity (normalized by  $0.6 \text{ s}^{-1}$  and  $1.01 \text{ s}^{-1}$ , color incr. = 0.1), valid at  $t_{KH} = 6.9$  (a and b) and 8.3 (c, d) min, respectively.

There are 3-6 waves in the horizontal domain (Fig. 1.13) with a shorter wavelength (i.e.,  $\lambda \sim 247$  m). However, the angle between wave phase lines and the geostrophic wind direction is about the same (see Table 1.1) and the ratios between the wavelength and the boundary layer depth are comparable as well. Just as in the control, three-dimensional secondary shear instability occurs first along the braids, when the primary KH billows are mature enough (Fig. 1.13 and 14). At their mature phase, the KH cores are located between 10-40 m, and the ratio between the core height and horizontal wave length is around 0.12, which is substantially smaller than those reported in previous studies with KH billows away from any solid surface. The same happens in other simulations, including a weaker geostrophic wind ( $U_g = 6$  m/s) simulation and a larger domain simulation (i.e., identical to B but with  $dx = 5$  m and therefore both domain length and width are doubled). No vortex pairing is observed in any sensitivity simulations, likely due to the close proximity of KH billows to the ground surface, which tends to suppress the subharmonic instability (Liu et al. 2023).

The phase lines in Fig. 1.13 are not as perfectly straight as in the control simulation. KH billows with variable phases that are misaligned along their axes have been documented in laboratory experiments (Thorpe 1985, 1987) and numerical simulations (Fritts et al. 2022). The misaligned portions, also known as “tubes” and “knots”, may evolve more rapidly than a typical secondary instability and accelerate turbulent breakdown of KH billows (e.g., Fritts et al. 2022). It seems to be the case in Fig 13 (highlighted by ovals). It is worth noting that the misalignment of KH billows in simulation B is due to the double-periodic lateral boundary conditions. In nature, this could happen associated with spatial variability in the large-scale flow.

### 1.2.5 Summary and Concluding Remarks

We have examined KH instability in an Ekman type stable atmospheric boundary layer using large eddy simulations. The LES model is initialized using a geostrophic wind which is constant with height without a prescribed shear layer. The vertical wind shear develops naturally as the SABL spins-up. While the wind component in the geostrophic wind direction (i.e.,  $u$ ) is substantially stronger than the  $v$  wind, KHI takes place associated with an inflection point in the weaker  $v$  wind profile.

The simulated KH waves are initially two-dimensional with their horizontal phase lines tilting with an acute angle from the  $u$  wind direction. As demonstrated in the control simulation, the KH event can be divided into four phases, namely the growth phase (i.e., exponential growth of KE in the primary KH waves,  $\sim 12$  min), mature KH billow phase (slow growth of KE associated with mature KH billows and secondary instability,  $\sim 6$  min), turbulent breakdown phase (fast exponential decay of KE due to turbulent breakdown of KH billow cores,  $\sim 4$  min) and recovery phase associated with slow decay of KE primarily due to turbulent dissipation.

The general characteristics of the simulated KH waves (or billows) and their subsequent turbulent breakdown are in qualitative agreement with previous analytical, numerical and observational studies. For example, KE grows exponentially with time as predicted by linear theory, and the wavelength to shear layer depth ratio is between 6 and 10, which is consistent with previous studies. The simulated KH wave length and the presence of two stacked KE maxima are in qualitative agreement with the Doppler lidar analysis of a KH event by documented during CASES-99 (Newsom and Banta 2003). In the meantime, several unique aspects and new insights from this study are worth noting. First, the simulated KH billows are centered at 20-30 m above the ground surface, and the resulting  $z_{IP}/D$ , where  $D$  is the half depth of the shear layer, is around unity or less. Regardless of the close proximity to a solid frictional surface, the KH billows and the general characteristics of the billow cores are consistent with previous numerical studies, and the impact of the ground surface on the KHI in the SABL is more quantitative. For example, it has been demonstrated that the integrated KE near the surface grows slower than its counterpart aloft, suggesting that a frictional surface tends to slow down the growth of KH billows, which is consistent with the recent DNS study by Liu et al. (2023). In addition, the ratio between the height of the mature KH billow cores and the horizontal wavelength appears to be smaller, suggesting that the ground surface plays a role in limiting the height of KH billow cores.

This study also suggests that the turbulent breakdown of KH billows in the SABL starts with the onset of secondary shear instability, which first develops along the S-shaped braids. In addition to the SSI, “tubes” and “knots” associated with misalignment of phase lines appear in some simulations, which seems to accelerate the secondary instability. Other types of secondary instability identified in previous studies are not observed in our simulations.

As shown in previous studies, KH billows and their subsequent breakdown may dramatically enhance vertical transport (and mixing) or momentum and heat. The magnitudes of KHI induced vertical fluxes peak during the onset of SSI and breakdown of the KH billows, and then slowly decay with time. This leads to significant increase in the SABL depth, weakened wind shear and stratification throughout most of the SABL, and increased Richardson number in the upper portion of the SABL. It is worth noting that KHI-

induced the momentum and heat fluxes exhibit substantial spatiotemporal variability. For example, the maximum momentum flux doubles itself from  $t_{KH}=15$  to 17 min (Fig. 1.6b), and for  $t_{KH}=16$  min, the momentum flux at  $z=40$  m is twice as large as at  $z=20$  m. This large variability suggests that one has to be careful when evaluating the KHI impact on boundary layer mixing from limited observations.

## 2. INTERACTION BETWEEN THE ARCTIC SABL AND SNOW OR ICE SURFACE

### 2.1 Background and Motivation

Over the Arctic area, the atmosphere is stably stratified due to the low temperature at the surface. The atmospheric stable boundary layer (ASBL) appears at night and during the daytime year around (Thorsten, 2007; Steeneneld, 2014). It is difficult to have robust parameterizations to represent the development and maintenance of the ASBL due to multiple physical processes involved and complex interactions among them. One of these essential and significant processes is the coupling between the lower atmosphere and the land surface with vegetation and soil. A realistic physical representation of the interdependence of the atmosphere and the underlying soil and vegetation can better model the surface temperature and sensible heat flux in the stable boundary layer. It is essential to have proper feedbacks between the surface temperature, the surface sensible heat flux and the heat flux from the underlying soil and vegetation towards the surface (Steenefeld et al., 2006).

Strong cooling is found in some numerical weather prediction (NWP) models for stable boundary conditions due to unrealistic decoupling of the atmosphere from the surface, further stabilizing the boundary layer and leading to a shallower boundary layer (Walsh et al., 2008). It is described as a positive feedback between the stratification and the sensible heat flux and results in a collapse of turbulence and a decoupling of the actual ASBL from the surface (Steenefeld et al., 2006). In this case, Mahrt (1999) suggested that similarity theory and the traditional concept of a boundary-layer break down under very stable conditions. The strongest turbulence may be detached from the surface and generated by shear associated with a low-level jet, gravity waves, or meandering motion. Delage et al. (2002) indicate that strong decoupling occurs during a weak wind when turbulent heat flux cannot oppose the net longwave radiation. Turbulence dies out from the ground upwards, permits dramatic falls in surface temperature, and results in a positive feedback. Decoupling between the first model level and the surface is due to a missing process or some other model deficiency and is initially attributed to the turbulent transfer formulation. Such decoupling arises from the physics processes rather than the finite-difference schemes and appears to occur in the real atmosphere (Derbyshire, 1999).

No theory is available to predict this phenomenon since it needs to be better understood and categorically satisfy traditional definitions of turbulence (Van de Wiel et al., 2004; Mahrt, 2014). Some ASBL schemes derived from micrometeorological research seem to allow a ‘decoupling’ behavior. Such decoupling of the surface from atmospheric fluxes can permit dramatic and possibly unrealistic falls in surface temperature. An inadequate formulation of turbulent transfer can result in the occurrence of decoupling in the model (Delage et al., 2002). Using downward longwave rather than net longwave radiation to drive the land surface process can benefit the model and prevent it from decoupling. This modification allows the model to produce a natural negative feedback between the surface air temperature and the turbulent heat flux and reduce the cold bias. A formulation giving greater downward flux values in situations of strong stability is better at preventing the occurrence of complete decoupling. More frequently observed in nature is the negative feedback when the wind speed stays above  $5 \text{ ms}^{-1}$  with the intermittency of turbulence.

For snow-covered surfaces, as in the Arctic land area, correct thermal insulation of the snowpack is essential to improve the realism of the near-surface atmospheric temperature (Dutra et al., 2012, Steeneveld, 2014). Snow melt and sublimation are important processes for manipulating the ground surface temperature and heat flux. Barlage et al. (2010) have modified the Noah land surface model (LSM) to improve snowpack prediction in the Colorado Rocky Mountains. Their modification includes time-varying snow albedo and increasing fresh snow albedo, adjusting solar radiation for terrain slope and orientation, reducing the surface exchange coefficient for the stable boundary layer, and adjusting surface roughness length when snow is present. It has diminished excessive sublimation and early snowmelt and improved the magnitude and timing of seasonal maximum snow water equivalent (SWE).

To improve the stable surface layer in the NCEP Global Forecast System over the deep snowpack and snow-free conditions, Zheng et al. (2017) have implemented a stability parameter constraint for preventing the land-atmosphere system from fully decoupling with the modification of the roughness length formulation. It reduces the excessive near-surface cooling since the new modification prevents the collapse of turbulence in the stable surface layer over land and decoupling between the atmosphere and the surface. Consequently, it improves the forecast skill for light and medium precipitation amounts.

Wang et al. (2010) have found that the shading of the forest canopy is not considered for physical processes associated with deep snow under dense boreal forests in the Noah land surface model (LSM). It causes most solar radiation reaches the ground under the canopy and increases surface energy available for snow sublimation and melting. Furthermore, the downward sensible heat flux under very stable conditions is also overestimated, contributing to more snow sublimation and less snow depth and SWE. Snow physics under the canopy is added to Noah LSM to solve these issues to improve snow depth and SWE prediction. The new snow physics is verified with the observations from the high-altitude Niw Ridge forest site (40.03 °N, 105.55 °W) and a boreal forest site (53.9 °N, 104.7 °W). The new revisions improve simulations for all snow processes, including SWE, snow depth, and sensible and latent heat fluxes. As expected, the canopy shading effect on the underlying snow is most important for improving overall snow simulation. Erlandsen et al. (2016) have also included this snow physics in the Advanced Research version of the Weather Research and Forecasting (WRF) Model (ARW) for studying the sensitivity of the terrestrial surface energy and water balance estimates for the south Norway snow area. It has shown decreased net downward surface radiation flux and sensible heat flux due to increasing ground snow by altering the snow/rain criterion.

The Naval Research Laboratory (NRL) Coupled Ocean/Atmosphere Prediction System (COAMPS, Hodur, 1997; Hodur, 2002) has implemented the Noah LSM for the land surface processes. Operational forecasts have shown the cold bias over the Arctic land area with the older Noah LSM version (Noah3.2). The new revisions for the snow processes and physics (Wang et al., 2010) in the Noah LSM (Noah3.6) are added in COAMPS to improve Arctic forecasts. The initial conditions for COAMPS LSM are derived from the National Aeronautics and Space Administration (NASA) Land Information System (LIS, Kumar et al. 2006) analysis. The LIS is a high-resolution land data assimilation system integrating advanced LSM and high-resolution satellite and observational data (Kumar et al., 2008). It uses the Noah LSM for the analyses, and so does the COAMPS forecast. Therefore the two systems provide consistency in the LSM.

## **2.2 Description and applications of the new snow physics**

This section briefly describes the main changes related to the new snow physics, and more details can be found in Wang et al. (2010). The control version of LSM (Noah3.2) in COAMPS is without the new snow physics

### *2.2.1 Adding vegetation shading effect on snow surface*

The new snow physics aims to reduce the overestimation in potential evapotranspiration so that snow sublimation and snow melt are not excessive. For this purpose, the new snow physics includes the fraction of vegetation with snow below ( $F_{vb}$ ).

$$F_{vb} = GVF \bullet F_{g,sn} (1 - F_{bur}) \quad (2.1a)$$

$$F_{g,sn} = \frac{sneqv}{sneqv_{cr,g}} \quad (2.1b)$$

$$F_{bur} = \frac{H_{sn} - Z_{bot}}{Z_{top} - Z_{bot}} \quad (2.1c)$$

where  $GVF$  is the green vegetation fraction,  $F_{g,sn}$  is the fraction of ground covered with snow and is between 0 and 1,  $sneqv$  is the SWE,  $sneqv_{cr,g} = 0.02$  m is the critical SWE for fully snow-covered ground,  $F_{bur}$  is the snow burial fraction and is between 0 and 1,  $H_{sn}$  is the snow depth, and  $Z_{top}$  and  $Z_{bot}$  are the canopy top and bottom heights, respectively, prescribed as a function of the vegetation type.

Since the vegetation shading effect for the under-canopy snow is important for the solar radiation during the day, the net shortwave radiation ( $SW_{net}$ ) is changed to be a new weighted average ( $SW_{netn}$ ) between the vegetation shaded fraction  $F_{vb}$  and the non-shaded fraction  $1 - F_{vb}$ :

$$SW_{netn} = \{ (1 - \alpha_v)(1 - \alpha_{usn}) \bullet SW \downarrow \bullet \gamma \} \bullet F_{vb} + (1 - \alpha) SW \downarrow \bullet (1 - F_{vb}) \quad (2.2a)$$

where  $SW \downarrow$  is the downward solar radiation,  $\alpha$  is the grid cell average albedo with vegetation albedo  $\alpha_v \approx 0.13$  and under-canopy snow albedo  $\alpha_{usn} \approx 0.5$ , and  $\gamma = \exp(-LAI)$  with  $LAI$  being the leaf area index. The  $LAI$  is related to the  $GVF$  as:

$$LAI = (1 - GVF) * LAI_{min}(vegt) + GVF * LAI_{max}(vegt) \quad (2.2b)$$

where  $vegt$  is vegetation type,  $LAI_{min}(vegt)$  and  $LAI_{max}(vegt)$  are from a table of vegetation parameters for Noah3.6. The  $LAI$  is adjusted for vertical burying by snow in the Noah3.6:

$$LAI = LAI * (1 - F_{bur}) \quad (2.2c)$$

The  $SW_{netn}$  is combined with the net longwave radiation  $LW_{net}$  for the new net radiative flux received by the surface  $R_{netn} = SW_{netn} + LW_{net}$ , which is used to compute potential evapotranspiration ( $E_p$ ) using the Penman equation (Penman, 1948; Mahrt and Ek, 1984).

$$E_p = \frac{\Delta}{1 + \Delta} (R_{netn} - G) + \frac{\rho L_v}{r_a (1 + \Delta)} (q^* - q) \quad (2.3a)$$

$$\Delta = \frac{0.622 L_v}{p} \frac{de^*(T)}{c_p dT} \quad (2.3b)$$

where  $G$  is the ground soil heat flux,  $\rho$  is the surface air density,  $L_v$  is the latent heat of condensation,  $c_p$  is the specific heat of air,  $r_a$  is the aerodynamic resistance [ $r_a = 1 / (C_h u)$ ] with  $C_h$  being the turbulent

exchange coefficient and  $u$  the wind speed,  $q^*$  is the saturated atmospheric specific humidity,  $e^*$  is the saturated vapor pressure,  $p$  is surface pressure, and  $T$  is near-surface air temperature.

The new  $E_p$  is used to determine the snow sublimation rate ( $E_{sn}$ ):

$$E_{sn} = E_p \bullet F_{sn} \quad (2.4a)$$

$$F_{sn} = \alpha_s \frac{sneqv}{sneqv_{\max}} \exp\left[-\alpha_s (sneqv / sneqv_{\max})\right] + \frac{1}{sneqv_{\max}} \exp(-\alpha_s) \quad (2.4b)$$

where  $F_{sn}$  is the fractional snow coverage of the model grid cell ( $0 \leq F_{sn} \leq 1$ ),  $sneqv_{\max}$  is the vegetation type-dependent maximum SWE for complete snow fraction, and  $\alpha_s$  is a distribution shape parameter.

The new  $R_{netn}$  is also used in computing the snowmelt rate ( $M_s$ ) based on mass and energy balance in the snowpack,

$$M_s = \frac{R_{es}}{L_f} = \frac{1}{L_f} (R_{netn} - G - LH - SH - F_1 - F_2) \quad (2.5)$$

where  $R_{es} > 0$  is the energy available for the snowmelt,  $L_f$  is the latent heat of fusion,  $LH$  is the latent heat flux,  $SH$  is the sensible heat flux,  $F_1$  is the heat flux from newly accumulating precipitation, and  $F_2$  is the freezing rain latent heat flux.

### 2.2.2 Snow density

To avoid the abrupt change of the snow density when the surface temperature is near melting point (i.e., 0 °C), the portion of liquid water ( $DW$ ) stored in the snowpack during snowmelt ( $M_s$ ) is limited by:

$$DW = 0.13\Delta t / D_{hr} \quad (2.6a)$$

$$DW = \min(DW, 0.13M_s / (sneqv + 0.13M_s)) \quad (2.6b)$$

where  $DW$  is the portion of liquid water stored in the snowpack during snowmelt,  $D_{hr} = 24h$  and  $\Delta t$  is the time step in hours. For both conditions  $T_1 \geq 0$  °C and  $T_{soil(1)} \geq 0$  °C, the snow density ( $\rho_{sn}$ ) is

$$\rho_{sn} = \rho_{sn} (1 - DW) + DW \quad (2.6c)$$

The change in snow density modifies snow depth ( $snowd$ ):

$$snowd = \frac{sneqv}{\rho_{sn}} \quad (2.6d)$$

The limitation setting of snow water helps the maintenance of snow depth.

### 2.2.3 Under-canopy resistance and sensible heat flux

The new snow physics has increased the under-canopy aerodynamic resistance  $r_u$  to prevent overestimation of downward sensible heat flux (SH) under stable conditions (i.e.,  $T_a > T_1$ ). The new formula of under-canopy aerodynamic resistance is based on the study of Sakaguchi and Zeng (2009):

$$r_u = r_{u \max} \bullet GVF \bullet F_{g,sn} \Delta T \times [1 - \exp(-LAI_{eff})] \quad (2.7a)$$

$$\Delta T = \min\left(\frac{T_a - T_1}{5}, 1\right) \quad (2.7b)$$

$$LAI_{eff} = LAI(1 - F_{bur}) \quad (2.7c)$$

where  $r_{u \max}$  is  $100 \text{ sm}^{-1}$ .  $F_{g,sn}$  and  $F_{bur}$  are defined in equations (2.1b) and (2.1c), respectively. The surface exchange coefficient ( $C_H$ ) for heat and moisture is then modified according to the change of the under-canopy aerodynamic resistance:

$$C_{hn} = C_H / (1 + r_u C_H)$$

This, combined with the vegetation shading effect, leads to a modified sensible heat flux ( $SH$ ) as shown below:

$$SH = -\frac{C_{hn} C_p P_s}{RT_v} (T_a - T_1) + SH_c \quad (2.8a)$$

where  $C_p$  is the heat capacity of dry air,  $R = 287.04 \text{ J K}^{-1} \text{ kg}^{-1}$  is the gas constant for dry air,  $P_s$  is surface pressure,  $T_a$  is air potential temperature,  $T_v$  is virtual potential temperature,  $T_1$  is surface skin temperature, and  $SH_c$  represents the vegetation shading effect:

$$SH_c = R_{net} - R_{netn} \quad (2.8b)$$

Most solar heating for forests with underlying snow is used to heat the atmospheric boundary layer through sensible heat. To maintain the energy balance, the different net radiation ( $SH_c$ ) between the non-shaded ( $R_{net}$ ) and shaded effect ( $R_{netn}$ ) is added to the sensible heat flux ( $SH$ ). Since the ( $SH$ ) is also constrained by the solar energy absorbed by the surface (vegetation, snow, and soil), the ( $SH_c$ ) is restrained to be less than half of the solar energy absorbed by the surface (vegetation, snow, and soil) in absolute value in equation (2.2a) (Wang et al., 2010).

