DOCTORAL DISSERTATION

The Behavior of the Atmospheric Boundary Layer in the Vicinity of the Gulf Stream Sea Surface Temperature Front

by

Hanyuan Liu

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Submitted to the Earth, Atmospheric and Planetary Sciences on August 11, 2023, in partial fulfillment of the requirements for the degree of Doctor of Philosophy

Abstract

The evolution of the marine atmospheric boundary layer (MABL) in the vicinity of a sea surface temperature (SST) front is of particular research interest, as the large air-sea temperature and humidity differences at the surface fuels various physical processes inside the boundary layer, causing intense heat and momentum exchange. Such processes make the mesoscale MABL an ocean-drive-atmosphere scenario. Dominant mechanisms, although having been studied intensively, are still yet to be fully understood due to the highly turbulent nature of the MABL. Previous studies often relied on satellite-derived SST and wind fields to investigate boundary layer dynamics, yet the coarse spatial and temporal resolution of such a method limits the understanding of the MABL evolution on shorter timescales.

In this thesis, a combination of in situ data and model simulations is used to investigate the MABL response to the SST front in the Gulf Stream region on a timescale of one day or less. Analysis of MABL structure is divided into three categories depending on the background wind strength and its direction relative to the front: cold to warm, parallel/weak, and warm to cold. Two mechanisms identified in previous studies, vertical mixing and thermally induced pressure gradient, and their role in MABL evolution, are studied quantitatively. A comparison between observations and model simulations allows further analysis of the contribution of moist processes that were often considered to be of secondary importance in the past even over the ocean. Results show that vertical mixing is responsible for the majority of the MABL deepening, while the pressure adjustment’s effect is more significant when the cross-frontal wind is weak. Sensitivity tests conducted in the Weather Research and Forecast (WRF) also show that moisture processes, including surface latent heat, boundary layer transport of moist, and cloud formation, further enhance the mixing that drives MABL changes.

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Chapter 1
Introduction

The ocean-atmosphere coupling due to surface variability has been categorized into two regimes over the past decades of intensive field observations and numerical modeling efforts (Seo et al., 2023), the first being the larger scale (>1000km) ocean response to atmospheric variabilities, and the second being the mesoscale (10-1000km) atmospheric response to the underlying ocean surface variabilities including mesoscale eddies, sea surface temperature (SST) anomalies and currents. While the first regime is considered being relatively passive, the second regime is of particular interest in this thesis, as the atmospheric response to a SST front involves various processes, including wind modulation, cloud formation, and precipitation, that are crucial to understand mesoscale climate variabilities (Seo et al., 2023). These responses also provide feedback to the ocean and fuel the evolution of ocean circulation, which further affects the aforementioned atmospheric response, forming a strongly coupled ocean-atmosphere system. (e.g. Small et al., 2008; Seo et al., 2023) This coupling has been determined to have a significant impart on regional weather and climate, and further investigations on it are much needed for both better performance in coupled weather and climate model simulations and deeper understanding of the mesoscale ocean-atmosphere dynamics. Less work has been done to analyze the role of changes in the near-surface atmosphere and how these may impact through to the weather and climate scales.

To understand to such a process, studying the evolution of the marine atmospheric boundary layer (MABL) in the vicinity of an SST front is crucial, as it contains processes that are directly related to the said atmospheric response to the underlying ocean surface that contains distinct variabilities. MABL is defined as the part of the troposphere that is directly influenced by the presence of the sea surface and responds to surface forcings with a timescale of about an hour or less (Stull 1988). It is directly affected by the air-sea temperature difference induced by the SST gradient (Stull 1988; Small 2008), mainly via
processes including surface sensible and latent heat exchange, surface layer modulation due to stability change, and currents effect.

Mesoscale variability across western boundary currents (WBC) and their meanders impacts atmospheric circulation, but questions remain about the role of the MABL in modulating the air-sea exchange. The Gulf Stream, a western boundary current (WBC) that brings warm water from low to high latitudes along the east coast of the United States, is characterized by a sharp sea surface temperature (SST) difference. This SST front is often greater than 10°C between the cooler shelf water and the core of the Gulf Stream (Wai and Stage 1981). Changes in the boundary layer structure due to such a high SST difference lead to significant MABL evolution that alters the system state and responses in both the atmosphere and the ocean surface (Stull 1988), and thus the evolution of the MABL in the vicinity of the Gulf Stream SST fronts is a case of interests in terms of understanding the mesoscale atmospheric response.

Main processes have been proposed and studied in past decades regarding the changes in the MABL across these strong SST gradients (e.g., Spall 2007; Small 2008; Seo et al., 2023), but the role of surface forcing, clouds, and MABL entrainment and boundary layer growth are still relatively understudied. Field experiments in regions where a sharp SST front exists have been conducted over the last decades (e.g., Frihe 1991, de Szoeke 2005), while numerical modeling that simulates the MABL near an SST front has also been intensively used (e.g., Spall et al. 2007; Skyllingstad et al. 2007; Kilpatrick et al., 2014; Schneider and Qiu, 2015). These studies used more quantitative and theoretical analysis on thermal, momentum, and turbulent kinetic energy (TKE) budgets from idealized model results to study the governing dynamics that lead to observed changes in MABL.

In the previous literature, several important mechanisms have been identified to have significant impact on the observed boundary layer growth and surface wind change (e.g., Small 2008; O’Neil et al. 2010; Kilpatrick 2014; Schneider 2015). Stability induced surface stress change across the SST front (e.g., Wai and Stage 1989), vertical mixing due to enhanced surface fluxes (e.g., Spall 2007), thermally induced pressure gradient (e.g., Lindzen and Nigam 1987), and a surface current effect (e.g., Kelly et al. 2001) are
the most significant mechanisms that account for ocean-driven surface wind change near an SST front (Small 2008). Previous studies have shown that the background wind direction and strength is crucial (Friehe 1991), as the cross-frontal wind component directly determines the strength of advection and the adjustment time for an air mass to react to the SST (Kilpatrick 2014). Thus, it is important to study the MABL boundary in different wind cases. While previous studies have made expensive progress over the past decades, the rapidly changing nature of the MABL near an SST front makes comprehensive observation of the boundary layer a challenging task. For example, the spatial and temporal interval between observations is inevitably large with traditional moving platforms (e.g., ship) that often aim to cover a large area in the vicinity of the SST front, which results in low resolution measurements. As a result, in situ observations in the MABL’s two-dimensional structure often face the problem that they are most useful for the longer timescale of 7-90 days (e.g., Plagge et al., 2016), but become less useful when the timescale is close to 1 day or less, Unfortunately this is the scale over which MABL adjustment often occurs (Stull 1988; Spall 2007; Kawai et al. 2014). For this reason, previous studies often relied on numerical simulations to resolve the fine dynamics inside the boundary layer. Even over the ocean, it is common in these simulations to neglect moist processes (e.g., latent heat flux, boundary layer transfer of atmospheric moisture, and cloud formation), as the dry dynamics driven by sensible heat exchange are considered to be the main driver of boundary layer change in their simplified setup (e.g., Kilpatrick 2014; Schneider 2015). Schneider (2015), however, pointed out that moist process should be a focus for MABL studies. Over the sea surface, moist processes including the surface moisture flux that enhances mixing that can directly affect the overall strength of vertical mixing throughout the MABL (Garratt 1985). de Szoeke (2005) also showed that the formation of clouds can lead to enhanced entrainment via cloud top radiative cooling using measurements from the EPIC experiment. These and other moist processes, however, are still less emphasized in many models simulations. The lack of moist processes also makes it difficult to compare model results to realistic observational data.

The goal of the thesis is to use a combination of observations and model simulation results to study the behavior of the MABL near the SST front in the Gulf Stream region.
In Chapter 2, in situ observations from several field campaigns in this region are documented and studied. The observation window of each shipboard transect is 12-24 hours, aiming to capture the boundary layer evolution within this time scale. For each transect, the major mechanisms affecting the boundary layer change are identified and their contribution is quantified. In Chapter 3, a series of idealized 2-D Weather Research and Forecasting (WRF) model simulations are studied to further differentiate the strength of each mechanism under different wind conditions. Model dynamics are turned on and off to investigate and distinguish the effect of moist versus processes. These results are compared with observations in Chapter 2. In Chapter 4, a fully-dynamic 3-D WRF simulation is used to bridge the gap between the idealized 2-D simulations and field observations. In Chapter 5, a summary of results, as well as potential future directions, is presented.
Chapter 2
Field Observations