#### 2.2.4 Roughness length for momentum under snow conditions

The roughness length for momentum is modified for vegetation under snow conditions to differ from the snow-free roughness length ( $z_0$ ). The new effective roughness length ( $z_{0n}$ ) for the grid cell includes the impact of snow coverage on vegetation:

$$\ln z_{0n} = \left[ (1 - F_{g,sn}^2) \ln z_{0g} + F_{g,sn}^2 \ln z_{0s} \right] (1 - F_v)^2 + \left[ 1 - (1 - F_v)^2 \right] \ln z_{0v} \quad (2.9a)$$

where  $z_{0g} = 0.01$  m is for bare soil,  $z_{0s} = 0.001$  m for snow, and  $z_{0v}$  is a vegetation type-dependent roughness length for momentum and is independent of snow coverage. Value  $F_v$  is the exposed maximum  $GVF$  and is between 0 and 1 as

$$F_v = GVF_{\max} \bullet (1 - F_{bur}) \quad (2.9b)$$

where  $GVF_{\max}$  is the maximum  $GVF$  prescribed for each grid cell based on satellite data.

### 2.3 1-D new snow physics with COAMPS forcing

Before implementing Noah3.6 to the COAMPS Noah LSM, we examined it using a time series of atmospheric forcing and compared the results with those from Noah3.2. The atmospheric forcing is obtained from COAMPS forecasts during October 01-31, 2015, with every 12-h data assimilation cycle. They are the outputs from the innermost nest with a horizontal resolution of 5 km. The time series have a 1-h time interval and are taken at a location (66.2192°N, 150.96°W) inside Alaska. The forcings include wind speed and direction, air temperature and mixing ratio, surface pressure, shortwave and longwave radiation, and precipitation at the first model level (10m). Figure 1 displays variations of these variables during the whole month of October 2015. There are a few high wind periods with the most significant and longest duration during Oct 7-14 (Fig. 1a) associated with significant drops in the surface pressure (Fig. 1f). Air temperature (Fig. 1b) and mixing ratio (Fig. 1c) show cold and dry trends during this particular period. The shortwave radiation (Fig. 1d) shows a strong diurnal cycle. Fig. 1g illustrates a few occasions of precipitation during October.

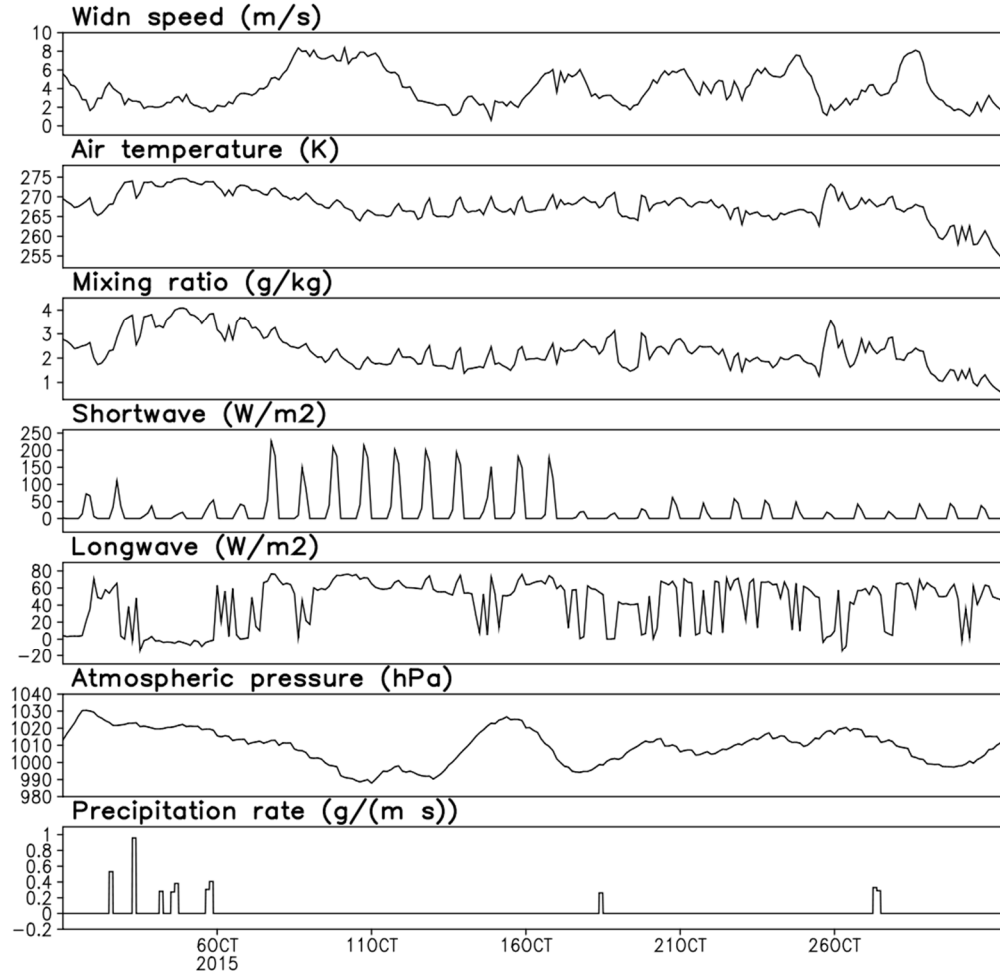


Fig. 2.1. Atmospheric forcings from the lowest COAMPS model level: (a) wind speed, (b) air temperature, (c) mixing ratio, (d) shortwave radiation, (e) longwave radiation, (f) atmospheric pressure, and (g) precipitation rate.

The COAMPS forcings are applied to LSM Noah3.2 and Noah3.6 to run with 20 different options of MODIS vegetation types to examine the effect of the new snow physics with various vegetation heights. As an example, the predicted variables from the two LSM versions are displayed in Figure 2.2 for two vegetation types: a tall evergreen broad-leaf forest and a short wooded tundra. Similar variation features are found for other vegetation types. By adding the vegetation shading effect, the net shortwave radiation (Eq. 2.2) and the net radiative flux received by the surface are reduced in the calculation of potential evapotranspiration (Eq. 2.3). A decrease of the potential evapotranspiration directly reduces the snow sublimation (Eq. 4) along with the reduction of the snowmelt rate associated with the change of net radiative flux (Eq. 2.5). It leads to more snow being retained when using the new snow physics, indicating considerable deeper snow depth than the old LSM, as seen in Fig. 2.2a. After combining it with the limitation on liquid water stored in the snow (Eq. 2.6b), the snow density is reduced in Noah3.6 (Fig. 2.2b).

The new snow roughness length with the vegetation shading effect is compared with the one without the vegetation impact and displayed in Fig. 2.2c. The roughness lengths increase when the vegetation shading effect is considered. The roughness length increases by about 30% for both short and tall vegetation. Without adding vegetation effect, the roughness length mainly depends on climatology under snow

condition that tends to be smaller. The roughness length of evergreen broadleaf remains unchanged, likely because the broadleaf evergreens usually holding their foliage year-round.

Wang et al. (2010) showed that downward sensible heat can be significantly overestimated under stable conditions with strong winds when snow physics is not included. Increasing under-canopy resistance in the new snow physics aims to correct this overestimation. As a result, the significant negative sensible heat flux during the intense wind periods (i.e., 7-14 Oct) from Noah3.2 is reduced by at least 30% from Noah3.6 (Fig. 2.2d). There is no apparent difference between short and tall vegetation, so the sensible heat flux is shown for only one vegetation type for clarity. The downward sensible heat flux reduction ranges are 20-60  $\text{Wm}^{-2}$  for all vegetation types.

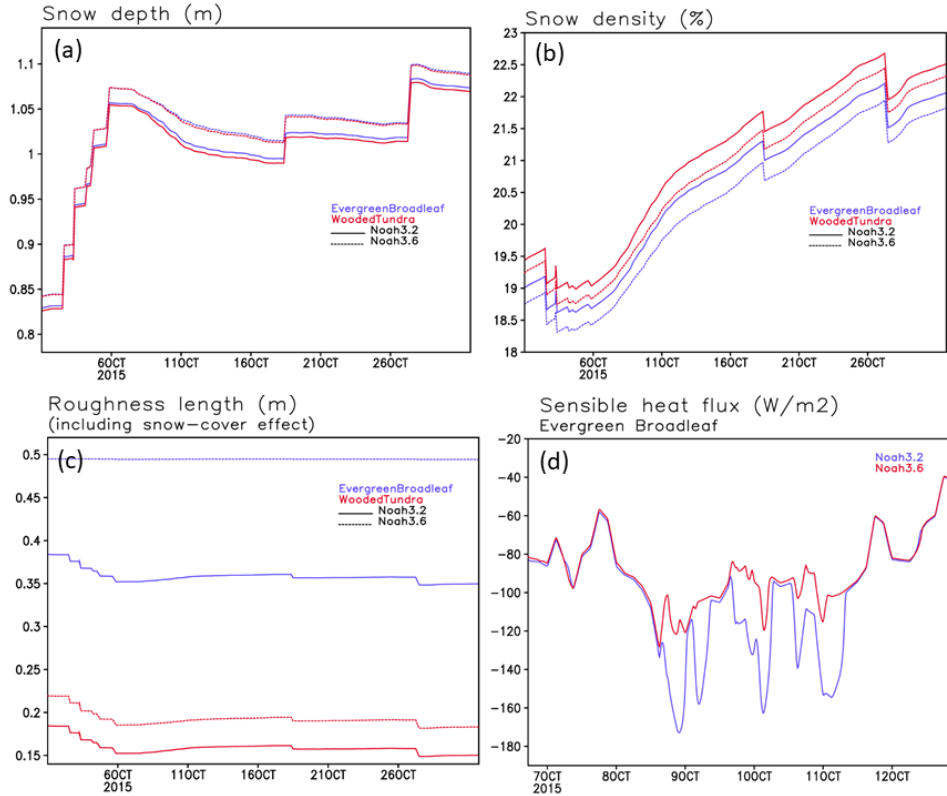


Fig. 2.2. Time series of (a) Snow depth (m), (b) snow density (%), and (c) roughness length (m) for a tall (evergreen broad leaf forest, blue) and short (wooded tundra, red) vegetation type using Noah3.2 (solid) and Noah3.6 (dashed). Sensible heat flux ( $\text{Wm}^{-2}$ ) is displayed in (d) only for evergreen broad-leaf forests using Noah3.2 (blue) and Noah3.6 (red) due to similarity to the features from other types of vegetation. Only time series for October 6 -12 are displayed for (d) since there aren't significant modifications for the other period of October for this particular situation.

## 2.4 Application of the new snow physics in COAMPS Arctic forecast

The new snow physics (Noah3.6) is implemented in COAMPS LSM to replace the old version LSM Noah3.2 for improving Arctic forecasts. Verification of COAMPS real case forecasts over the Arctic area with Noah3.6 is provided using all the available land surface observations. Various relationships associated with new snow physics and stable boundary layers are also discussed in this section.

### 2.4.1 COAMPS and experiments

The COAMPS atmospheric component is a three-dimensional, fully compressible, non-hydrostatic, primitive equation model (Hodur, 1997; Hodur et al., 2002). Subgrid-scale physical parameterizations for

COAMPS include Kain-Fritsch (Kain and Fritsch, 1990) scheme for convection, Rutledge and Hobbs(1983) single-moment scheme for cloud microphysics, Fu-Liou scheme for shortwave and longwave radiation (Fu and Liou,1992; Liu et al., 2009). The planetary boundary layer and free-atmospheric turbulent mixing and diffusion are modeled using a prognostic equation for the turbulent kinetic energy (TKE) budget based on a method derived from the level 2.5 formulation of Mellor and Yamada (1982). Atmospheric data assimilation is the NRL Atmospheric Variational Data Assimilation System (NAVDAS, Daley and Barker 2001) used to assimilate the observations and provide the initial conditions for the COAMPS forecasts. The lateral boundary conditions are derived from the Navy Global Environmental Model (NAVGEN, Hogan et al., 2014) atmospheric forecast fields. The sea surface temperatures used in the COAMPS forecasts are analyses from the Navy Coupled Ocean Data Assimilation (NCODA, Cummings, 2005).

The surface layer (SL) parameterization in COAMPS follows a modified Monin-Obukhov similarity theory (MOST; Monin and Obukhov, 1954) in Louis (1979) and Louis et al. (1982). The Louis scheme matches the MOST with gustiness effects for convective regimes (Bradley et al., 2000; Fairall et al., 2003). An analytical method is developed to solve the stability parameter  $z/L$  ( $z$  is height and  $L$  is Monin-Obukhov length) in the MOST for stable condition (Holtslag and De Bruin, 1988). Detailed formulations can be found in Wang et al. (2002). The surface exchange coefficients for momentum ( $C_d$ ), heat ( $C_h$ ), and moisture ( $C_q$ ) are formulated as follows:

$$C_d = \frac{u_*^2}{U^2} = \frac{k^2}{\ln(1 + z_a / z_{0M})^2} F_M(z_a / z_{0M}, R_b) \quad (2.10a)$$

$$C_h = \frac{u_* \theta_*}{U \Delta \theta} = \frac{k^2}{\ln(1 + z_a / z_{0M}) \ln(1 + z_a / z_{0H})} F_H(z_a / z_{0M,H}, R_b) \quad (2.10b)$$

$$C_q = \frac{u_* q_*}{U \Delta q} = \frac{k^2}{\ln(1 + z_a / z_{0M}) \ln(1 + z_a / z_{0q})} F_q(z_a / z_{0M,q}, R_b) \quad (2.10c)$$

where  $U$ ,  $\theta$ , and  $q$  indicate the total wind speed, potential temperature, and water vapor mixing ratio at the lowest model level  $z_a$ , respectively. The differences in potential temperature and moisture deficit between  $z_a$  and the surface are denoted as  $\Delta \theta$  and  $\Delta q$ . The  $z_{0M,H,q}$  terms are the aerodynamic, thermal, and moisture roughness lengths. The bulk Richardson number ( $R_b$ ) is calculated by

$$R_b = \frac{g \Delta z \Delta \theta_v}{\theta_v \Delta U^2} \quad (2.11)$$

where  $\Delta \theta_v$  and  $\Delta U$  are the differences in virtual potential temperature and wind speed between the top and bottom of the layer of thickness  $\Delta z$ . Solutions of  $F_M$ ,  $F_H$ , and  $F_q$  are solved for stable and unstable conditions as illustrated in Wang et al. (2002). In the Noah LSM, it, like most numerical models, assumes that the surface exchange coefficient for moisture ( $C_q$ ) is assumed to be identical to the heat exchange coefficient ( $C_h$ ) (Chen et al., 1997). Therefore, the surface exchange coefficient for heat ( $C_h$ ) is passed to the Noah LSM subroutines to calculate sensible and latent heat flux in the COAMPS. Before providing it to the LSM subroutines,  $C_h$  has been multiplied by wind speed:

$$C_H = U C_h$$

where  $C_H$  is technically a conductance. The formulation of sensible heat flux has been discussed in Eq. (8) to include the vegetation shading effect. Total evaporation ( $E$ ) for latent heat flux is summed from the direct evaporation of bare soil ( $E_s$ ), the evaporation of precipitation intercepted by the canopy ( $E_w$ ), and transpiration via the canopy and roots ( $E_c$ ) as follows:

$$E = E_s + E_w + E_c \quad (2.12a)$$

in which only the  $E_c$  formulation has used the moisture exchange coefficient ( $C_q$ ) to calculate the plant coefficient ( $P_c$ ) for the reduction in transpiration due to internal plant physiology:

$$E_c = \sigma_f (1 - \beta_c) E_p P_c \quad (2.12b)$$

Here  $P_c$  is a function of the thermodynamic properties ( $\Delta$  and  $R_r$ ), the surface exchange coefficient for moisture ( $C_q$ ), and the canopy resistance ( $R_c$ ) as

$$P_c = \frac{1 + (\Delta / R_r)}{1 + (\Delta / R_r) + R_c C_q} \quad (2.12c)$$

The detailed formulations for computing these variables can be found in (Chen et al., 1997) and (Koo et al., 2017).

The Noah land surface model (LSM), along with the surface layer parameterization in COAMPS as described above, are initialized using global analyses from the NASA Land Information System (LIS). The U.S. Air Force 557th Weather Wing (557WW) produced the analyses and transferred daily to the FNMOC for COAMPS operational forecasts. The analyses were generated using the LIS7.0 with a horizontal resolution of  $0.25^\circ$  in the past and updated using the LIS7.2 with the LSM Noah3.6 and a horizontal resolution of  $\sim 10$  km currently. Therefore, the COAMPS LSM is updated to Noah3.6 to match the physics with the LIS7.2 analyses.

Numerical simulations are conducted to examine the impact of the new snow physics on the COAMPS forecast under stable atmospheric conditions during a winter month. The COAMPS domain is designed for two nests with horizontal resolutions of 45 and 15 km, respectively. Distributions of vegetation and soil types, sea and land ice, are shown in Fig. 2.3, with a large variety of vegetation types from the shortest to the tallest. The soil types are most loam and sandy loam. The different vegetation and soil types, and sea and land ice are considered in the computation of the surface fluxes in Noah LSM. The simulation period is from January 01 to 31, 2021, with every 6-h data assimilation cycle. The forecast length for each simulation initiated every 6-h is 48-h. Two experiments are conducted with the control run using Noah3.2 and the sensitivity run using Noah3.6.

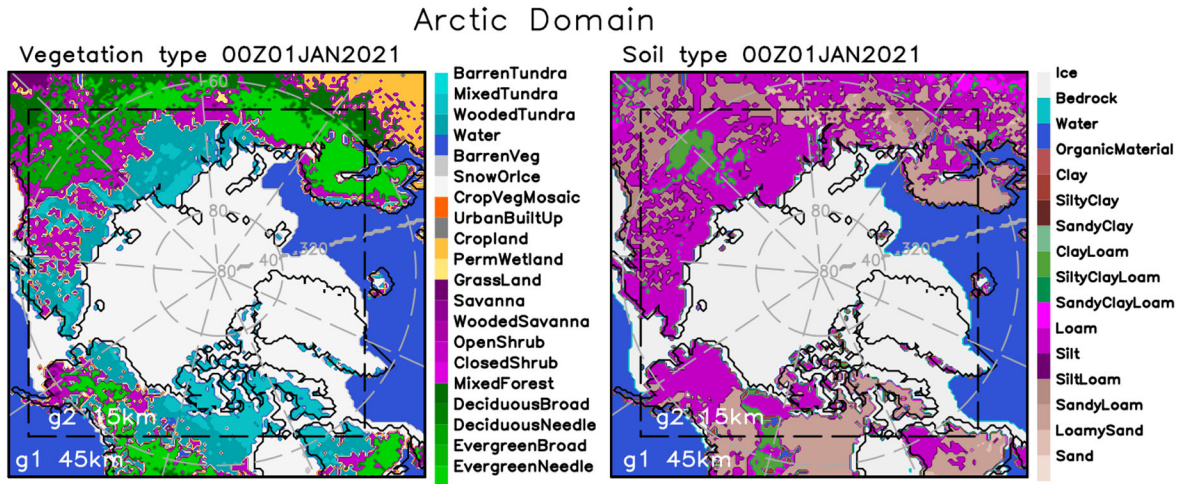


Fig. 2.3. The two-nested domain with (a) MODIS (Moderate Resolution Imaging Spectroradiometer) vegetation type and (b) soil type. Horizontal resolutions are 45 and 15 km for nests one and two, respectively.