Abstract

In situ observational data are analyzed to study the behavior of the marine atmospheric boundary layer in the vicinity of the Gulf Stream, where a sea surface temperature front with a sharp temperature difference larger than 6 degrees C is present. Ship transects from the CLIVAR Mode Water Dynamic Experiment (CLIMODE) and the Processes Driving Exchange at Cape Hatteras (PEACH) campaigns provide both surface measurements and atmospheric observations via Rawinsonde launches. The evolution of the MABL near the SST front is categorized into three cases based on the background strength and its direction relative to the SST front: strong cold side to warm side wind, wind parallel to front/weak wind, and strong warm to cold wind. Different analysis methods are used in each case to differentiate the role of the SST front in boundary layer adjustments. Two mechanisms are examined as the possible main contributors to observed MABL changes: vertical mixing and thermally induced pressure adjustment. Both qualitative and quantitative analyses show that the majority of observed boundary layer height growth and entrainment are due to vertical mixing. Surface buoyancy flux induced by the air-sea temperature difference near the SST front provides the bulk of the estimated entrainment at the top of the mixed layer, while the shear driven entrainment and cloud top radiative cooling provide a lesser portion. The pressure gradient adjustment is not significant in the strong cross-frontal wind case but can be responsible for observed surface wind acceleration in the weak wind case. Results also show that the stability induced surface stress change only contributes a minor amount to surface wind change. Only qualitative analysis can be performed on the warm to cold case due to limited data availability, but the formation of the stable boundary layer and its evolution is consistent with previous studies and model results.

2.1 Introduction

The Gulf Stream is a western boundary current that brings warm water from lower latitudes to northward along the coastline of the United States. In this region, a sharp sea surface temperature difference often exceeding 10 °C between the cooler shelf water and the core of the Gulf Stream is observed (Wai and Stage 1989). Due to this sharp SST gradient, air-sea interaction inside the overlying boundary layer is often enhanced significantly through intense sensible and latent heat exchange, which drive conditions where the ocean forces a response in the atmosphere (Warner et al., 1989; Small et al. 2008; Plagge et al., 2016). Thus, it is of great interest to study the response and dynamics
inside the boundary layer in the vicinity of the Gulf Stream, where strong SST gradients are found.

While this topic has been studied intensively in the past decades, previous studies often relied on numerical simulations and large-scale satellite data due to the scarcity of in situ measurements (Skyllingstad et al. 2007; Doyle et al. 1993) These studies found a significant correlation between the SST gradient, surface stress, wind speed and pressure gradient (Plagge et al., 2016; O’Neil 2003, 2005, 2010) on time scales of 7 to 90 days, However, it is not known whether these findings are robust on time scales of 1 day or less, which is the scale over which the marine atmosphere boundary layer (MABL) adjustment often occurs (Spall et al. 2007; Kawai et al. 2014) The spatial and temporal resolution of satellite data is often insufficient to resolve boundary layer adjustments, which can occur within 1 hour and over 10km (Stull 1988).

Previous field experiments conducted in other regions where sharp SST fronts exist have provided valuable observations over time scales shorter than those that satellite observations can resolve. For example, studies on the Frontal Air-Sea Interaction Experiment (FASINEX) examined the boundary layer stability and height adjustment in different wind regimes in Bermuda. (Frihe 1991). The Marine ARM GCSS Pacific Cross-Section Intercomparison (GPCI) Investigation of Clouds (MAGIC) documented the evolution of MABL clouds during several ship transits near California and Hawaii (Xiaoli et al. 2015). Observations from the Eastern Pacific Investigation of Climate 2001 (EPIC2001) described the atmospheric boundary layer adjustments and tropical instability waves (TIW) in the equatorial ITCZ region where a northward SST gradient exists (de Szoeke et al. 2005; McGauly et al. 2004). In the Agulhas Current Air–Sea Exchange Experiment, the response of the MABL under along-front winds was examined (Rouault et al. 2000). These in situ measurements included aircraft flights, Rawinsonde launches, and direct covariance/bulk estimation of surface heat fluxes and produced datasets with both high temporal and spatial resolution near an SST front, which are essential to resolve the evolution of the boundary layer in response to the underlying SST gradient.
Building on this earlier work, this study aims to investigate boundary layer evolution by documenting and analyzing in situ measurements including ship-board flux measurements and Rawinsonde launches in the region of the Gulf Stream from three field campaigns: the CLIVAR Mode Water Dynamic Experiment (CLIMODE) pilot experiment in January 2006, the CLIMODE main cruise in February and March 2007 (Marshall et al., 2009), and the Processes Driving Exchange at Cape Hatteras (PEACH; Seim et al. 2022) recovery cruise in November 2018. One objective of this study is to analyze the evolution of the boundary layer adjustment with several possible dynamics, including pressure adjustment, vertical mixing, and stability induced surface stress change.

Monin-Obukhov Similarity Theory, Surface Heat Flux and Turbulence induced Boundary Layer Modification

The atmospheric boundary layer (ABL) is defined as the part of the troposphere that is directly influenced by the presence of the earth’s surface and responds to surface forcings with a timescale of about an hour or less (Stull 1988). One major characteristic of the boundary layer is the presence of turbulent flows. The generation and dissipation of turbulent kinetic energy (TKE) contributes significantly to the flow structure and mixing inside the MABL. The surface layer, which spans approximately the bottom 10% of the total boundary layer thickness, is directly affected by sea surface friction and experiences almost constant fluxes of heat, moisture, and momentum across it. Within the surface layer but outside the wave boundary layer where turbulence is more complicated due to surface waves, the turbulent flow can be described with the Monin-Obukhov similarity theory (MOS), whereby a set of scaling parameters are used to non-dimensionlize variables inside the surface layer including velocity, moisture and heat flux (Stull 1988, Plagge 2016). With MOS, a friction velocity $u_*$ can be defined as:

$$u_* = \left[ \left( \overline{u'w'} \right)^2 + \left( \overline{v'w'} \right)^2 \right]^{1/4} \quad (2.1)$$

where $u'$, $v'$ and $w'$ are horizontal and vertical velocity fluctuations relative to a temporal mean. With this friction velocity, the surface wind stress can then be expressed as:
\[ \tau = -\rho_a (u'w') \cong \rho_a u_*^2 \quad (2.2) \]

where \( \rho_a \) is the air density. Within the boundary layer, dimensional-analysis provides a length scale \( L \) known as the Monin-Obukhov Length scale, such that:

\[ L = -\frac{\Theta_v u_*^3}{g \kappa \omega' \theta_v'} \quad (2.3) \]

where \( \Theta_v \) is the mean virtual potential temperature, \( \kappa \) the Von Karman constant, \( g \) the local gravitational acceleration and \( \omega' \theta_v' \) is the buoyancy flux. The MO length represents the ratio of shear-induced turbulence to buoyancy-driven turbulence. The surface layer below the height where \( z = L \) is dominated by shear, while that above is dominated by buoyancy. Note that the virtual potential temperature is defined as:

\[ \Theta_v = \Theta (1 + 0.61r - r_L) \quad (2.4) \]

where \( \Theta \) is the potential temperature, \( r \) the mixing ratio of water vapor and \( r_L \) the mixing ratio of liquid water in the air. The use of virtual potential temperature is convenient as it includes the impact of moisture on the buoyancy of air parcels in the MABL. Within the surface layer, MOS predicts that the normalized wind shear is a function of \( z/L \):

\[ \frac{\kappa z U}{u_*} = \varphi_m (z/L). \quad (2.5) \]

Using these dimensionless parameters, the characteristics of the surface boundary layer, such as the logarithmic wind profile and drag coefficient, can be modeled as a function of stability. Methods are described in detail in previous studies (e.g., Fairall et al. 1996; Edson et al. 2003, 2004; Plagge 2016). In this study, the Coupled Ocean–Atmosphere Response Experiment (COARE) algorithm (Fairall et al. 1996, 2003; Edson et al. 2013) is used to estimate these variables, generating bulk estimates of surface fluxes and stability parameter \( z/L \). One important variable to note is the buoyancy flux, which has the form of:

\[ BHF = \rho_a C_p \omega' \theta_v' \cong -\rho_a C_p u_* T_{v*} \quad (2.6) \]
where $C_p$ is the specific heat at constant pressure, and $T_{v*}$ is the virtual temperature scaling parameter. The buoyancy heat flux combines the sensible heat and moisture exchange at the surface and serves as a measure of the strength of the buoyancy driven turbulence (Stull 1988).

In the region where a sharp SST gradient exists, the buoyancy flux is often positive (upward) on the warm side of the SST front, representing the convective boundary layer driven by the warming of overlying air by the warmer sea surface. In this case, a convective velocity scale can be defined as:

$$w_* = \left[ \frac{g}{\theta_v} \overline{w'\theta'_v} \right]^{1/3}$$

(2.7)

where $z_i$ is the average mixed layer (ML) height. The s subscript denotes the buoyancy flux at the surface. In strongly convective situations, the magnitude of the velocity of the uprising thermals caused by surface heating is close to the convective velocity scale. Strongly convective conditions are often observed in regions of the SST front where the wind is blowing from the cold side of the front to the warm side. This results in the advection of cold air above warm water, which drives an unstable convective boundary layer, enhances the vertical turbulent mixing, and increases the boundary layer height.