#### 2.4.2 Surface Condition

According to Wang et al. (2010), the most significant benefit of adding the new snow physics in LSM is substantial improvement in snow water equivalent and snow depth forecasts. In comparing results from one-dimensional simulations using Noah3.2 and Noah3.6 (Fig. 2.2), the new snow physics reduces the potential evapotranspiration and snow density. As a result, snow depth increases since the net shortwave radiation and snow melt decrease. The monthly mean 48-h forecast of snow depth shows that the snow depth is better retained in Noah3.6 with the new snow physics (Fig. 2.4b) than in Noah3.2 (Fig. 2.4a). The new net shortwave radiation in Noah3.6 is the weighted average of fluxes between the vegetation shaded and non-shaded areas. The reduction of the net radiation results in less potential evapotranspiration in equation (2.3a) and decreases the snow melt rate due to the available net heat flux in equation (2.5). Snow sublimation depends on potential evapotranspiration; therefore, it is reduced as described in equation (2.4). Snow water stored in the snowpack during snowmelt has set a limitation so that the snow density is not encountering an abrupt change as the adjustment in equation (2.8). The decreased snow melt and limited snow water in Noah3.6 increase snow depth. The domain-averaged snow depth has shown that the mean snow depth from Noah3.6 (~279 mm) has increased by 4.2% (~11 mm) over the land area, as compared to Noah3.2 (~268 mm).

Adding the vegetation shading effect on under-canopy aerodynamic resistance reduces the downward sensible heat flux for the stable condition, as discussed in Fig. 2.2c. The downward sensible heat flux reduction can be seen from the monthly mean of COAMPS 48-h forecasts using the LSM Noah3.6 (Fig. 2.4e) compared to Noah3.2 (Fig. 2.4d). The difference between using Noah3.6 and Noah3.2 show the decrease of the downward sensible heat flux over the most land area (Fig. 2.4f). The increased snow depth from Noah3.6 (Fig. 2.4b) prevents a reduction in the ground temperature and reduces the surface stability. Therefore, it reduces the overestimate of the downward sensible heat flux compared to Noah3.2.

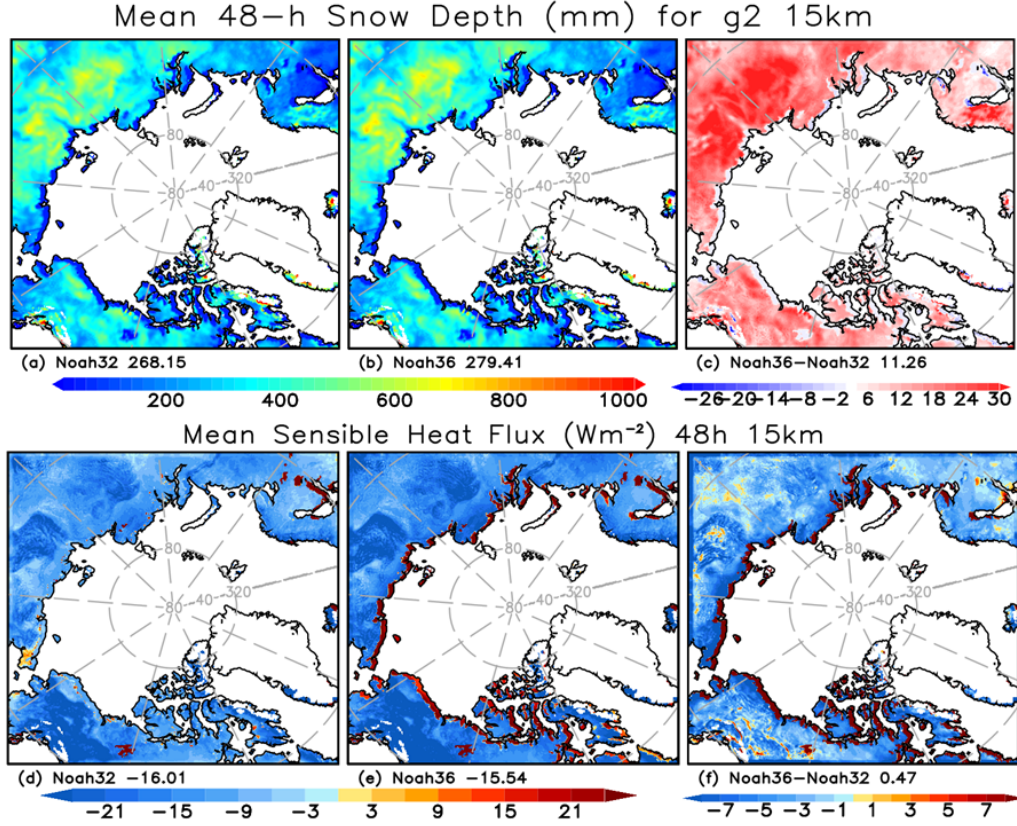


Fig. 2.4. Monthly mean 48-h forecasts of snow depth (upper panel) and sensible heat flux (lower panel) from COAMPS nest two over the land area for (a) control run using Noah3.2 (left column), (b) sensitivity run using Noah3.6 (middle column). The differences between the two experiments are displayed in the right column. The domain-averaged values are also included under each corresponding plot.

Corresponding to the less downward sensible heat flux, the 10-m air temperature (Fig. 2.5a) and surface temperature (Fig. 2.5b) are warmer from Noah3.6 than Noah3.2. On the other hand, compared to the 10-m surface air temperature, the surface temperature appears significantly colder from Noah3.2, which is consistent with less snow depth, as shown in Fig. 2.4a. The difference between the 10-m air and surface temperature is about 1-4 °C more from Noah3.2, which leads to more significantly larger downward sensible heat fluxes in Fig. 2.4d. It also implies a more robust surface temperature inversion that results from a more stable surface condition from Noah3.2 (Fig. 2.5d) than Noah3.6 (Fig. 2.5e). Here we refer the surface temperature inversion as the atmosphere at the surface ( $T_s$ ) is colder than the layer above it ( $T_{10m}$ ). The strength of the surface temperature inversion is defined as the magnitude of the temperature difference ( $T_{10m} - T_s$ ).

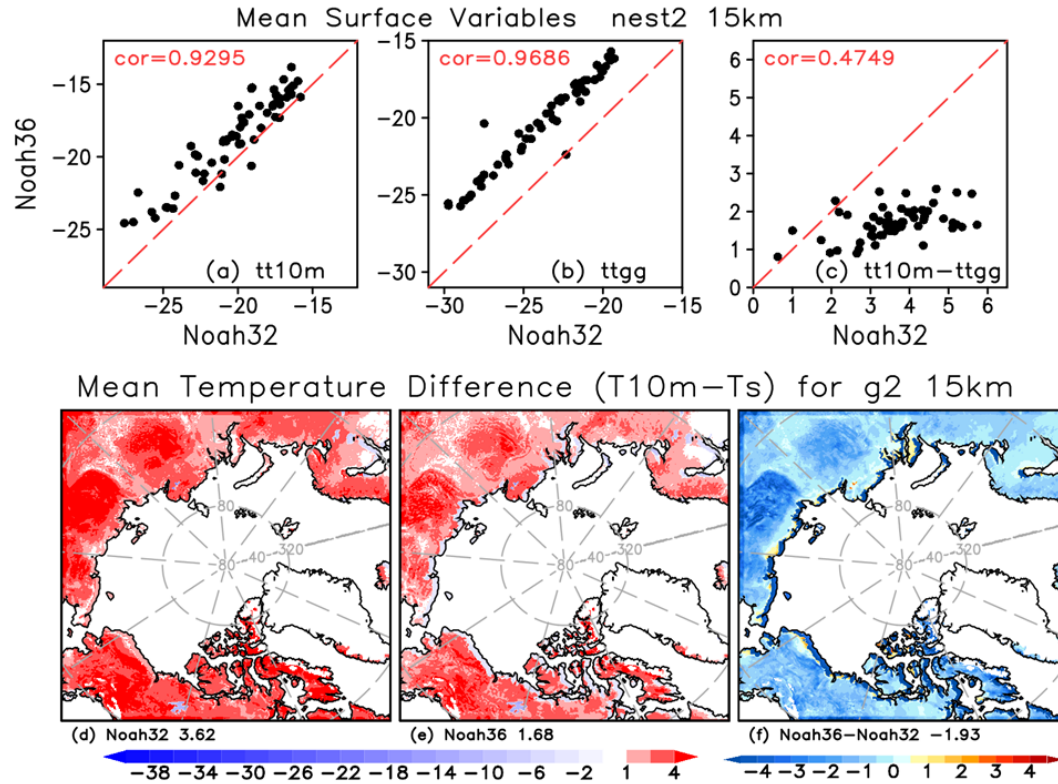


Fig. 2.5. The upper panel is scatter plots for 10-m air temperature (first atmospheric level) and surface temperature from (a) control run using Noah3.2, (b) sensitivity run using Noah3.6, and (c) the difference between (a) and (b). The lower panel is the mean temperature differences between 10 m and surface for (d) from Noah3.2, (e) from Noah3.6, and the differences between (e) and (d).

### 2.4.3 Verification and Discussion

Standard verification for the COAMPS forecasts uses the ADP observational data set. The observations include pressure, geopotential height, temperature, dew point temperature, wind direction and speed. All available observations are operationally collected from the surface station, radiosonde, buoy, pibal, and aircraft reports from the Global Telecommunications System (GTS), radar, SSM/I and satellite data (<https://doi.org/10.5065/Z83F-N512>). Figure 2.6 shows mean biases and root-mean-square (RMS) errors of 48-h COAMPS forecasts from 30 model-observation pairs for all surface station locations. Reduction of cold biases can be found by the sensitivity run using Noah3.6 (Fig. 2.6b), compared to the control run using Noah3.2 (Fig. 2.6a). Differences in the RMS error between using Noah3.6 and Noah3.2 in Fig. 2.6c show more negative values, indicating smaller RMS errors by using Noah3.6. Similar reductions of the RMS error are also evident for dew-point temperature (Fig. 2.6d), sea-level pressure (Fig. 2.6e), and 10-m wind speed (Fig. 2.6f). The new snow physics has improved COAMPS 48-h forecast skills for surface temperature, moisture, pressure, and wind.

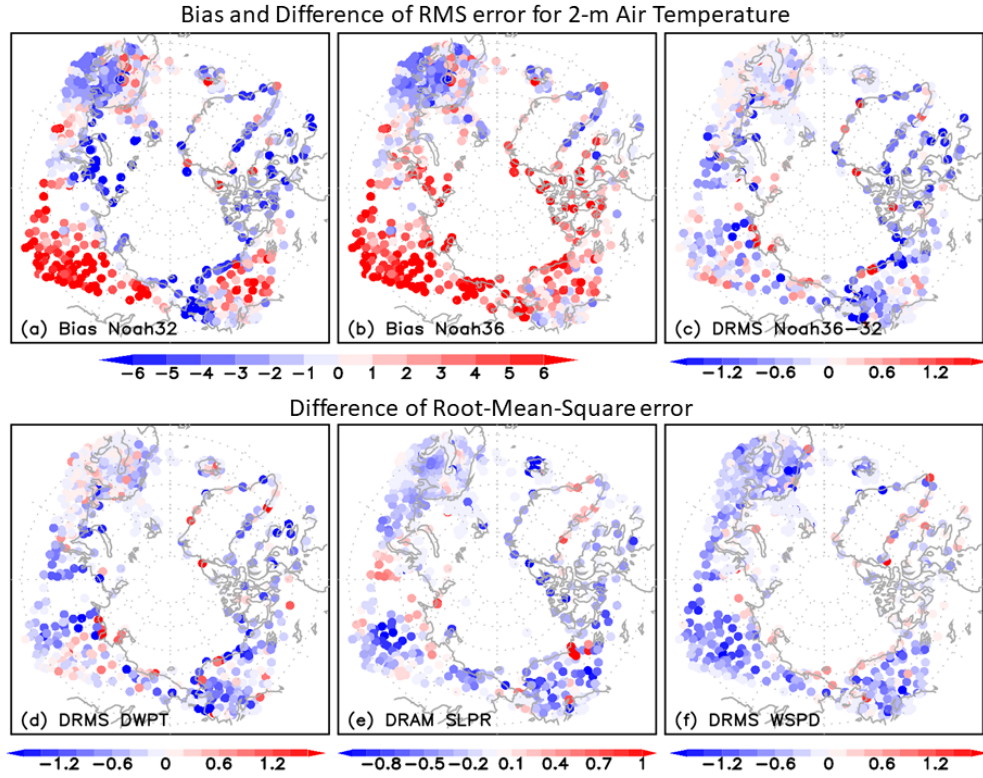


Fig.2.6. Biases of 48-h COAMPS 2-m air temperature (2mT) forecasts with station observations from using Noah3.2 (a) and Noah3.6 (b). Differences in the root-mean-square (RMS) errors for 48-h 2mT between control run using Noah3.2 and sensitivity run using Noah3.6 are displayed in (c). Differences in the RMS errors for 2-m dew-point temperature, sea-level pressure, and 10-m wind speed are displayed in (d), (e), and (f), respectively.

The verification also shows domain-averaged time series of mean biases and the RMS errors in Figure 2.7. The new snow physics in Noah3.6 has significantly reduced the biases and RMS errors for all forecast lengths from 6 to 48 h. Confidence levels reach 99% for the biases and RMS errors, indicating that the verifications are statistically significant.

The reduction of bias and RME errors for 2-m air temperature (Fig. 2.7a) is consistent with the effect of the new snow physics, as discussed in section 4b. The mean cold bias of  $\sim 1^\circ\text{C}$  using Noah3.2 is reduced to  $\sim 0^\circ\text{C}$  using Noah3.6. It is the most beneficial variable from the increased snow depth (Fig. 2.4b) and decreased downward sensible heat flux (Fig. 2.4e) with the new snow physics in Noah3.6. The 2-m dew-point temperature (Fig. 2.7b) and relative humidity (Fig. 2.7c), sea-level pressure (Fig. 2.7d), and 10-m wind speed (Fig. 2.7e) and wind direction (Fig. 2.7f) in the sensitivity run using the new snow physics in Noah3.6 (Fig. 2.7) have also shown improvements.

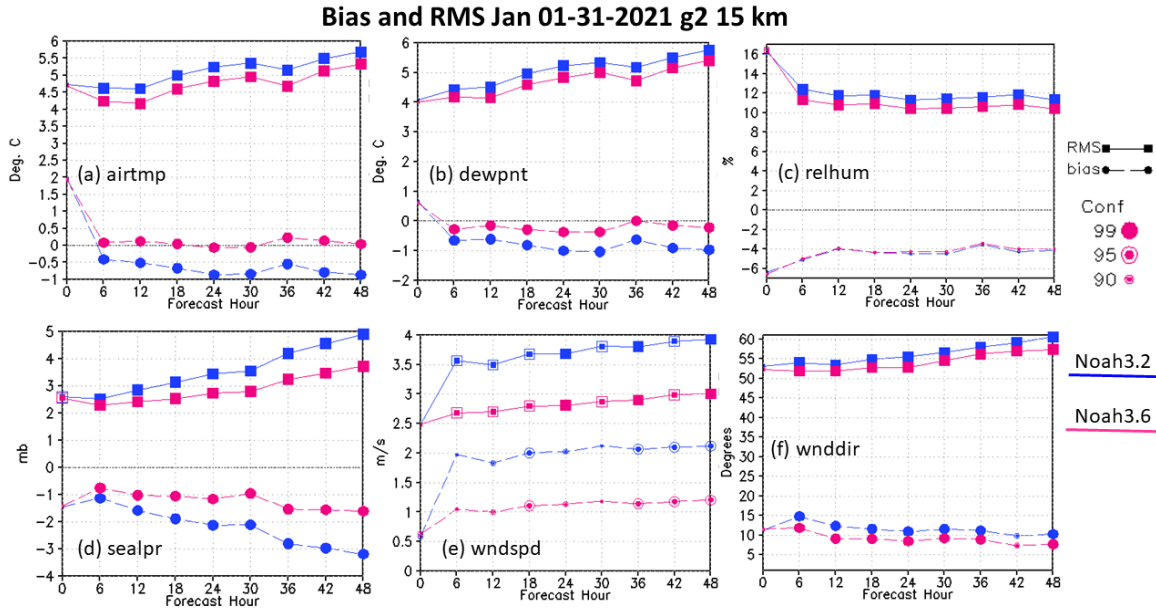


Fig. 2.7. Mean bias (dashed lines) and RMS (solid lines) of COAMPS surface forecasts from Jan 01-31, 2021, of (a) 2-m air temperature, (b) 2-m dew-point temperature, (c) 2-m relative humidity, (d) sea-level pressure, (e) 10-m wind speed and (f) 10-m wind direction for control run using Noah3.2 (blue color) and sensitivity run using Noah3.6 (red color). Statistic confidence levels are marked as solid circles for bias and squares for RMS.

Figure 2.8 shows the time series of domain-averaged snow depth from 48-h forecasts using Noah3.2 (blue line) and Noah3.6 (black line), compared to the LIS analyses (red line). The LIS data assimilation integrates observations with model forecasts to generate improved estimates of land surface conditions, including snow coverage (Kumar et al. 2006). The same snow depths from the LIS analyses are used to initialize LSM Noah3.2 and Noah3.6. However, the 48-h forecasted snow depths from Noah3.6 are closer to the LIS analyses than Noah3.2. The better match from Noah3.6 is reasonable since the LIS analyses have also used the same LSM Noah3.6. It confirms that without a good LSM, using the improved initial conditions does not ensure the improvement of model forecasts.

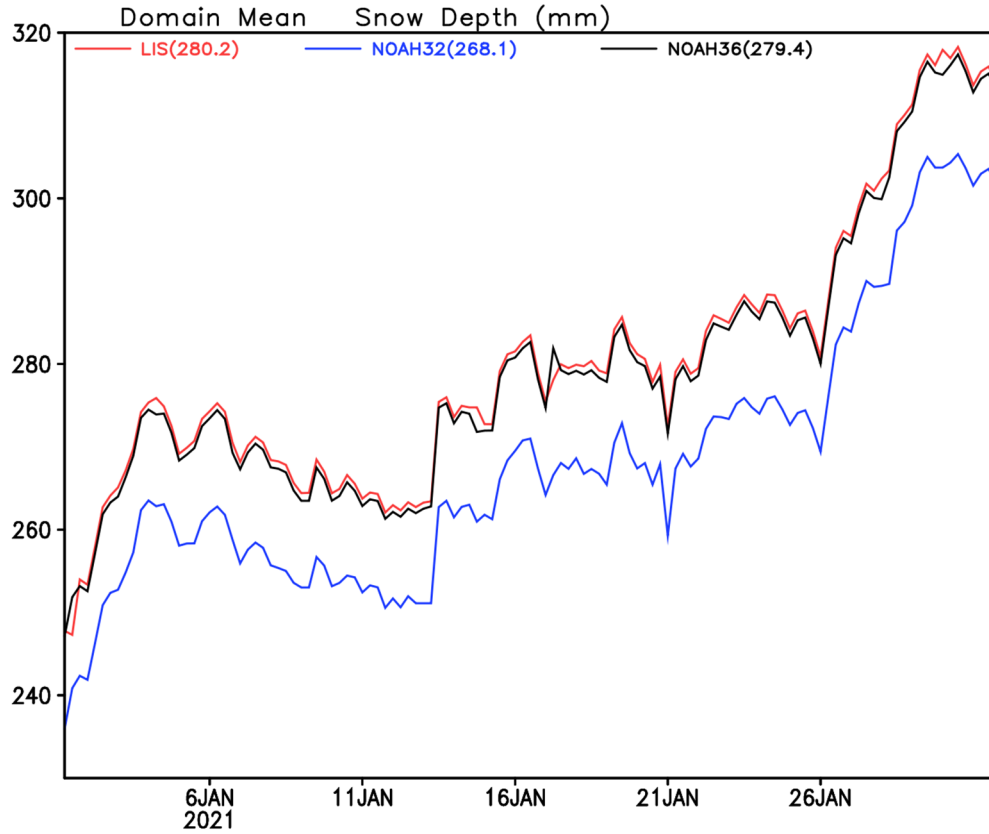


Fig. 2.8. Domain-averaged 48-h forecast snow depth from LIS analyses (red line), COAMPS LSM using Noah3.2 (blue line), and Noah3.6 (black line).