Uprising thermals reach the warmer free atmosphere (FA) at the top of the mixed layer where, although they may overshoot a short distance (Deardorff et al. 1969), their vertical motion becomes largely "capped" at the capping inversion. During this process, dry and warm air from the FA is rapidly mixed down to the MABL due to strong turbulence, causing the ML to erode into the FA and deepening the boundary layer. The entrainment flux of air, $w_e$, together with the subsidence velocity at the top of the boundary layer $w_L$, can be used to describe the boundary layer height development:

$$\frac{dz_i}{dt} = w_e + w_L.$$  

(2.8)

Previous studies have shown the importance of this mechanism (e.g., Frihe 1991; McGauy et al. 2004; Rouault et al. 2000; Hayes et al. 1989; Song et al. 2004) by separating the wind regime into different cases. In a case where cold air is above warm SST, a significant increase in boundary layer height and buoyancy flux is often observed.
The coupling between surface stress, wind field and SST has been observed and studied extensively (e.g., Giordani et al. 2001; Chelton et al. 2001; Wallace et al. 1989; O’Neil et al. 2010; Plagge 2016). Two major mechanisms related to the surface flux and stability have been shown to the increase in momentum exchange due to 1) enhanced mixing in the surface layer over warm water described by MOS, and 2) the mixing of high-momentum air from the FA due to entrainment (Spall 2007; Small 2008). The first mechanism can be investigated using the diabatic profile derived by rearrangement and integration of Eq. 2.5:

\[ U(z) = U(z_0) + \frac{u_\ast}{\kappa} (ln(z/z_0) - \psi_m(z/L)) \]  

(2.9)

where \( z_0 \) is the surface roughness length and \( \psi_m \) is a function derived from the integral of \( \varphi_m \). The surface wind stress can be defined as:

\[ \tau = \rho u_\ast^2 = \rho C_D U^2 \]  

(2.10)

where \( C_D \) is the drag coefficient. The drag coefficient can be defined using Eqs. 2.9 and 2.10 as:

\[ C_D = \kappa^2 (ln(z/z_0) - \psi_m(z/L))^2. \]  

(2.11)

Field experiments have shown that the \( \psi_m \) function is positive under unstable conditions such that it models the enhancement of the drag coefficient due to convective mixing in Eq. 2.11 resulting in a larger stress as predicted by Eq. 2.10. Coupling due to both mechanisms has been seen in previous studies (Sweet et al. 1981; Businger and Shaw 1984; Hayes et al. 1989; Nonaka et al. 2003). Vertical wind shear in a convective boundary layer is often observed to be reduced due to the well-mixed structure of the mixed layer. In a case where the stability parameter \( z/L \) is less than 1 or negative (unstable), the diabatic profile \( \psi_m(z/L) \) will act to increase the magnitude of \( C_D \) according to Eq.2.11, and the surface wind stress will increase due to the increase in \( C_D \), causing a surface wind acceleration, hence the increase in momentum due to stability induced surface stress. Coupling due to both mechanisms has been seen in previous studies (Sweet et al. 1981; Businger and Shaw 1984; Hayes et al. 1989; Nonaka et al. 2003). Vertical wind shear in a
convective boundary layer is often observed to be reduced due to the well-mixed structure of the mixed layer.

In a case where warm air is advected to a cooler sea surface, a stable internal boundary layer is likely to form due to the negative (downward) buoyancy flux (Small et al. 2008; Stull 1988; Hsu 1983). The difference between the SST and the air temperature modifies the surface layer to a neutral-to-stable layer, where strong stratification counteracts the mechanical production by wind shear. The reduced vertical mixing causes the top of the boundary layer to decouple from the surface layer and becomes a residual layer, where the subsidence velocity becomes dominant and acts to lower the residual layer height. At the lower 50m - 200m of the boundary layer, a stable layer with $\theta_v$ increasing with height (inversion) is often present. The growth of the stable boundary layer (SBL) can be modeled as

$$ H_{\Delta \theta} = (-Q_T \cdot t \cdot B)^{1/2} $$

(2.12)

where $H_{\Delta \theta}$ is the SBL height, $Q_T$ the total heat flux acting on the SBL, $t$ the time elapsed and $B$ the bulk turbulence scale, which is the ratio of the SBL height to the difference between residual layer air and the near-surface air (Stull 1988).