#### 2.4.4 Turbulent mixing in the surface layer

A histogram of 48-h 2-m air temperature over the land points shows that a higher frequency of colder temperature occurs at -25 to -35 °C from Noah3.2 (Fig. 2.9a) than Noah3.6 (Fig. 2.9b). It provides detailed information for the frequency range of decreased air temperature. Since the 2-m air temperature is derived from the 10-m air temperature using the similarity theory, they have similar features and appear colder than the 2-m air temperature. It is consistent with the stronger and higher inversion from Noah3.2, as presented in Fig. 2.5. Response to this change is the decreased boundary layer height (BLH); as seen in Fig. 2.9e, its frequency is higher under 100 m and lower above 100 m.

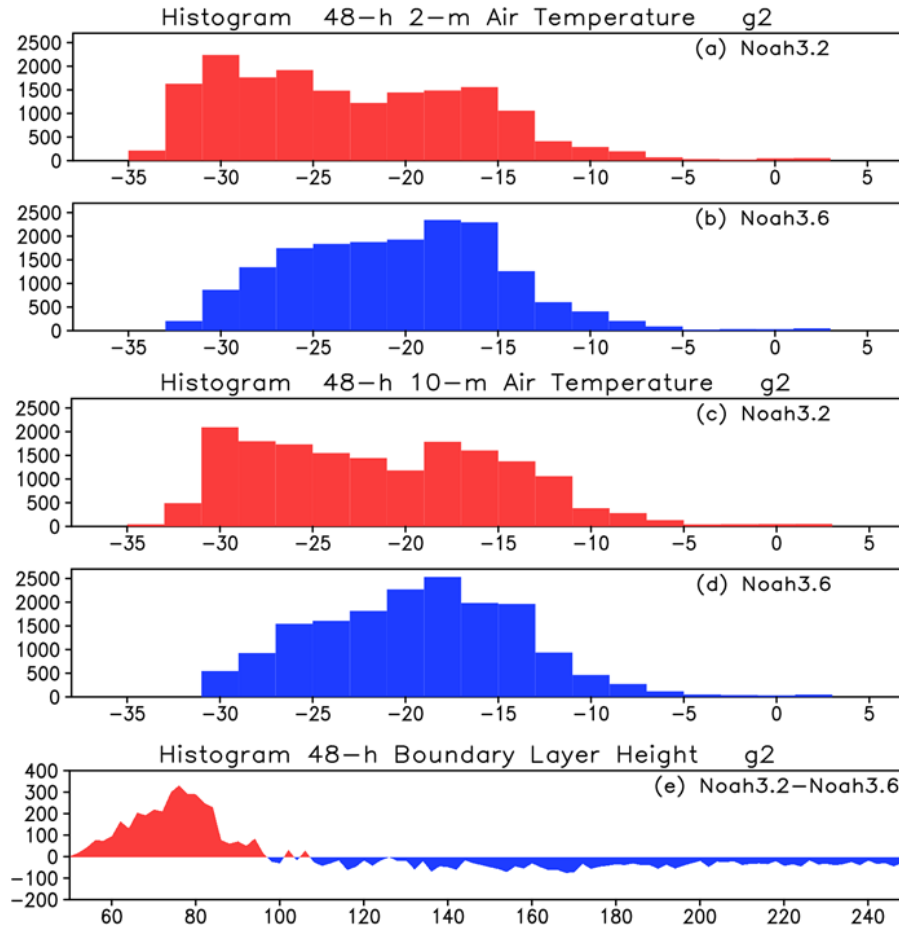


Fig. 2.9. Histogram of 48-h 2-m (upper two panels) and 10-m (middle two panels) air temperature, and differences of boundary layer height between the control run using Noah3.2 and the sensitivity run Noah3.6 (lower panel). All values are over the land points only.

Scatter plots for eddy mixing coefficients for momentum ( $K_m$ ) and heat ( $K_h$ ) at 1000 hPa show stronger mixing from the sensitivity run using Noah3.6 than the control run using Noah3.2 (Fig. 2.10a and 2.10b). The correlations of eddy mixing coefficients between Noah3.6 and Noah3.2 are only 0.49 for  $K_m$  and 0.51 for  $K_h$ . Both coefficients are around  $1.0 \text{ m}^2 \text{ s}^{-1}$  for Noah3.2 and in the range of  $1.0\text{-}3.2 \text{ m}^2 \text{ s}^{-1}$  for Noah3.6, indicating that much weaker turbulent mixing occurs in Noah3.2 than in Noah3.6. Turbulence kinetic energy (TKE) has a consistent change in magnitude and a high correlation between these two experiments. However, more substantial turbulence is from Noah3.6 (Fig. 2.10c), consistent with the warmer air temperature, as shown in Fig. 2.6.

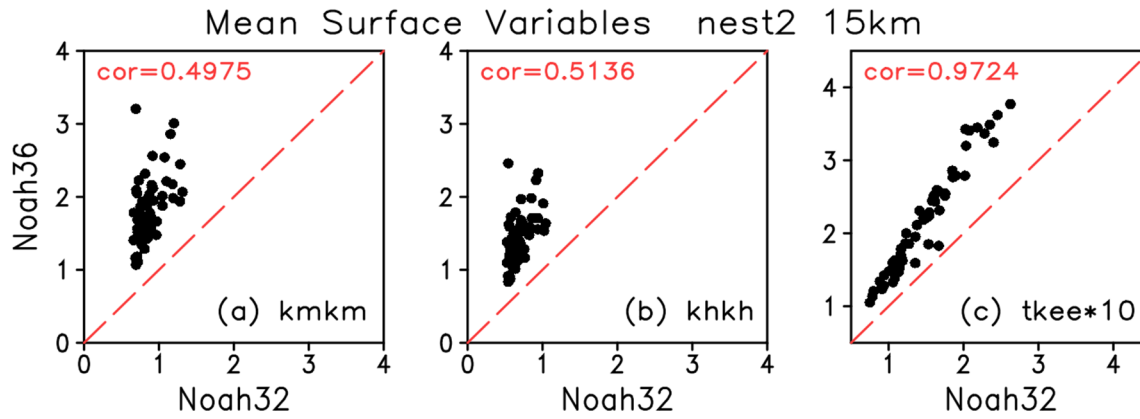


Fig. 2.10. Scatter plots between the sensitivity run using Noah3.6 and the control run using Noah3.2 for eddy mixing coefficients ( $m^2 s^{-1}$ ) of (a) momentum and (b) heat and for (c) turbulence kinetic energy ( $\times 10 m^2 s^{-2}$ ) at 1000 hPa over the land points.

#### 2.4.5 Influences to the lower boundary layer

Time series of vertical profiles for potential temperature and turbulence kinetic energy (TKE) show warmer temperatures and more vigorous vertical mixing from using the new snow physics (Fig. 2.11b) than the old LSM (Fig. 2.11a). Correspondingly, the boundary layer is about 100-500 meters higher (Fig. 2.11c), consistent with the higher frequency for the deeper boundary layer (Fig. 2.9c). The warmer surface temperature and less stable boundary layer promote vertical mixing and prevent the decoupling between the land surface and the atmospheric surface layer. The improvement is significant in the second half of the month when cold weather is more pronounced.

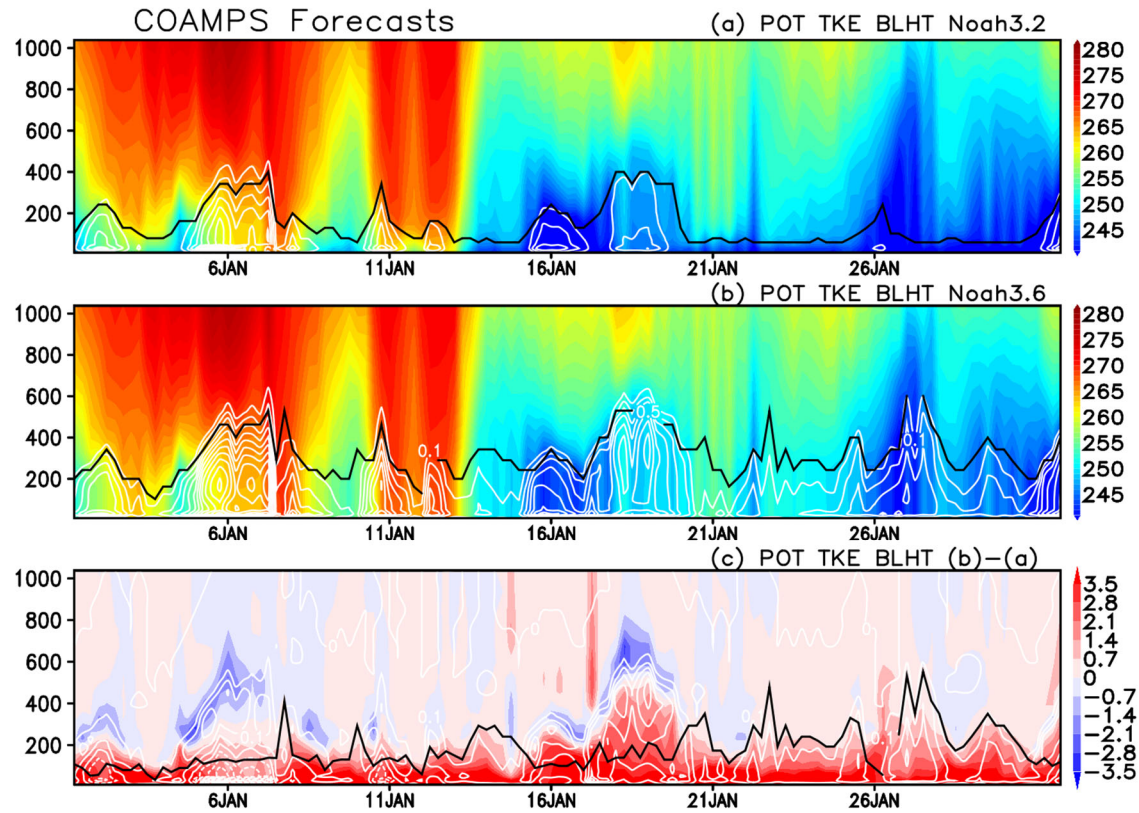


Fig. 2.11. Time series of COAMPS forecasts of vertical profiles for potential temperature ( $^{\circ}\text{K}$ , color shaded), TKE ( $\text{m}^2\text{s}^{-2}$ , white contours), and boundary layer height (m, black contour) from using (a) the control run using Noah3.2, (b) the sensitivity run using Noah3.6 and (c) difference between (b), and (a) at a location ( $72.6804^{\circ}\text{N}$ ,  $77.3677^{\circ}\text{E}$ ).

Vertical profiles of potential temperatures further prove the overall less stable lower boundary layer from the sensitivity run using Noah3.6 than the control run using Noah3.2 (Fig. 2.12a). Influences of modifying the land surface model with the new snow physics have reached  $\sim 300$  m height. Vertical mixing is about 1-2 times stronger from Noah3.6 than Noah3.2 (Fig. 2.12b). The bulk Richardson number ( $R_b$ ) shows a significantly larger value from Noah3.2 than Noah3.6 (Fig. 2.12c), indicating less mixing and result in much more stable lower boundary condition. It leads to very stable (Fig. 2.12a) and robust stratified conditions (Fig. 2.11a) without using the new snow physics.

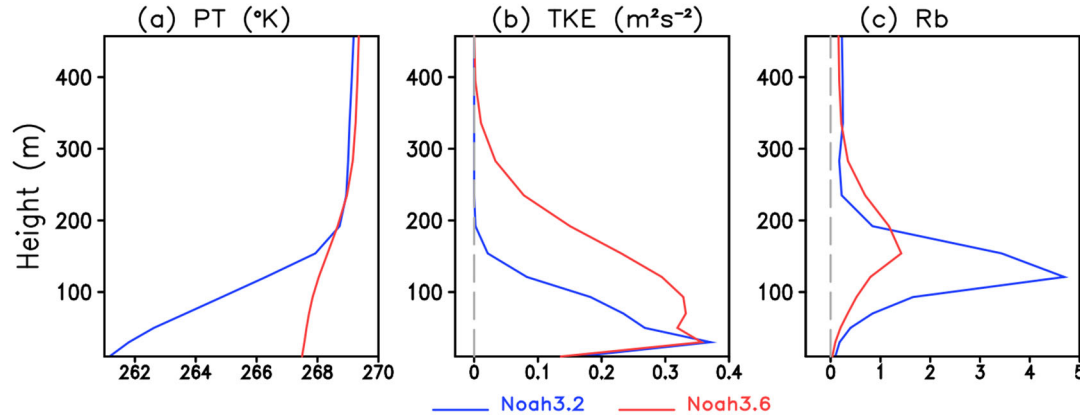


Fig. 2.12. Vertical profiles of (a) potential temperature ( $^{\circ}\text{K}$ ), (b) TKE ( $\text{m}^2\text{s}^{-2}$ ), and (c) bulk Richardson number ( $R_b$ ) from the control run using Noah3.2 (blue) and the sensitivity run using Noah3.6 (red) at 06Z January 11, 2021.

## 2.4 Impact of lateral boundary conditions

Benefits from larger area forecasts to obtain smaller area, higher quality forecasts using lateral boundary conditions (LBCs) derived from the larger-area model have been recognized decades ago (Mesinger, 2001). The values of the dependent variables at the lateral boundary grid points are used to solve the governing equations for the limited area model (LAM) (Davis, 1976; Perkey and Kreitzberg, 1976). For operational atmospheric forecasts, the LBCs are obtained by interpolation from values of dependent variables at lateral grid points of a previously run global atmospheric forecast model. It becomes a standard for various operational centers that a nested LAM is driven by boundary conditions of another global model (Chandrasekar, 2022). The operational COAMPS forecasts conducted at the FNMOC have used the most recent global forecasts from NAVGEM to obtain the LBCs (Hodur, 1997). As presented in the previous section 4, the newly implemented snow physics has been verified in the standard manner using NAVGEM forecasts for the LBCs to fulfill the operational forecasts.

For non-operational research purposes, the LBCs for the LAM have options to use archived regional or global analyses due to their availability since the analyses have combined the optimal atmospheric model output with all possible atmospheric observations (Chandrasekar, 2022). It is widely thought that the LBCs are a source of significant errors in LAMs, but the LBC errors are shown to be a tiny part of the overall error (Davies, 2014). To examine the impact of the LBCs on the COAMPS forecast errors, we have performed two additional simulations, using the same settings as for the simulations presented in Section 4, except that the LBCs have been obtained from the global NAVGEM analyses. The results of using different LBCs between the NAVGEM forecasts in the previous two simulations and the NAVGEM analyses in the additional two simulations are compared and discussed below.

Standard verifications using the ADP observational data set are performed for these two additional simulations using the method described in Figure 2.6 and 2.7. Table 2.1 displays the mean biases and RMS errors of COAMPS 48-h forecasts for all four simulations. They include simulations using the LBCs obtained from the global NAVGEM forecasts (i.e., the two simulations in Section 4) and analyses (i.e., the two additional simulations) with the LSM Noah3.2 and Noah3.6, respectively. Differences in the biases and RMS errors between using the NAVGEM forecasts and analyses for the LBCs are displayed in columns 4 and 7 for using the LSM Noah3.2 and Noah3.6, respectively. Overall, the differences in 48-h forecast biases and RMS errors between using two different LBCs are insignificant, which agrees with the previous

investigation (Davies, 2014). The results also show that using the LBCs from the analyses only sometimes provides smaller forecast biases and RMS errors than using the LBCs from the forecasts.

Table 2.1. Mean biases and RMS errors of COAMPS 48-h forecasts using LBCs from the global NAVGEM forecasts (fcst) and analyses (anal) with the LSM Noah3.2 and Noah3.6.

|                  | Noah3.2<br>(fcst) | Noah3.2<br>(anal) | Noah3.2<br>(fcst-anal) | Noah3.6<br>(fcst) | Noah3.6<br>(anal) | Noah3.6<br>(fcst-anal) |
|------------------|-------------------|-------------------|------------------------|-------------------|-------------------|------------------------|
| Bias tt2m (°C)   | -0.77             | -0.74             | -0.03                  | 0.19              | 0.46              | -0.27                  |
| Bias slpr (mb)   | -3.22             | -3.86             | 0.64                   | -1.64             | -3.37             | 1.73                   |
| Bias wspd (ms-1) | 2.07              | 2.11              | -0.04                  | 1.10              | 1.34              | -0.24                  |
| Bias dwpt (°C)   | -0.86             | -0.84             | -0.02                  | -0.10             | 0.23              | -0.32                  |
| Bias relh (%)    | -4.00             | -3.84             | -0.16                  | -3.46             | -3.28             | -0.18                  |
| RMS tt2m (°C)    | 5.50              | 5.41              | 0.10                   | 5.20              | 5.16              | 0.04                   |
| RMS slpr (mb)    | 4.84              | 5.07              | -0.22                  | 3.69              | 4.59              | -0.90                  |
| RMS wspd (ms-1)  | 3.88              | 3.87              | 0.02                   | 2.96              | 3.13              | -0.17                  |
| RMS dwpt (°C)    | 5.57              | 5.43              | 0.14                   | 5.25              | 5.20              | 0.05                   |
| RMS relh (%)     | 11.56             | 11.40             | 0.16                   | 10.38             | 10.63             | -0.26                  |

The impact of the LBCs on the forecasts of a stable boundary layer is examined from the results of the four simulations, as displayed in Table 2.1. Potential temperatures within the boundary layer, as shown in Fig. 2.7, have minimum values on the surface to indicate the strength of the stable condition. The potential temperatures are warmer using the LSM Noah3.6 than Noah3.2, and the results are consistent, whether using the LBCs with NAVGEM forecasts or analyses. The differences in minimum potential temperature between the two LBCs are negligible for both Noah3.2 and Noah3.6. The maximum bulk Richardson number within the boundary layer and the mean boundary layer height further explain the improvements using Noah3.6 through the increased turbulent mixing. They also show no significant impact when using two different LBCs obtained from the forecasts and analyses.

Table 2.2. Minimum of potential temperature, maximum of bulk Richardson number within the boundary layer and mean boundary layer height during the simulation periods.

|               | Noah3.2<br>(fcst) | Noah3.2<br>(anal) | Noah3.2<br>(fcst-anal) | Noah3.6<br>(fcst) | Noah3.6<br>(anal) | Noah3.6<br>(fcst-anal) |
|---------------|-------------------|-------------------|------------------------|-------------------|-------------------|------------------------|
| Min_pott (°K) | 261.17            | 261.16            | 0.01                   | 267.49            | 267.48            | 0.01                   |

|               |        |        |       |        |        |        |
|---------------|--------|--------|-------|--------|--------|--------|
| Max_Rich      | 4.69   | 5.73   | -1.04 | 1.42   | 1.26   | 0.15   |
| Mean_blht (m) | 137.81 | 136.80 | 1.01  | 309.52 | 345.19 | -35.67 |

## 2.5 Concluding remarks

The forecast over the Arctic region remains a great challenge to the NWP community due to constant stable boundary layer conditions during day and night. Decoupling between the land surface and atmosphere due to inadequate turbulent heat flux often causes cold bias and induces positive feedback on the surface air temperature. To prevent the decoupling and reduce the forecast errors over the Arctic region, the new snow physics is added to the COAMPS Noah LSM to improve the surface heat flux. The impact of the new snow physics on the stable boundary layer is also investigated over the Arctic land surface.