In the region where a clearly defined SST front is present, the large magnitude of air-sea temperature difference when warmer air is advected to the cooler sea surface often causes the stability parameter to change sign (Rouault et al. 2000). In theory, this should cause both a decrease in surface wind stress according to Eq.2.9 and a lack of vertical mixing of higher momentum air from the FA due to the formation of an SBL. Friehe (1991) and Song (2004) reported a slowing of wind on the cold side of the front. However, winds in SBL can have very complex characteristics (Stull 1988). The decrease in roughness and drag may act to accelerate the wind in cases where complex forcings are present (Small 2008), thus when investigating the wind field within the SBL it is important to consider the temporal and spatial scale, and the interaction with larger-scale processes such as cyclogenesis.

Another characteristic of an SBL is the formation of a low-level jet (LLJ). An LLJ is a thin stream of fast-moving air with maximum wind speed of 10 to 20 m/s usually located
100 to 300m above the ground (Stull 1988). It has been argued that within an SBL, the ML turbulence weakens and the pressure gradient tends to accelerate the wind to reach geostrophic balance, however the inertial oscillation due to the Coriolis force will cause the wind to be 2 to 5 m/s faster than the geostrophic wind (Garratt, 1985; Kraus et al. 1985). In the vicinity of an SST front, such a phenomenon is observed on the cold side of the front (Vihma et al. 1998), and the LLJ in a vertical wind profile can often help to identify the SBL height as it often resides on the top of the SBL (Stull 1988).

**SST Gradient Induced Pressure Gradient**

Another important process that often occurs in the vicinity of a sharp SST front is the thermally induced pressure gradient and the resulting secondary circulation. Lindzen and Nigam (1987) used a simple one-layer model of the trade cumulus boundary layer to examine the local influence of the SST gradient on the flow forced by an induced pressure gradient in the Pacific ITCZ (Inter Tropical Convergence Zone) region. They found that the local flow was very sensitive to the pressure gradient and the resulting low-level tropical flow and convergence were prominent, yet not related to net latent heat release. They suggested that the secondary flow could have a profound influence on the mass redistribution and wind field adjustment inside the boundary layer.

Wai and Stage (1989) used a numerical model to investigate the MABL adjustments in the Gulf Stream region. In their study, they found that the difference in SST created variations in the horizontal surface pressure perturbation, and as a result, a thermal cell was formed. The upward motion on the warm side led to the boundary layer deepening. They also suggested that the increased surface stress on the warm side was due to the baroclinic wind field rather than the change in drag coefficient due to stability change.

Small et al. (2005) used a numerical model to study the cross-frontal flow adjustment in the region of the Pacific cold tongue. Based on their analysis on the momentum budget from the model, they suggested that the hydrostatic pressure set up by the SST gradient significantly contributes to observed wind acceleration, which was consistent with observations from Tropical Atmosphere Ocean (TAO) moorings and the EPIC campaign.
O’Neill et al. (2010) used the Weather Research and Forecasting (WRF) model with monthly mean output to investigate the flow adjustment in the Agulhas return current region and found that the aforementioned vertical momentum mixing mechanism acts in concert with the pressure gradient to accelerate the flow downwind.

In terms of observations, Mahrt et al. (2004) used aircraft eddy-correlation data to investigate the evolution of MABL off the coast of North Carolina. A significant horizontal pressure gradient due to warming of the air was observed downwind of the SST front, and an acceleration in the wind field that was consistent with the local horizontal pressure gradient was also observed. They thus hypothesized a local low-pressure system superimposed on larger scale pressure system. The pressure perturbation induced thermally in their study was calculated in the form:

\[ P_t = \frac{g}{\theta} h \frac{\partial \theta}{\partial x} \] (2.13)

where \([\theta]\) is the potential temperature vertically integrated from the surface up to a height \(h\), and \(\Theta\) is a scale value of the potential temperature. They pointed out that, though there were inadequate observations of the vertical structure to resolve the process, the sign of the local pressure gradient was consistent with the flow acceleration. Note that in this study the value \(h\) is the MABL height, instead of a constant value as in Mahrt (1982) and Kawai et al. (2014), because we argue that in a rapidly growing boundary layer, the top of the MABL better represents the upper boundary of the surface-heating-induced flow.

More recently, Kawai et al. (2014) conducted observations to Mahrt in the Kuroshio region. They calculated the thermally induced pressure gradient using an approach similar to Eq.2.14. They found that the spatial perturbation in surface level pressure (SLP) tended to correlate with SST perturbation, and low-level convergence was suggested over regions with higher SST, in which case a sea-breeze like circulation would be formed. They also pointed out, however, that such a relationship was not always evident during the observation period.