The major additions in the new snow physics include adding the vegetation shading effect on snow sublimation and snow melt, limiting the liquid water stored in the snowpack during the snowmelt, differing the roughness length for vegetation under snow and snow-free conditions, and modifying under-canopy resistance to prevent overestimating downward sensible heat flux.

The new snow physics is examined using COAMPS forecasts during October 2015 as surface forcing in one-dimensional mode and compared with the old Noah LSM. The results confirm the expected improvements by the new snow physics for all MODIS 20 vegetation types. The vegetation shading effect increases the snow depth and decreases the snow density, reducing excessive snow melt in the old LSM. The downward sensible heat flux decrease is consistent with the increased under-canopy resistance for the snow condition. The new roughness length has shown the impact of snow coverage on vegetation to avoid small values under snow conditions.

The new snow physics is used in COAMPS Arctic forecasts for the winter of January 2021 to verify its improvement on the persistent cold bias over the land surface. Monthly mean 48-h predictions have confirmed the consistent results from the original paper on the new snow physics (Wang et al., 2010). Deeper snow depth and less downward sensible heat flux have resulted in warmer surface air and ground temperature. In addition, it has led to a more negligible temperature difference between surface air and ground and weaker surface inversion. Verification using all available surface land observations has shown a significant reduction of cold bias in 2-m air temperature, with the mean value changed from -1 to 0°C. The improvements can also be seen for variables such as 2-m dew-point temperature and relative humidity, sea-level pressure, and 10-m wind speed and direction. The 48-h forecast snow depths from Noah3.6 better match the Land Information System (LIS) analyses.

On the other hand, the histogram analysis indicates a higher, colder temperature frequency occurring at -25 to -35 °C when using the old LSM. Response to this change is the higher frequency of the boundary layer larger than 100 m from the new snow physics. Further analysis shows more significant surface eddy mixing coefficients for momentum and heat and more extensive turbulence kinetic energy produced by the new snow physics. Time series of vertical profiles have shown that warmer temperatures, more vigorous vertical mixing, and higher boundary layer have been forecasted when using the new snow physics, especially during the cold period of the second half of the simulation month. The analyses of vertical profiles for TKE and bulk Richardson numbers denote the impact of the snow physics reaching ~300 m in the lower boundary layer.

Two additional simulations are performed to investigate the impact of the LBCs on the forecasts. Instead of using the NAVGEM forecasts for the LBCs, the two new simulations use the LBCs obtained from the NAVGEM analyses. Overall, the differences in 48-h forecast biases and RMS errors from using two different LBCs are insignificant, which agrees with the previous investigation (Davies, 2014). The results also show that using the LBCs from the analyses only sometimes provides smaller forecast biases

and RMS errors than using the LBCs from the forecasts. No significant differences in the stable boundary layer prediction are found from using the two LBCs since the impact of the LBCs on the turbulent vertical mixings is similar.

The implemented new snow physics in COAMPS Noah LSM has passed the operational scorecard and transited to the Fleet Numerical Meteorology and Oceanography Center (FNMOC) for operational forecasts. It helps researchers investigate stable boundary layers' formation and persistency over higher latitude areas. It also has the benefit of gaining insight into the interaction between the atmosphere and land surface processes for stable and very stable boundary layer conditions during the snow season.

### 3. MESOSCALE VARIABILITY OF THE STABLE MARINE BOUNDARY LAYER (SMBL)

Mesoscale variability of EM duct characteristics over the Southern California Bight induced by interaction between the stable marine boundary layer (SMBL) and coastal terrain have been examined using CASPER-West observations and high-resolution air-sea coupled COAMPS simulations. Findings from this research can be found in the following peer-reviewed paper:

**Jiang, Q., Q., Wang, and S. Gaberšek, 2022:** Mesoscale variability of surface-based ducts during Santa Ana winds. *Journal of Geophysical Research: Atmospheres*. 127, <https://doi.org/10.1029/2022JD036698>.

#### ABSTRACT:

Mesoscale variability of the marine boundary layer characteristics and the corresponding radio wave ducting conditions over the Southern California Bight (SCB) during the Santa Ana wind (SAW) events have been examined based on analysis of field observations and diagnosis of mesoscale model simulations. Under the influence of a dry and the stably stratified air flow from inland, a shallow moist internal boundary layer (MIBL) develops over the coastal waters. The MIBL consists of a thin well-mixed moist layer capped by an inversion layer where the specific humidity decreases with height sharply. Accordingly, a surface duct (SD) is virtually ubiquitous with its depth comparable with the MIBL depth. The wind direction and the SD characteristics exhibit diurnal variations over the SCB area associated with land and sea breezes. In addition, the MIBL interacts with the coastal topography, Channel Islands, and sea surface temperature (SST) gradients, resulting in spatial mesoscale variabilities in SD characteristics. Specifically, the SD height and strength grow with the offshore distance owing to the surface latent heat flux over sea, and, in the meantime, exhibit substantial along-shore variation. The shallow MIBL tends to split and flow around the higher elevation of the islands, leading to the formation of wakes with reduced modified refractivity. These island wakes may extend dozens of kilometers downstream with the SD height and strength growing accordingly.

## 4. INTERNAL BOUNDARY LAYER OVER COASTAL WATERS

It is known that the atmospheric boundary layer is much more complicated over a heterogeneous surface such as over a littoral zone or across a sea surface temperature front than over a homogeneous surface. An internal boundary layer (IBL) may develop and the classical Monin-Obukhov similarity theory (MOST) may become invalid. However, MOST still serves as the foundation for many Navy related applications, such as surface layer parameterizations in NWP models Navy Vertical Surface Layer model (NAVSLaM) used for predicting EM ducts. The objective of this research is to a) explore methods for simulating IBLs over coastal waters using a large eddy simulation model, b) advance our understanding of internal boundary layer characteristics and physics, and c) develop a new scaling theory that better characterizes the IBL and improves Navy's battlespace environment forecasting capabilities over strategically important littoral zones. In section 4.1, two novel LES modeling strategies developed over this project have been described. Sections 4.2 and 4.3 include abstracts and references of a pair of peer-reviewed journal articles with emphasis on convective and stable internal boundary layers, respectively. A new scaling theory has been proposed in Section 4.4

### 4.1 Large eddy simulations of internal boundary layers over littoral zones

Large eddy simulation (LES) models with periodic laterally boundary conditions have been widely used for studying the atmospheric boundary layer (BL) over a homogeneous surface in a relatively small domain. The periodic lateral boundary conditions employed by large eddy simulation models become inadequate over coastal areas, across sea surface temperature fronts, or over other heterogeneous surfaces. A couple of modeling methods, namely an Eulerian method using a fringe upstream condition (EFBC) and a Lagrangian method with differential advection correction (LDAC) have been explored and tested over the past few years with encouraging results.

#### 4.1.1 Eulerian simulations with fringe upstream boundary condition (EFBC)

Recently, LES model had been used to examine the marine BL over a heterogeneous surface utilizing a fringe method to break the periodic lateral boundary conditions (e.g., Inoue et al. 2014; Sullivan et al. 2020). In order to investigate the atmospheric IBL over coastal waters, we have implemented the fringe method into the NCAR/NRL-LES code and applied it to the offshore internal boundary layer problem. The EFBC approach is described in the schematic figure below.

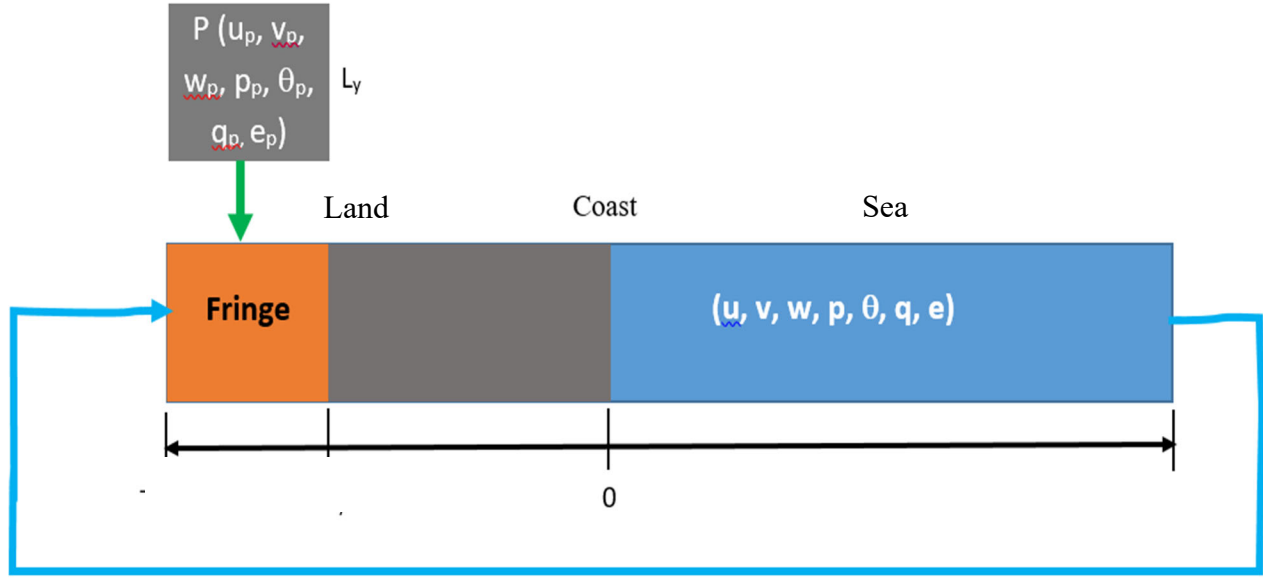


Figure 4.1. Sketch of the Eulerian/Fringe domains, namely a smaller periodic domain P and a long heterogeneous domain H. The domain sizes are  $(L_y, L_y, L_z)$  for P and  $(ML_y, L_y, L_z)$  for H with  $M=15-25$  has been tested. The H domain includes both the land and sea (in blue) portions. In the fringe zone (i.e., in orange in H, over the land surface), each state variable in H is nudged towards its counterpart from domain P at each time step following equation (1) to break the periodic boundary conditions in H.

First, a stable boundary layer over land is simulated using a periodic domain P until an equilibrium stable boundary layer is established. Next, the 3-D instantaneous state variables from domain P are used in the “fringe” zone of domain H to nudge the solutions in H toward their counterparts in P:

$$\frac{\partial \phi}{\partial t} = R_\phi + (\phi_p - \phi) / \tau(x, z) \quad (4.1)$$

where  $\phi = u, v, w, p, \theta, q$ , and  $e$ , denote the wind components, perturbation pressure, potential temperature, water vapor mixing ratio, and subgrid-scale turbulence kinetic energy, respectively, and  $\phi_p = u_p, v_p, w_p, p_p, \theta_p, q_p$ , and  $e_p$  denote their counterparts from domain P. Here  $R_\phi$  denotes all the right-hand-side terms in the LES governing equations and  $\tau(x, z)$  is the nudging time given by

$$\tau^{-1} = 2 \exp[-(x - x_c)^{2n} / a^{2n}] f(z) \quad (4.2)$$

where  $x_c$  is the center of the fringe zone,  $a = L_y / 3$  is an adjustable length scale, and  $n=2$  is constant. Here,  $f(z)=1$  for  $z \leq z_{ref}$  and increases with  $z$  as  $(z / z_{ref})^3$  aloft, where  $z_{ref} = 600$  m is a reference height.

The example reported here is a 5-h LES simulation of a SIBL inside a 30.5 km long domain, which includes a 7.75 km-long portion over a warm and rough land surface (i.e., 2 K warmer than the air and the roughness length  $z_0 = 0.1$  m) and a 22.75 km portion over a cooler and smoother sea surface (i.e., 1 K cooler and  $z_0 = 0.0002$  m). There are  $2048 \times 128 \times 220 \sim 6 \times 10^8$  grid points with a relatively coarse resolution ( $\Delta x = 15$  m). The flow is driven by uniform geostrophic winds  $U_g = 5$  m/s. The surface fluxes and friction are calculated using the bulk formula following Fairall et al. (2003). A fringe method is applied to the leftmost 2 km of the land domain (Fig. 4.2). As expected, the friction velocity is much larger over land than over sea (Fig. 4.2a) with substantially larger perturbations over land. The heat flux is positive over land and

negative over sea (Fig. 4.2b). The negative heat flux tends to decrease with the offshore distance, presumably due to the decrease of the air-sea temperature difference.

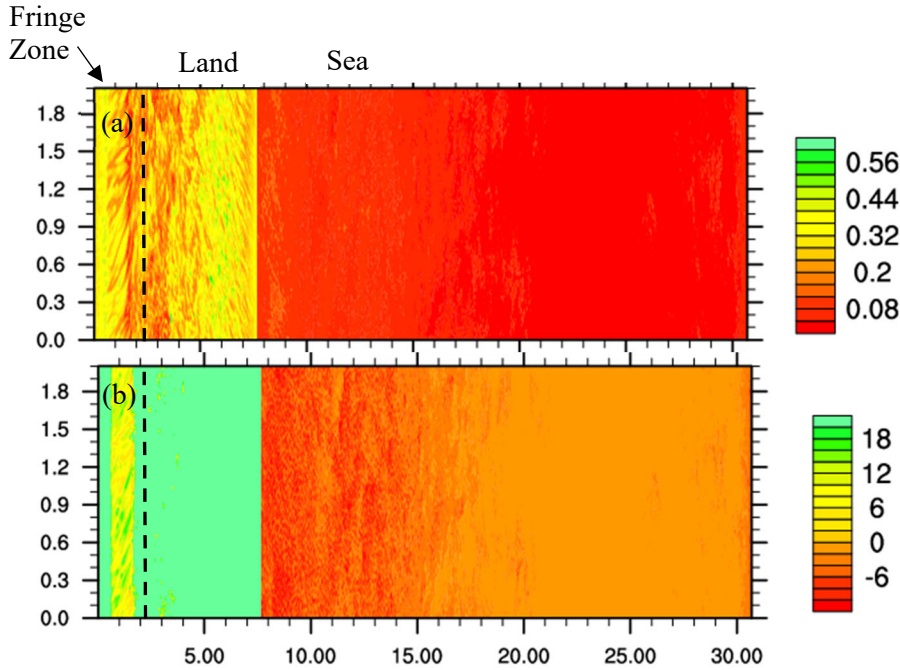


Figure 4.2. Snapshots of (a) the surface friction velocity (m/s) and (b) surface fluxes ( $\text{W/m}^2$ ) in the LES domain valid at 5 h. The domain size is approximately  $L_x \times L_y = 30.5 \times 2 \text{ km}^2$ . The geostrophic wind component is oriented along the x direction (from left to right). The coastline is located at  $X=7.75 \text{ km}$  with land to the left and sea to the right. Over “land”, the surface temperature is  $T=T_0+2 \text{ K}$  where  $T_0$  is the initial air temperature near the surface and the roughness length is  $z_0 = 0.1 \text{ m}$ . Over sea, the SST is  $T=T_0-1 \text{ K}$  and  $z_0 = 0.0002 \text{ m}$ . Over the “Fringe Zone”, the LES solution is being “nudged” toward a convective boundary layer solution derived from a separate LES simulation (Sullivan et al. 2020).

As an example, the vertical cross-section snapshots of the latitudinal winds, potential temperature, specific humidity and modified refractivity are shown in Fig. 4.3. The new fringe approach is able to minimize the influence from periodic boundary conditions and maintain a drier well-mixed layer over land. In addition, the simulation successfully captures the development of a stratified and moist marine internal boundary layer (SIBL). The SIBL here is evidenced by weaker winds, cooler and more stable air, and markedly increased specific humidity and modified refractivity. The growth of the SIBL depth with offshore distance approximately follows the reference curve,  $z_i = 0.33\sqrt{X_{\text{offshore}}}$  (Jiang and Wang 2021).

(a)

(b)

(c)

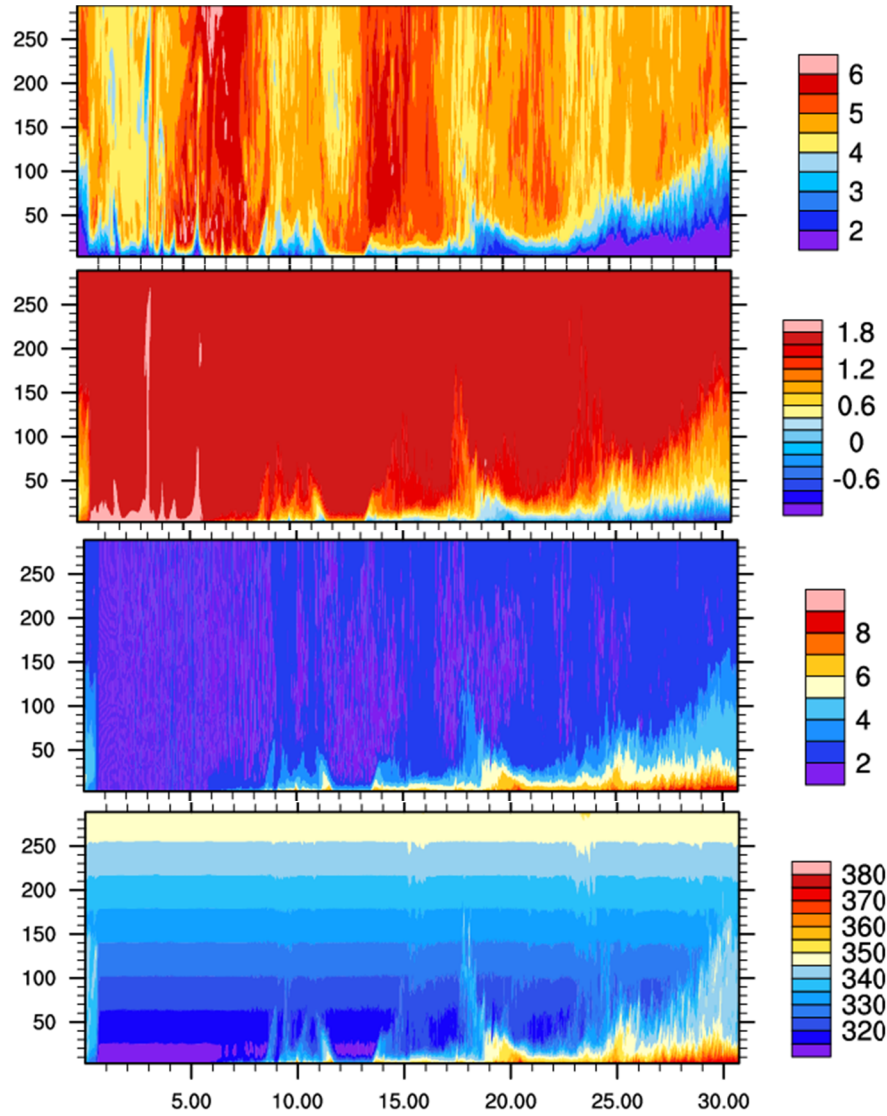


Figure 4.3. The vertical cross-section snap shots of (a) the latitudinal wind component ( $u$ , incr. = 0.5 m/s), (b) potential temperature ( $\theta - \theta_0$ , incr. = 0.2 K), (c) specific humidity ( $q$ , incr. = 1g/kg), and (d) modified refractivity ( $M$ , M units) from the large eddy simulation valid at  $T = 5$  h. Only the lowest 280 m is shown. The geostrophic winds blow from left to right. The stable internal boundary layer depth from Jiang and Wang (2021) is included as white dashed curves for comparison.

The latitudinal winds above the SIBL exhibit some notable quasi-periodic variability, likely due to gravity waves with wavelengths between 5-10km. It is worth noting, though, that such gravity wave response to an abrupt change in surface properties is physical rather than numerical. The potential influence of these waves on the scaling of the nearshore marine surface layer deserves to be further examined.

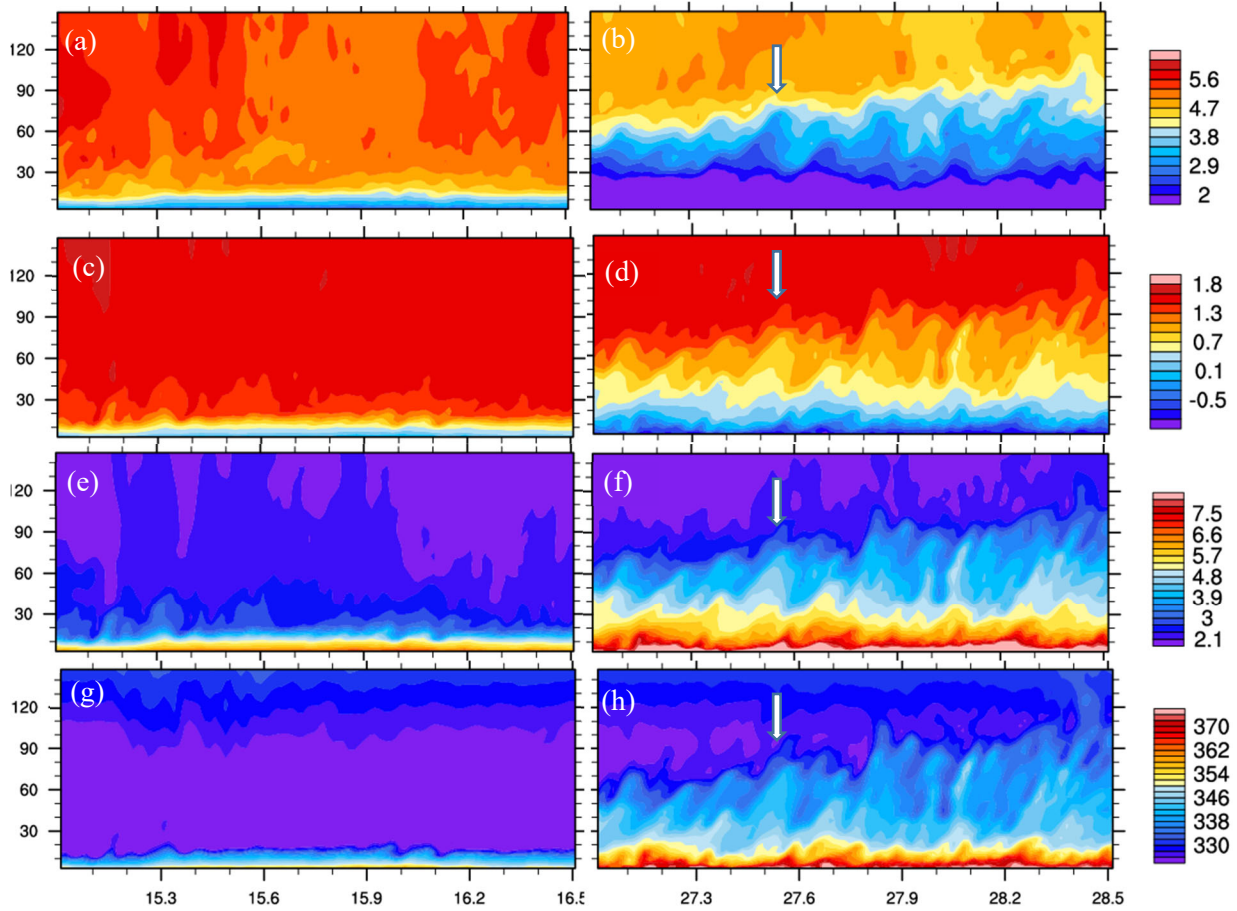


Figure 4.4. Snapshots of vertical cross-sections of (a, b) the meridional wind component (incr. = 0.3 m/s), (c, d) potential temperature ( $\theta - \theta_0$ , incr. = 0.2 K), (e, f) specific humidity ( $q$ , incr. = 0.3 g/kg), and (g, h) modified refractivity ( $M$ , incr. = 3 units) valid at  $T = 5$  h. The left and right columns correspond to 1.5 km-long segments of the domain centered at  $X = 15.75$  and  $27.75$  km (or offshore distance of 8 and 20 km), respectively. Only the lowest 150 m is shown.

It is interesting to compare the SIBL and turbulent characteristics as well as duct conditions at different offshore locations (Fig. 4.4). At the nearshore location (left columns), the SIBL is shallow and turbulent eddies are smaller. Larger eddies above the SIBL, presumably advected from land, are notable in the wind field and indistinguishable in the well-mixed scalar fields. The surface layer with larger refractivity is about 10-15 m deep, above which  $M$  becomes smaller. In the right column (i.e.,  $\sim 20$  km offshore), the SIBL is much deeper and characterized by larger moist eddies. A few aspects of these large eddies are worth noting. First, these large eddies have horizontal scales between 100-200 m, which are typical for a stable internal boundary layer and much smaller than those in a convective boundary layer. Second, unlike the moist plumes in a convective boundary layer (e.g., Franklin et al. 2022), the large eddies in the SIBL tend to attach to the surface, likely due to the negative buoyancy effect. Therefore, the classical attached eddy hypothesis (e.g., Townsend 1976) applies. Third, majority large eddies tilt forward, in consistence with generally positive wind shear in the SIBL. Furthermore, coherent perturbations associated with large eddies appear in all the fields and accordingly induce substantial changes in the modified refractivity (e.g., the eddy highlighted by the white arrows in Fig. 4.4).

#### 4.1.2 Lagrangian simulations with novel differential advection correction (LDAC)

Lagrangian modeling method is computationally efficient and allows us to simulate the boundary layer adjustment to spatially-evolving forcings over a substantial distance with a relatively high model resolution. However, an intrinsic weakness of a traditional Lagrangian approach is that differential advection associated with vertical wind shear in the Lagrangian column is ignored, an approximation that becomes problematic near the sea surface where wind shear is typically strong. The LDAC method was first proposed in Jiang and Wang (2021) to simulate the turbulent marine boundary layer over coastal waters under the influence of prevailing offshore winds. According to Jiang and Wang (2021), the prognostic equation for a momentum component (i.e.,  $u$ ,  $v$ , and  $w$ ) or scalar (i.e., potential temperature, water vapor, and turbulence kinetic energy),  $C$ , in a Lagrangian air column moving at a speed  $U_m$  can be written as,

$$\frac{\partial C}{\partial(\lambda t)} = -U_m \frac{\partial C}{\partial x} + R.H.S., \quad (4.3)$$

where  $\lambda = U_m / U_R$  was referred to as the “shear factor”. Here  $U_R$  denotes a reference wind speed in the prevailing wind direction. The shear factor is larger than unity near the surface where the wind is weaker and tends to unity at the boundary layer top. The two terms on the right hand side represents Lagrangian advection and typical right-hand-side terms in the traditional prognostic equations (e.g., flux divergences), respectively. In the classical Lagrangian approach,  $\lambda = 1$  in the whole air column, and the differential advection associated with vertical wind shear and horizontal gradients is neglected. A unique aspect of the LDAC approach is that it minimizes the well-known problems associated with the neglect of differential advection in the classical Lagrangian approach. In addition, it is self-contained (no additional information from observations is needed to estimate the differential advection). In Jiang and Wang (2021), the reference wind  $U_R$  was obtained from domain average of the  $u$ -wind component over the coastal area and was computed offline. Accordingly,  $U_R$  is only a function of height and needs to be estimated from a previous simulation.

We have explored a new method for calculating the shear factor  $\lambda$ , namely,  $\lambda = U_m / \langle U \rangle$ , where  $\langle U \rangle$  denotes the domain average of the  $u$ -wind component at each time step, which represents the true mean wind shear at each location. This approach (referred to as inline method hereafter) allows us to accurately assess the differential advection term. A qualitative comparison between two LES simulations with identical model configuration except for the method to obtain  $U_R$  are shown in Fig. 1. In general, the simulated potential temperature and TKE sections are comparable; both showing a sudden decrease of TKE offshore and the growth of a stable internal boundary layer due to surface cooling over sea. In addition, the simulation with the new in-line method exhibits more structure than its offline counterpart. A group of four test simulations were conducted using this new approach. We found that a) the new formulation has little impact on the numerical stability of the LES code, b) the additional cost for computing  $\langle U \rangle$  is negligible, c) the simulated fields are consistent with previous simulations using  $U_R$  estimated offline.

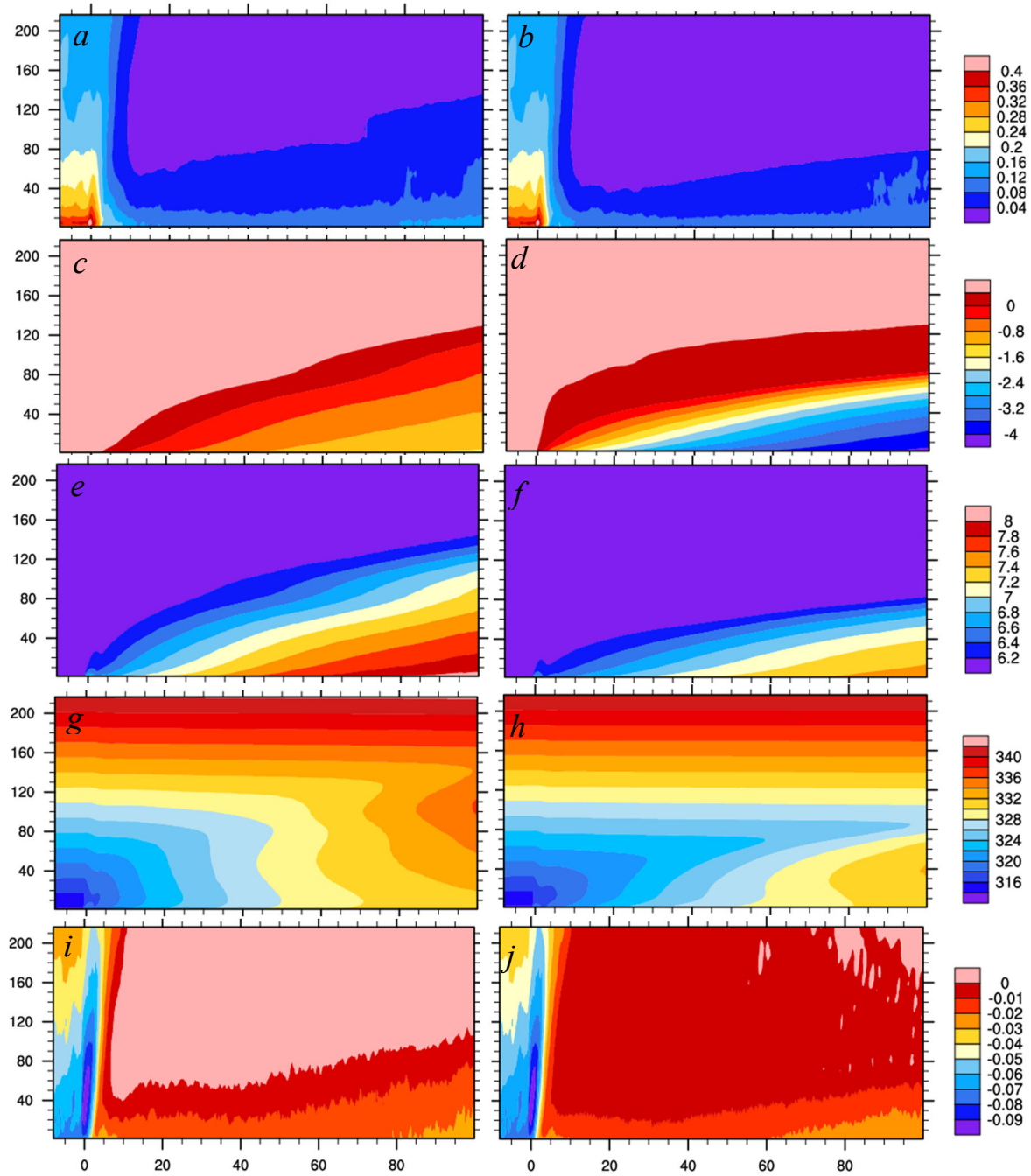


Figure 4.5. Vertical cross-sections of (a, b) TKE ( $\text{m}^2 \text{s}^{-2}$ ), (c, d) potential temperature ( $\theta - \theta_0$ , K), (e, f) specific humidity ( $q$ , g/kg), (g, h) modified refractivity (M units), and (i, j) zonal momentum flux ( $\overline{u'w'}$ ,  $\text{m}^2 \text{s}^{-2}$ ), constructed from two nearly identical LES simulations except that the land-sea temperature differences are 6 (left column) and 10 K (right column), respectively.

We have conducted three additional simulations using the new inline method to explore the sensitivity of the internal boundary layer characteristics and scaling to the surface cooling over sea. These

are idealized simulations driven by geostrophically-balanced offshore winds,  $U_g = 9$  m/s. The land and sea surfaces are defined by different roughness lengths and surface temperatures. Surface fluxes are calculated using bulk flux formula (e.g., Fairall et al. 2003). The land surface is 2 K warmer than the initial air temperature near the surface. Accordingly, a weakly convective boundary layer develops over the land. These simulations are virtually identical except that the sea surface temperature varies from simulation to simulation. Preliminary analysis of these simulations find that the general characteristics of the SIBL and the associated duct conditions are moderately sensitive to the sea surface temperature over the parameters examined. Shown in Fig. 4.5 are vertical cross-sections from two such simulations with a land-sea surface temperature difference of 6 K (left) and 10 K (right column), respectively.

As expected, the boundary layer structure is identical over land. Immediately offshore, TKE nearly collapses (Fig. 4.5a and 4.5b) with an SIBL developing further offshore in both simulations. The SIBL are evident in the TKE, potential temperature (i.e., cooler), specific humidity (i.e., more moist), and momentum flux cross-sections. While the SIBL in these simulations are qualitatively similar, the SIBL over a cooler sea surface (right column) tends to be shallower and less turbulent. In addition, the air in the SIBL over a cooler sea tends to be drier. Consequently, the modified refractivity patterns and ducting conditions in the two simulations are markedly different (Fig. 4.5-h). Further analysis is underway focusing on the scaling of mean profiles and higher-order moments.

One of the objectives of this project is to evaluate the influence of mesoscale forcing on the structure of coastal marine boundary layers. It has been demonstrated by Jiang et al. (2022) that ducting conditions in a stable marine boundary layer exhibited substantial spatial variability over the southern California coastal area owing to strong mesoscale forcing. Mesoscale forcing may induce strong convergence or divergence in the marine boundary layer and how the SIBL responds to such forcing remains poorly understood. We conducted a pair of large eddy simulations, which are identical except that a specified subsidence associated with divergence in the boundary layer is applied to one of them. The divergence is  $-3 \times 10^{-5} \text{ s}^{-1}$  and constant with height in the lowest 500 m.

A qualitative comparison between the two simulations are shown in Fig. 4.6. The vertical cross-sections from these two simulations are quite similar. A major difference between the two is the SIBL depth. With the imposed subsidence (right column), the SIBL is noticeable shallower. For example, at 60 km offshore, the SIBL defined based on the momentum flux (see white-dashed lines in Fig. 4.6g-h) is about 20% shallower with imposed subsidence than without. This implies that the constant flux layer could be much shallower under the influence of mesoscale subsidence. It is worth noting that the magnitude of the divergence used in this example is relatively small for typical mesoscale processes.

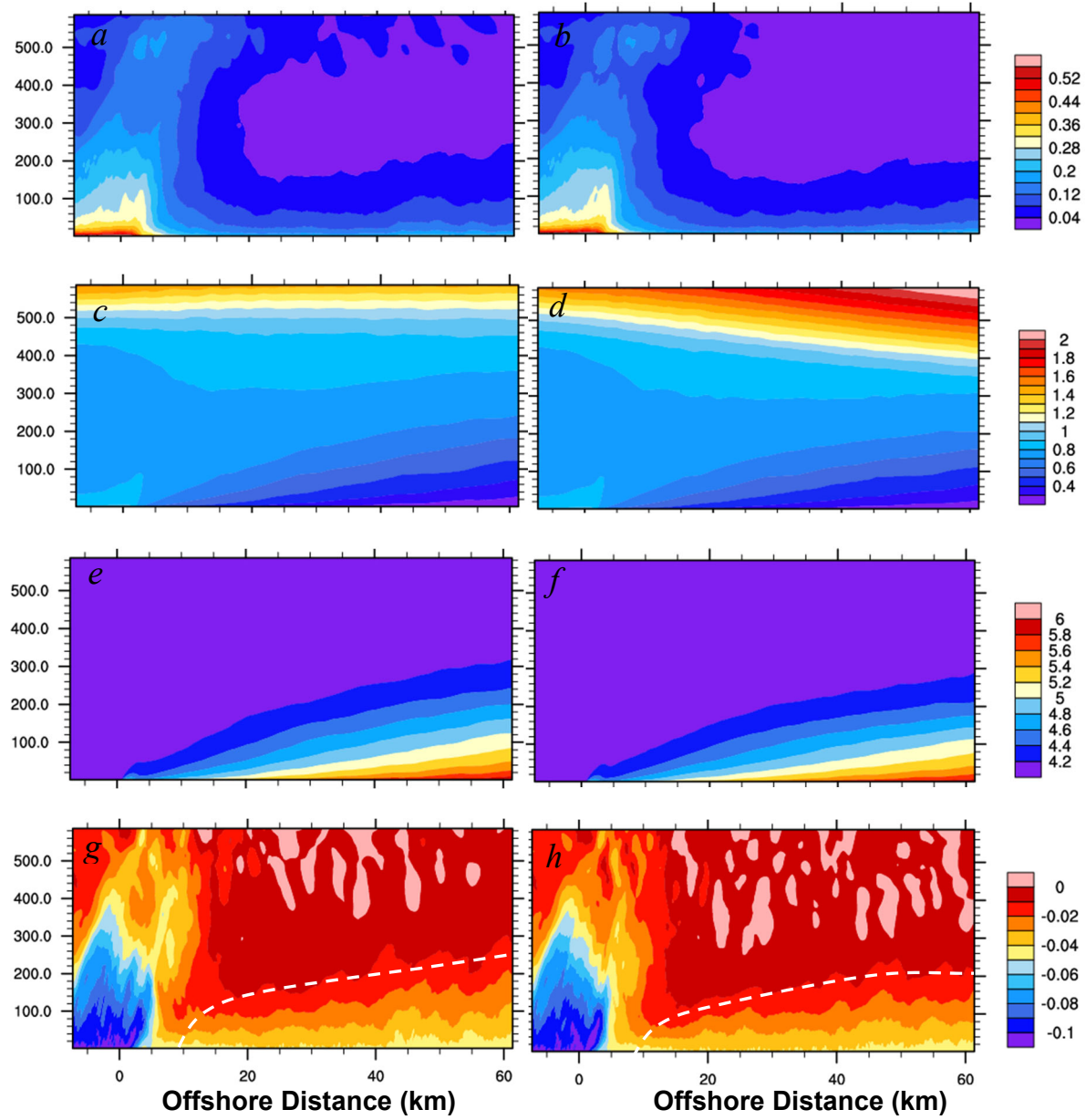


Figure 4.6. Vertical cross-sections of (a, b) TKE ( $\text{m}^2 \text{s}^{-2}$ ), (c, d) potential temperature ( $\theta - \theta_0$ , K), (e, f) specific humidity ( $q$ , g/kg), and (g, h) zonal momentum flux ( $\overline{u'w'}$ ,  $\text{m}^2 \text{s}^{-2}$ ), constructed from two nearly identical LES simulations except that a constant divergence ( $-3 \times 10^{-5} \text{s}^{-1}$ ) is applied to the one corresponding to (b, d, f, and h) to mimic subsidence induced by mesoscale forcing. The white-dashed curves in g and h indicate the approximate top of the SIBL.