Note that in Mahrt (2004), the potential temperature, \(\theta\), is used in Eq.2.13, while in Kawai (2014) the virtual potential temperature is used. The use of a virtual potential
temperature accounts the additional effect of moisture in the atmosphere, and should be used to quantify the impact of moisture on the thermally induced pressure gradient. In this thesis, we use virtual potential temperature for observational data including sounding profiles and pressure gradient calculation. But in later chapters where a model simulation is considered, the potential temperature is used instead to better compare with previous studies where the atmosphere was set to be completely dry (e.g, Kilpatrick 2014). In general, it is more appropriate to use virtual potential temperature in future studies of the MABL, as this study will clearly show the importance of moisture in MABL processes.

Plagge et al. (2016) investigated the coupling between surface wind stress and SST gradient using in situ buoy measurements in the Gulf Stream region from CLIMODE and wind measurements from the QuickSCAT satellite. They observed an atmospheric pressure perturbation of 0.5 hPa/ºC along the SST pressure gradient on a time scale of 10-100 days. A peak negative correlation between pressure and wind speed was observed on a 50–70-day scale. They pointed out that the inferred strength of coupling between wind field, surface stress and the SST gradient was highly dependent on the analysis timescale. In general, previous studies have come to agreement on the contribution to the wind speed and surfaced stress from the local pressure on time scales longer than a week and in numerical simulations. However, this correlation between surface level pressure and SST is less obvious in in-situ observations on shorter time scales of 1 day or less (Kawai et al. 2014). One plausible explanation for such a weaker correlation can be the superimposition of a larger scale pressure system in the region of interest, e.g., atmospheric tide or synoptic storms. Pressure perturbations from these larger scale systems can often counteract small-scale processes and modify flows in a less predictable way on shorter time scales (Bluestein, H. B., 1992).

Cloud Formation and Radiative Cooling

It has been recognized that the Gulf Stream has a profound influence on cloud formation in its vicinity (e.g., Wai and Stage, 1989). Bands of stratocumulus clouds are often observed along the north wall of the Gulf Stream due to the large latent heat flux over the warmer ocean surface. It is very common to observe formation of stratocumulus over the
top of a mixed layer over warmer water, and its feedback to the ML is not negligible when we study the evolution of the boundary layer (Stull 1988). After the formation of boundary layer clouds, the surface will be shaded and, as a result, less of solar radiation is received at the surface. This often creates negative feedback over land, as fewer thermals can be generated due to reduced surface heating, and the growth of the mixed layer will be slowed. However, over a warm sea surface, the reduction in solar radiation often causes much less temperature variation than it does over land due to the relatively stable SST. As a result, shading caused by cloud cover often has a less pronounced feedback on boundary layer growth over the ocean than over land.

Another important mechanism associated with boundary layer clouds is the cloud-top radiative cooling. This process creates upside-down “thermals” of cold air sinking from the cloud top and acts as a form of entrainment (Stull, 1988). The entrainment velocity when cloud-top radiative cooling dominates can be approximated by:

$$w_e \cong \frac{\Delta I^*}{\Delta E_Z \theta_v}$$

(2.14)

where $\Delta I^*$ is the net longwave radiative flux divergence near the cloud top and $\Delta E_Z \theta_v$ the virtual potential temperature difference across the entrainment zone (Stull 1988). In boundary layer height development, such an entrainment can be as important as surface heat fluxes. de Szoeke et al. (2005) used the following equation to estimate the entrainment velocity in the ITCZ zone during EPIC2001 campaign:

$$U \frac{\partial z_i}{\partial x} = w_e + w_L$$

(2.15)

where $U$ is the averaged meridional wind advecting the inversion. They estimated $w_L$ with regional analysis and calculated $w_e$ with the equation below, assuming that only surface heat flux forced the entrainment

$$w_e = \frac{A w_*^3}{h \Delta b}$$

(2.16)

where $\Delta b = g \Delta \theta_v / \theta_v$ is the buoyancy jump capping the inversion, and $A$ is the fraction of surface flux contributing to entrainment. Their estimation of entrainment rate through Eq.2.17 with $A = 0.2$ was much lower than the observed entrainment rate, i.e., the left-
hand side of Eq.2.16. After conducting a thermodynamic budget analysis, they pointed out that the difference was primarily due to the cloud top radiative cooling, which contributed nearly half of the total surface heat flux. Thus, the existence of clouds and related buoyancy flux from the top of the mixed layer is not negligible and should not be neglected when investigating the MABL structure in the vicinity of the Gulf Stream.