## 4.2 Convective internal boundary layer (CIBL) over coastal waters

The development and characteristics of convective internal boundary layers has been examined using the CASPER-East observations, air-sea coupled COAMPS simulations, and large eddy simulations. Major findings have been reported in the following peer-reviewed journal paper:

**Jiang, Q.**, Q. Wang, S. Wang, and S. Gaberšek, 2020: Turbulence adjustment and scaling in an offshore convective internal boundary layer: A CASPER Case Study. *J. Atmos. Sci.*, 77, 1661-1681.

#### ABSTRACT

The characteristics of a convective internal boundary layer (CIBL) documented offshore of the US East Coast during the field campaign CASPER-EAST has been examined using field observations, a coupled mesoscale model (i.e., Navy's COAMPS) simulation, and a couple of surface-layer resolving large eddy simulations (LES). The Lagrangian modeling approach has been adopted with the LES domain being advected from a cool and rough land surface to a warmer and smoother sea surface by the mean offshore winds in the CIBL. The surface fluxes from the LES control simulation are in reasonable agreement with field observations, and the general CIBL characteristics are consistent with previous studies.

According to the LES simulations, in the nearshore adjustment zone (i.e., fetch < 8 km), the low-level winds and surface friction velocity increase rapidly, and the mean wind profile and vertical velocity skewness in the surface layer deviates substantially from the Monin-Obukhov similarity theory (MOST) scaling. Further offshore, the nondimensional vertical wind shear and scalar gradients and higher-order moments are consistent with the MOST scaling. An elevated turbulent layer is present immediately below the CIBL top, associated with the vertical wind shear across the CIBL top inversion. Episodic shear instability events occur with a timescale between 10-30 min., leading to the formation of elevated episodic maxima in turbulence kinetic energy and momentum fluxes. During these events, the turbulence kinetic energy production exceeds the dissipation, suggesting that the CIBL remains in non-equilibrium.

### 4.3 Characteristics and Scaling of the Stable Marine Internal Boundary Layer

A stable internal boundary layer (SIBL) offshore of the North Carolina has been studied using CASPER-East observations, air-sea coupled COAMPS simulations, and large eddy simulations using the LDAC method. Major findings have been reported in the following peer-reviewed journal paper

**Jiang, Q.** and Q. Wang, 2021: Characteristics and scaling of the stable marine internal boundary layer. *JGR: Atmosphere*, 126. <https://doi.org/10.1029/2022JD036698>.

## ABSTRACT

The characteristics of a stable internal boundary layer (SIBL) offshore of Duck, North Carolina documented during CASPER-EAST have been examined using surface layer-resolving large eddy simulations. Under the influence of steady offshore winds, a SIBL develops over the coastal waters with its depth growing with the offshore advection time ( $T_A$ ) as  $\sim \sqrt{T_A}$ . The SIBL includes an internal equilibrium layer above the surface where the heat and momentum fluxes linearly decrease with height and a transition layer aloft. The offshore area can be partitioned into three distinct zones, namely, a nearshore zone (NSZ,  $T_A < T_d/3$ ), an intermediate offshore zone (IOZ,  $T_d/3 < T_A < T_d$ ), and a far-offshore zone (FOZ,  $T_A > T_d$ ), where  $T_d$  is the vertically-averaged turbulent dissipation time. The advection of turbulence from land plays a significant role in the momentum and scalar budgets over the NSZ. Consequently, the Monin-Obukhov similarity theory (MOST) becomes invalid, and local similarity theory (LST) is applicable only above the SIBL in the NSZ. In the IOZ, the advection of turbulence from land is less important but still non-negligible. As a result, while the dimensionless standard deviations of winds and potential temperature converge toward the MOST curves, the dimensionless gradients are smaller than their counterparts in the FOZ. In the FOZ, the advection from land becomes negligible, and MOST is applicable throughout the constant flux layer. Over the IOZ and the FOZ, LST works reasonably well within the SIBL.

## 4.4 A new theory for SIBL over coastal waters

A new physics-based scaling laws for the marine boundary boundary layer over coastal waters under different meteorological conditions has been proposed and extensively tested with multiple high-resolution LDAC large eddy simulations.

### 4.4.1 Issues related to applying MOST over coastal waters

In order to characterize offshore stable internal boundary layers, test the validity of classical MOST and develop new scaling laws, several Lagrangian large eddy simulations have been conducted over a range of meteorological parameters (i.e., wind speed, air-sea temperature difference, etc.). These simulations capture most of the salient features of the offshore stable internal boundary layer, such as the square-root growth of the SIBL depth with the offshore advection time (or fetch). A few examples are shown here, including two new simulations with different geostrophic winds (Table 4.1).

Table 4.1. Parameters from a group of three surface layer resolving Lagrangian large eddy simulations of the offshore stable internal boundary layer. The parameters are air-land temperature difference ( $\Delta T_{A-L}$ , K), air-sea temperature difference ( $\Delta T_{A-S}$ , K), geostrophic wind speed, the minimum dynamic heat flux offshore, and the minimum Monin-Obukhov length ( $L_{\min}$ , m).

| Exp. | $\Delta T_{A-L}$ (K) | $\Delta T_{A-S}$ (K) | $U_g$ (m/s) | $H_{smin}$ | $L_{\min}$ (m) |
|------|----------------------|----------------------|-------------|------------|----------------|
| A    | 3                    | -3                   | 6           | -0.0452    | 11.8           |
| B    | 1                    | -3                   | 6           | -0.0314    | 17.0           |
| C    | 3                    | -3                   | 3           | -0.0165    | 3.2            |

Several theoretical and practical issues arise when applying MOST to an offshore stable internal boundary layer, as shown in the examples below (Figs. 4.7).

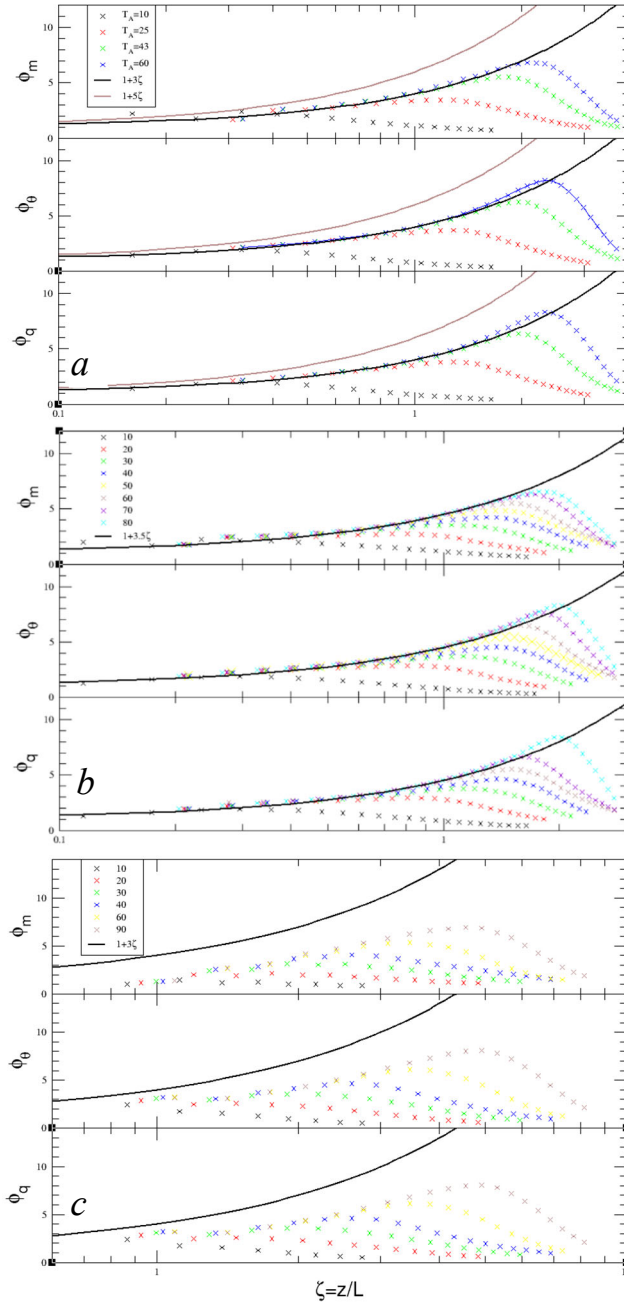


Figure 4.7. Dimensionless vertical gradients of horizontal winds (top), potential temperature (middle), and specific humidity (bottom) versus the stability parameter ( $\zeta = z/L$ , where  $L$  is the Monin-Obukhov length) derived from simulations a) A, b) B, and c) C are shown, with data points in different colors corresponding to different offshore advection times ( $T_A$ ). Only the portion inside the SIBL is shown. The two reference curves in a) correspond to  $1+3\zeta$  and  $1+5\zeta$ , respectively, and only the former is plotted in b) and c).

According to Figure 4.7, in general, the dimensionless wind speed and scalar gradients over coastal waters are smaller than the widely used profile-flux relation over land (i.e.,  $1+5\zeta$ ). For simulations A and B the

dimensionless gradients near the surface tend to follow the  $1+3\zeta$  curve up to approximately  $z = 0.2-0.3 h$ , where  $h$  is the SIBL depth. At each offshore location (i.e., a given offshore advection time  $T_A$ ), the depth of the layer with gradients converging toward  $1+3\zeta$  increases with  $T_A$ , in proportion to the offshore increase of the SIBL depth. Further aloft, the nondimensional gradients drop rapidly with the vertical distance above the water surface. Finally, the nondimensional gradients seem to be smaller in simulation C, which has the smallest Monin-Obukhov length (Table 4.1) over the nearshore zone.

The rapid falloff the dimensionless gradients aloft is likely due to the fast decrease of the turbulent fluxes with height inside the SIBL, which violates one of MOST's key assumptions, the constant flux assumption. As demonstrated in Jiang and Wang (2021), the momentum and heat fluxes in a SIBL approximately linearly decrease with height, i.e.,  $F(z) = F_0(1-z/h)$ , where  $F_0$  denotes the surface momentum or scalar flux. For example, if the SIBL is 40 m deep, by definition, the constant flux layer is around 4 m, which is too shallow for many practical applications. The scaling challenge associated with the rapid decrease of turbulent fluxes with height in a shallow stable boundary layer has been addressed in several studies in the past four decades. A widely tested remedy to MOST was referred to as the local scaling theory (LST, e.g., Grachev et al. 2021), which uses local fluxes (i.e., fluxes at level  $z$ ) instead of surface fluxes for scaling. Examples of LST scaling are shown in Fig. 4.8.

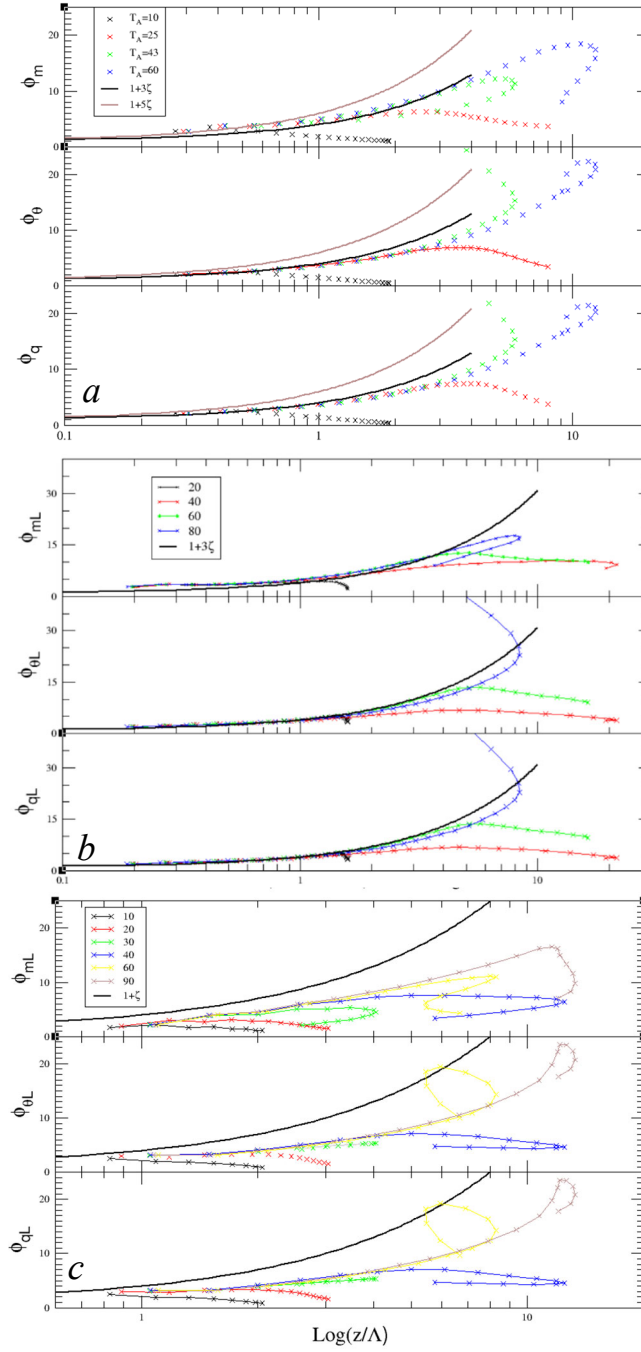


Figure 4.8. Same as Fig. 4.7 except that the gradients are non-dimensionalized using local fluxes, the Obukhov length ( $L$ ) is replaced with its local fluxes-based counterpart ( $\Lambda$ ), and the horizontal axis is shown in logarithm.

Figure 4.8 suggests that the gradients non-dimensionalized using local fluxes converge better than the traditional MOST over a layer approximately between  $z=0$  and  $0.4h$ . For the upper portion of the SIBL (i.e., approximately  $z > 0.4h$ ), the dimensionless gradients exhibit large scatter presumably due to the small flux amplitudes near in the upper portion of the SIBL, which was referred to as a transition layer in Jiang and Wang (2021). It is worth noting that it is rather inconvenient to use LST for applications such as predicting modified refractivity profiles, as turbulent fluxes at multiple levels need to be provided.

#### 4.4.2 A new scaling theory for SIBL

Based on multiple LES simulations and theoretical considerations, we have proposed a set of new flux-gradient relationships proposed for SIBL over coastal waters. The new flux-gradient relations can be written as

$$\frac{\kappa z}{u_*} \frac{\partial \bar{U}}{\partial z} = \varphi_M (1 - z/h) \quad \text{and} \quad (4.4)$$

$$\frac{\kappa z}{\theta_*} \frac{\partial \bar{\theta}}{\partial z} = \varphi_H (1 - z/h) \quad (4.5)$$

for winds and scalars (i.e., temperature, specific humidity, etc.), respectively. Several aspects of the new scaling law are worth noting:

- 1) Equations (4.4) and (4.5) are identical to the classical MOST except for the additional correction factor  $(1-z/h)$  on the right hand sides.
- 2) The correction factor represents the effect of the decrease of turbulent fluxes with height. The constant flux assumption in MOST is replaced with the linear decrease of fluxes with height assumption. The linear decrease of turbulent fluxes assumption is supported by large eddy simulation results and limited field observations (e.g., Tjernström and Smedman 1993) and accordingly, the new scaling may be applicable to beyond the traditional surface layer.
- 3) Near the surface, i.e.,  $z/h \ll 1$ , equations (4.4) and (4.5) reduce to the classical MOST flux-gradient relations. The correction factor only becomes significant in the upper portion of the SIBL.
- 4) The new scaling relations can also be derived from the local scaling theory by assuming the linear decrease of fluxes with height and slow variation of the local Obukhov length with height.

The new scaling with  $\varphi_m = \varphi_H = 1 + 5\zeta$  are included in Figure 4.9 as solid curves along with the  $1 + 3\zeta$  curve to compare with the dimensionless gradients at several different offshore advection times from the three LES simulations. In general, the new scaling law performs better than the traditional MOST, especially for the upper portion of the SIBL.

In summary, the new scaling law is based on the assumption that turbulent fluxes in a relatively shallow stable internal boundary layer tend to decrease linearly with the vertical distance above the water surface, an assumption that is supported by large eddy simulations and field observations. In (4.4-4.5), the correction factor can be written into a more general form,  $(1 - z/h)^\alpha$ , where the constant  $\alpha$  determines how fast turbulent fluxes decrease with height and is between 0.8-1.75 based on previous field observations. In the near surface and deep boundary layer limits, the correction factor becomes unity and equations (4.4-4.5) reduce to the classical MOST. Therefore, the proposed scaling is physically sound and is superior to MOST for a shallow stable boundary layer such as a SIBL.

#### 4.4.3 Implications of the new scaling law and future work

According to Jiang and Wang (2021), a stable internal boundary layer can be partitioned into the lower equilibrium layer where LST is applicable and the upper transition layer where LST becomes invalid. The results shown in Fig. 4.9 suggest that equations (1-2) can be used for

predicting wind and scalar profiles up to the top of the SIBL. These new findings have important implications on the surface parameterizations for Navy's numerical weather prediction model and EM duct predictions over littoral zones.

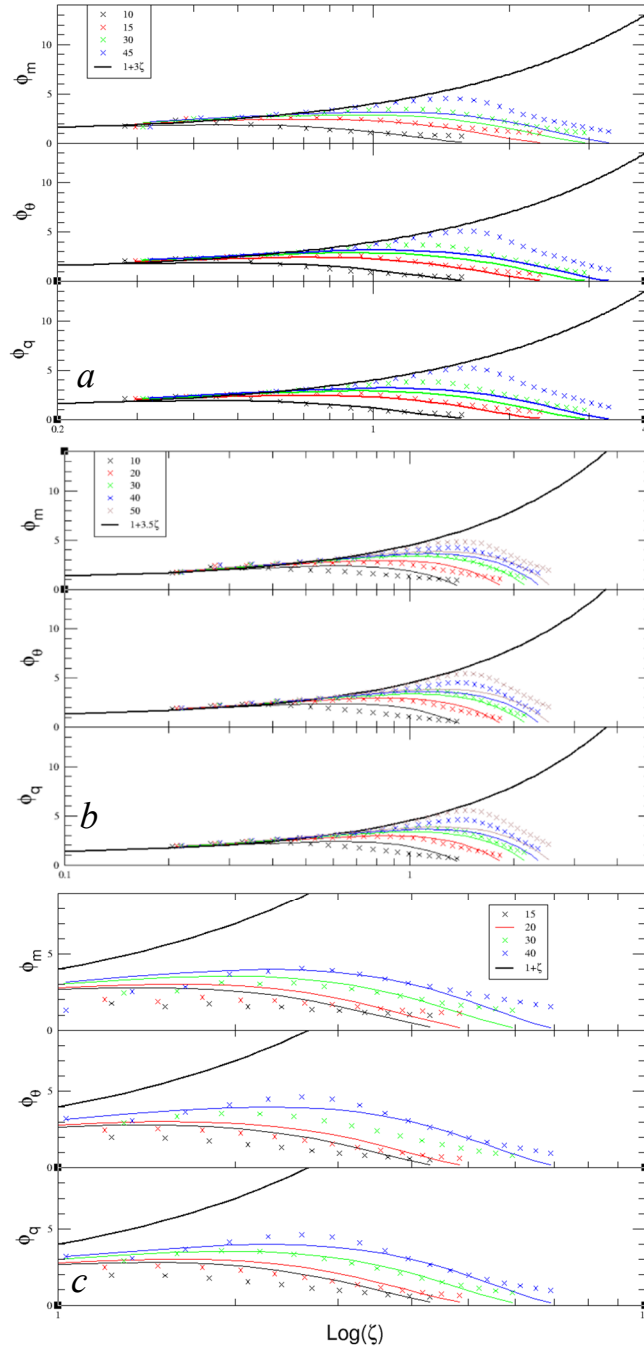


Figure 4.9. Similar to Fig. 4.7 except that results from equations (4.4) and (4.5) are plotted as solid lines for different offshore advection times.

Whether MOST is applicable to the surface layer over coastal waters remains an open question. According to our study, a better question seems to be whether a sizable surface layer exists in a SIBL so that MOST can be useful for practical applications.

For example, for a SIBL with its depth  $h = 40$  m, a constant flux layer may exist with its depth  $z_{sfc} = 0.1h = 4$  m. Such a  $z_{sfc}$  is far below the first model level (i.e., 10 m) in a typical numerical weather prediction (NWP) model. For surface layer parameterizations in COAMPS, MOST is assumed to be valid up to model level and, according to equations (4.4-4.5), the surface fluxes evaluated at  $z = 10$  m based on MOST may produce  $\sim 40\%$  positive biases over coastal waters under stable conditions. For a 200-m-deep boundary layer, the biases reduce to  $\sim 5\%$ , which is less an issue for NWP forecasts. Therefore, equations (4.4-4.5) can be readily incorporated into the COAMPS surface layer parametrization (Wang et al. 2002) and only makes noticeable differences in a relatively shallow stable internal boundary layer.

The Navy Vertical Surface Layer model (NAVSLaM) is used for predicting EM ducts by employing MOST up to 60 m. Similarly, EM ducts are diagnosed from COAMPS by blending MOST predictions in the lowest 40 m with COAMPS forecasts aloft. Apparently, these approaches become problematic when applying to a stable internal boundary layer. This study suggests a blending between surface layer predictions based on (4.4-4.5) up to the SIBL depth and COAMPS forecast aloft is likely a better alternative for the EM duct prediction purpose.

It is worth noting that equations (4.4-4.5) are in qualitative agreement with some available observations, including a couple of stable profiles sampled near the North Carolina coastline during CASPER-East and earlier observations over the Baltic Sea (Termstrom and Smedman 1993). Overall, direct observations of vertical wind and scalar gradients over coastal waters are still rare. The upcoming REDSAW field campaign may provide valuable data for testing some key findings from this research. Specifically, test or validate the following:

- the square-root dependence of the SIBL depth on the offshore distance
- decrease of turbulent fluxes with height and its dependence on SIBL depth (i.e.,  $(1 - z/h)^\alpha$ )
- the new scaling relations (4.4-4.5),
- new surface parameterization based on (4.4-4.5) for Navy's NWP model
- and new blending algorithm over coastal waters based on (4.4-4.5).

## 5. IMPACT OF OCEAN WAVES ON SML STRUCTURE AND EVAPORATION DUCTS

This was the first study of ocean wave impact on the vertical distribution of scalars in the marine surface layer, which is crucial for EM wave ducting in the marine boundary layer and therefore crucial for forecasting over the horizon radio wave propagation. This issue has been explored through a group of well-designed surface layer resolving large eddy simulations using a state-of-science LES model. The findings from this research have implications on air-sea coupling parameterization and EM duct prediction over ocean waves. Some of the results have been reported in the following peer-reviewed journal paper.

**Jiang, Q.**, 2020: Influence of Swell on Marine Surface Layer Structure. *J. Atmos. Sci.*, 77, 1865-1885.

### ABSTRACT

The influence of swell on turbulence and scalar profiles in a marine surface layer and underlying physics is examined in this study through diagnosis of large eddy simulations (LES) that explicitly resolve the surface layer and underlying swell. In general, under stable conditions, the mean wind and scalar profiles can be significantly modified by swell. The influence of swell on wind shear, turbulence structure, scalar profiles and ED characteristics becomes less pronounced in a more convective boundary layer, where the buoyancy production of turbulence is significant. Dynamically, swell has little direct impact on scalar profiles. Instead it modifies the vertical wind shear by exerting pressure drag on the wave boundary layer. The resulting redistribution of vertical wind shear leads to changes in turbulence production and therefore turbulence mixing of scalars.

Over swell, the LES-simulated eddy diffusivities systematically deviate from the Monin-Obukhov similarity theory (MOST) prediction, implying that MOST becomes invalid over a swell-dominated sea. The deviations from MOST are more pronounced in a neutral or stable boundary layer under relatively low winds and less so in a convective boundary layer.

## 6. COMPUTING REFRACTIVE INDEX STRUCTURE PARAMETER CN2 IN A NUMERICAL WEATHER PREDICTION MODEL

Through close collaboration with our NRL colleagues and Professor Qing Wang's group at Naval Postgraduate School, a second-order turbulence closure Cn2 model has been formulated and developed within the framework of the Navy's COAMPS model (i.e., COAMPS-Cn2). COAMPS-Cn2 explicitly predicts water vapor and temperature variances and the covariance, which are crucial for adequately computing Cn2. This predictive approach substantially improves the Cn2 prediction by partially taking into account non-local turbulent mixing processes. Our group has conducted a number of high-resolution LES simulations of the SML over waters, based on which the eddy diffusivity for the variance and covariance mixing has been optimized. The new Cn2 model predictions compare favorably with the observations and additional high-fidelity large eddy simulations. These results have been published in a peer-reviewed journal article and the Cn2 module has been transitioned.

**Wang, S.,** Q. Wang, B. J. Wauer, and **Q. Jiang**, 2020: Computing refractive index structure parameter Cn2 in a numerical weather prediction model. *Geophysical Research Letters*. 47, e2020GL089168.

### ABSTRACT:

The refractive index structure parameter Cn2 provides crucial information regarding turbulent effects on electromagnetic wave propagation. We propose to use second-order turbulence closure techniques to predict Cn2 based on boundary layer parameterizations in numerical weather prediction (NWP) models. This predictive approach provides additional degrees of freedom for improving Cn2 simulations. The model is evaluated using a single column version of Coupled Ocean-Atmosphere Mesoscale Prediction System in a case study of convective boundary layer over the ocean. The predicted Cn2 is generally consistent with results from large-eddy simulations, well-known scaling theories, and limited surface layer observations. This study offers a general approach for computing Cn2 for the entire boundary layer and various turbulence regimes in a NWP model.

## 7. SUMMARY AND CONCLUDING REMARKS

The overarching objective of this project is to identify yet-to-be discovered key fine-scale processes in the SABL, including intermittent turbulence and internal wave generation, turbulent adjustment to a wavy or heterogeneous surface, and their role in regulating the SABL. These science issues are of fundamental scientific importance and great significance to Navy battlespace environment forecasting capability. This research has focused on two broad categories of science issues, namely, a) shear instability, intermittent turbulence and internal gravity wave generation, and SABL-snow coupling, and b) internal boundary layer (IBL) physics and simulations associated with the SABL over a heterogeneous surface with emphasis on IBL over coastal waters and the marine boundary layer over ocean waves. Major achievements of this project are summarized below.

### 7.1 Shear instability, intermittent turbulence, and internal waves in the SABL

Kelvin-Helmholtz instability (KHI) in the stable atmospheric boundary layer (SABL): KHI is considered a major source of intermittent turbulence and internal waves in the SABL, which play a role in invalidating the classical Monin-Obukhov similarity theory (MOST) and pose a challenge to the SABL parameterization in numerical weather prediction (NWP) models. The objective of this effort is to shed light on a) the generation mechanisms of intermittent turbulence and gravity waves in the SABL and their role in regulating the SABL structure, and b) the dynamics of KHI in the close proximity of ground surface in the SABL.

These issues have been addressed using eddy-resolving large eddy simulations, which explicitly resolve gravity waves, KHI billows, secondary billows, and energy-containing turbulent eddies and the interactions among them. Novel and significant findings from this research include:

- Kelvin-Helmholtz instability occurring in an elevated shear layer interacts strongly with the underlying SABL, leading to wavy motions, flow separation from the surface, and intermittent turbulence in the SABL.
- Vertical mixing in the SABL is enhanced directly by elevated KH billows and by interaction between KH billows aloft and SABL turbulence.
- KHI could frequently occur near the ground surface in an Ekman-type SABL associated with the presence of an inflection point in the wind profiles, located in the first several meters or tens of meters above the ground surface under very stable stratification conditions.
- A KHI event in the SABL include the exponential growth of quasi-two-dimensional KH waves (or billows), the onset of secondary shear instability along S-shaped braids, and three-dimensional turbulent breakdown of KH billows.
- Significant vertical transport of momentum, heat, and moisture takes place associated with KHI processes, which weakens vertical wind shear and stratification and induces explosive growth of the SABL depth.
- It has been demonstrated that internal gravity waves and intermittent turbulence are generated during KHI events.

Previous numerical studies focused on KHI in turbulence-free shear layer away from any solid surface. This is the first high-resolution large eddy simulation study that reproduces KHI in a turbulent boundary layer in the close proximity of the ground surface. It addresses some key science issues related to KHI, secondary instability, and generation of intermittent turbulence and internal waves near the

ground surface in the SABL. Insights from this study have the potential to improve stable boundary layer parameterizations in Navy mesoscale weather prediction systems. Some results from this effort have been reported in a peer-reviewed journal paper and a second manuscript based on this research has been submitted.

## **7.2 Interaction between the Arctic SABL and a snow or ice surface**

Accurate numerical weather forecasts over the Arctic region remain a great challenge due to our limited understanding of the stable atmospheric boundary layer physics and its coupling with a snow or ice surface. Particularly, decoupling between the land surface and the SABL under very stable conditions often causes cold biases and induces positive feedback on the surface air temperature. To reduce the forecast errors of NWP models over Polar Regions, a new snow physics package has been added to the Noah Land Surface Model (LSM) in the Navy Coupled Ocean/Atmosphere Mesoscale Prediction System (i.e. COAMPS) by this group, and its impact on the SABL characteristics has been investigated over the Arctic land surface. The major additions to the new snow physics package include i) adding the vegetation shading effect on snow sublimation and snowmelt, ii) limiting the liquid water stored in the snowpack during the snowmelt, iii) modifying under-canopy resistance to prevent overestimation of downward sensible heat flux, and iv) modifying unrealistically small ground soil heat flux to improve the skin temperature calculation.

COAMPS hindcasts using the new snow physics package have been systematically evaluated using available observations and compared with their counterparts using the traditional Noah LSM for two separate months (i.e., October 2015 and January of 2021). Substantial improvements have been found in variables such as 2-m dew-point temperature and relative humidity, sea-level pressure, and 10-m wind speed and wind direction, which have marked impact on the SABL characteristics.

The implemented new snow physics in COAMPS Noah LSM has passed the operational scorecard and been transited to the FNMOC for operational forecasts. It is also a powerful tool for NRL researchers to investigate the formation and persistency of stable boundary layers over higher latitude areas. It has the benefit of gaining insight into the interaction between the atmosphere and land surface processes under stable and very stable boundary layer conditions during the winter season. A manuscript based on these new results has been submitted to the Journal of Hydrometeorology for publication and is currently under revision.

## **7.3 Mesoscale variability of the stable marine boundary layer (SMBL)**

The stable marine boundary layer (SMBL) is relatively shallow and can be significantly disturbed by coastal terrain and (or) islands. A number of mesoscale or sub-mesoscale circulations, such as gravity waves, topographic wakes and shedding vortices can be generated around or downwind of the topography. While it is known that terrain-induced circulations have significant impact on the stable marine boundary layer, the turbulent adjustment processes in the SMBL in response to mesoscale circulations and their implications on electromagnetic (EM) wave ducting conditions and optical turbulence remain poorly understood. Under this project, these issues have been examined using field observations and high-resolution air-sea coupled COAMPS simulations.

Three SMBL events documented offshore of southern California during the CASPER -West field campaign have been investigated through the collaboration with Dr. Qing Wang's group at the Naval Postgraduate School. A number of physical and dynamical processes (e.g., the formation and maintenance of the shallow moist surface layer and the influence of orographic flow dynamics on the SMBL and EM duct characteristics) have been investigated based on two-way air-sea coupled COAMPS simulations. Main findings from this research have been reported in a peer-reviewed paper

published on the Journal of Geophysical Research: Atmosphere. Insights from this study have the potential of improving EM duct prediction over coastal areas and around islands.

#### 7.4 Large eddy simulations of the SABL over a heterogeneous surface

While large eddy simulation (LES) codes have been widely used for studying the atmospheric boundary layer over a homogeneous surface, it remains technically difficult to model a turbulent boundary layer over a heterogeneous surface (e.g., coastal areas and SST fronts) with LES. Two different modeling approaches have been adopted and tested under this project, namely a) the Lagrangian modeling strategy with a novel differential advection correction factor (Jiang and Wang 2021) and b) a fringe method to provide a turbulent upstream boundary condition (similar to Sullivan et al. 2020). The team has implemented the fringe method (hereafter referred to as Eulerian/Fringe method) into the NCAR/NRL-LES code and applied it to the study of the offshore internal boundary layer problem. This new capability allows us to examine the boundary layer and EM duct variation over SST fronts and coastal waters. Insights from this study have the potential to improve EM duct predictability over coastal waters and SST fronts.

Several high-resolution LES simulations have been conducted successfully and these valuable datasets have been used by three other NRL groups to explore the modified refractivity profiles over coastal waters and to assess the impact of turbulence and offshore variability in ducting conditions on EM propagation. At least two manuscripts from these efforts are in preparation. In addition, LES datasets from two high-resolution simulations had been analyzed by LCDR Kyle Franklin at Naval Postgraduate School for his radio wave propagation study, which constituted a major part of his PhD thesis and was published in a peer-reviewed journal article.

#### 7.5 Stable internal boundary layer (SIBL) over coastal waters

The characteristics and physical processes associated with the SABL over coastal waters are still poorly understood and consequently, accurate forecasts of the atmospheric environment over littoral zones remains a challenge to the NWP community. Under this 6.1 project, a number of fundamental science issues associated with this difficult problem have been examined utilizing CASPER-East observations, air-sea COAMPS simulations, and surface-layer resolving large eddy simulations. The LES simulations reveal the development of a SIBL with the surface fluxes and mean profiles comparing favorably with observations. The classic Monin-Obukhov similarity theory (MOST) and its extension, local similarity theory (LST), have been systematically evaluated using LES datasets. This research has shed new light on SIBL characteristics and physics, and a new scaling theory for the SIBL has been proposed. Major findings from this research include:

- A shallower equilibrium layer exists inside the SIBL and both the SIBL and the equilibrium layer depths tend to grow offshore, in proportion to the square-root of the offshore distance (or fetch).
- An offshore area can be divided into three zones, namely i) a nearshore zone (NSZ) where the turbulence advection from land plays a dominant role and accordingly neither MOST nor LST is valid, ii) an intermediate offshore zone (IOZ) where the land advection is less important but still non-negligible and MOST is only valid for higher order moments, and iii) a far offshore zone (FOZ) where, to a good approximation, the mean profiles in the SIBL tend to agree with the corresponding MOST prediction.

- In the NSZ and IOZ, the nondimensional wind and scalar gradients are in general smaller than the corresponding MOST predictions and the departure of dimensionless mean gradients from MOST tends to become larger at a higher level.
- Turbulent fluxes in the SIBL decrease nearly linearly with the vertical distance between the water surface and the top of the SIBL. The rapid decrease of turbulent fluxes in the vertical and the shallowness of the SIBL imply that the constant flux layer assumption for MOST is questionable.
- A new scaling theory has been proposed which takes into account the rapid decrease of turbulent fluxes with height. It has been demonstrated that the new scaling theory is substantially superior to the classical MOST over the coastal waters and consistent with the latter in the deep boundary layer and open ocean limits.

These findings have the potential to improve Navy's EM duct prediction capabilities and air-sea coupling parameterizations in Navy's NWP models over coastal waters. The proposed scaling theory is being evaluated for different meteorological conditions through an ongoing 6.2 ONR project, and will be tested using new observational data from the upcoming ONR-funded REDSAW field campaign (California, April 2024). Some results from this research have been presented in American Geological Union annual meetings and reported in a pair of peer-reviewed journal papers. The new scaling theory will be reported in a separate paper.

## 7.6 Impact of ocean waves on SIBL structure and evaporation ducts

While previous studies showed that the wind profiles in a marine boundary layer may be substantially modified by underlying ocean waves, the impact of surface waves on vertical distribution of scalars in the marine surface layer and therefore evaporation ducts for EM waves has never been studied. The latter are crucial for over the horizon radio wave propagation. This issue has been explored through a group of well-designed surface layer resolving large eddy simulations using a state-of-science LES code initially developed by NCAR scientists, which facilitates the inclusion of propagating waves at the surface. Moist processes have been incorporated into this LES code by this group to conduct the proposed research, which leads to the following major findings:

- Mean wind and scalar profiles and their variances in the marine surface layer are strongly regulated by ocean waves.
- With waves propagate against the mean winds in the SIBL, the SIBL depth can be significantly increased owing to enhanced turbulence. When the waves propagate along the mean wind direction, the SIBL depth becomes much shallower with weakened turbulent mixing.
- The conventional similarity theory (i.e., MOST) based evaporation duct prediction algorithms (i.e., currently used by the Navy) become inadequate. The impact of waves on scalars is more dramatic when the marine boundary layer is more stable.

The findings from this research have implications on air-sea coupling and EM duct prediction over ocean waves. Some of the results have been reported in a peer-reviewed journal paper.

## 7.7 Optical turbulence ( $C_n^2$ ) in the SIBL

Optical turbulence in the atmosphere is best represented by the refractivity structure parameter (Cn2). Forecast of Cn2 profiles in the marine boundary layer at both the radio and optical frequencies is needed in order to properly represent the turbulence effects on EM/EO propagation. Through close collaboration with our NRL colleagues (i.e., Dr. Shouping Wang's group), a second-order turbulence closure Cn2 model has been formulated and developed within the framework of the Navy's COAMPS model (i.e., COAMPS-Cn2). COAMPS-Cn2 explicitly predicts the water vapor and temperature variances and covariance, which are crucial for adequately computing Cn2. This predictive approach substantially improves the Cn2 prediction by partially taking into account non-local turbulent processes. Our group has conducted a number of high-resolution LES simulations of SABL over waters, based on which the eddy diffusivity for the variance and covariance mixing has been optimized. The Cn2 profiles predicted by the COAMPS-Cn2 have shown reasonable agreement with those from high-resolution large eddy simulation results and the classical MOST predictions. The new Cn2 model predictions compare favorably with the observations and additional high-fidelity large eddy simulations. These results were published in a peer-reviewed journal article and the Cn2 module has been transitioned.

In summary, we have explored the fundamental physical and dynamical issues associated with a number of important physical processes in the SABL. Significant findings from this research have been presented in workshops and scientific meetings and/or published in top-tier peer-reviewed journals. Some new findings and high-resolution large eddy simulation datasets have high transition potentials. For example, the SABL over wave study and the new scaling theory over coastal waters imply that the air-sea coupling parameterizations in Navy's NWP models and EM duct prediction tools need to be improved over ocean waves and over coastal waters. These issues will be further explored under a new NRL base project and an ongoing ONR project.

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