Over cooler water, the reduced latent heat flux often leads to a boundary layer with less cloud formation. In this case fog can form if SST is much cooler than the moist air above and heav condensation occurs. Similar to the mixed layer case, intensive radiative cooling at the top of the thickened fog layer can occur and acts to mix the fog layer (Stull 1988).

Due to the lack of dedicated cloud measurements and long-wave divergence in the datasets presented here, this study aims to discuss qualitatively the formation of clouds and their potential impact on the boundary layer adjustments based on the vertical profiles of humidity and temperature acquired from Rawinsonde launches. An estimate of cloud top radiative flux is made based on observations and values obtained from previous studies (e.g., Zheng et al. 2018).

**Surface Current and Its Effects on Estimations**

Due to the nature of a WBC, the current speeds within the Gulf Stream can be relatively fast compared to the speeds in the neighboring cooler slope seawater, sometimes reaching maximum at order of 1m/s (Joyce and McDougall, 1992). The effect of such a fast moving surface current on the measured wind field and surface stress has been long studied. Kelly et al. (2001) discussed the difference in satellite-based surface wind measurements and direct measurements from TAO moorings’ anemometers and pointed out that the satellite scatterometer measured surface stress instead of surface wind, which provided a better estimate of the relative surface wind by taking the surface current speed into account. They suggested a 50% error or even reversed sign could be generated if the relative motion was not considered when setting boundary condition for oceanic and atmospheric models. Similar note was also made in Rouault et al. (2000), where they proposed to estimate surface fluxes with following equations:
\[ \tau = \rho_a C_D (U - U_s)^2 \quad (2.17) \]

\[ Q_H = \rho_a C_p C_H (U - U_s)(T_s - \Theta) \quad (2.18) \]

\[ Q_E = \rho_a C_E (U - U_s)(q_s - q) \quad (2.19) \]

where \( C_D \) is the drag coefficient, \( C_H \) and \( C_E \) the transfer coefficient for sensible and latent heat, \( U_s \) the surface current speed, \( T_s \) and \( q_s \) the mean temperature and specific humidity at the surface. Based on previous observation of surface current speed of 2 m/s (Rouault et al. 1995), they suggested that a 50% error would occur in surface flux estimations under a 4 m/s measured surface wind.

Cornillon and Park (2001) used the NASA scatterometer, NSCAT to investigate the surface current within warm core rings in the Gulf Stream region. They again confirmed that scatterometer measured stress instead of wind and extra caution was needed treating surface condition within the boundary layer. In their later work (Park et al. 2006), they provided further evidence of the MABL modification induced by surface currents and SST gradients in Gulf Stream warm rings.

Seo et al. (2016) studied eddy-wind interaction in the California Current system with a high-resolution regional model. They found that the surface current modification on the eddy was significant, primarily through increased surface drag. They also pointed out that the eddy induced surface current and SST gradient strongly modified the Ekman pumping, suggesting the importance of surface currents in ocean-atmosphere coupling.

While it is beyond the scope of this study to extensively and quantitatively examine the role of the observed surface currents on boundary layer modification, it is worth noting that surface current speed is taken into account when calculating the bulk fluxes, as relative wind speed is derived based on ship measurements of the current. Note of the surface current effect is taken here when comparing observations from cold and warm sides of the front, where the difference in current speed is significant.
The Half-power Dependence on Distance

The growth of the boundary layer height can be a function of distance from the front (fetch), according to Eq.2.15. Previous studies suggested a half-power dependence on fetch to take into account the sharp transition very close to the front where the air-sea temperature difference is large, and the near equilibrium state far away from the front where ML growth is slow. Venkatram (1977) proposed the following function over a convective boundary layer:

\[
Z_i = \left[ \frac{2CD|\bar{\theta}_{\text{warm}} - \bar{\theta}_{\text{cold}}|X}{\gamma(1-2F)} \right]^{1/2}
\]  

where \( \bar{\theta}_{\text{warm}} \) and \( \bar{\theta}_{\text{cold}} \) the potential temperature on the warm side and cold side, \( X \) the fetch, \( \gamma \) the lapse rate above the boundary layer and \( F \) an entrainment coefficient, ranging 0 to 0.22. Hsu (1983, 1984) used several observations to verify Eq.2.20 in both the convective boundary layer and the warm to cold case, and after considering variations in parameters, he proposed the following relation:

\[
Z_i = c X^{1/2}
\]  

where \( c \) is a dimensionless parameter depending on the overall boundary layer condition.

In a stable internal boundary layer, Garratt (1987) proposed that

\[
h = 0.014 \bar{U} \left[ \frac{X \bar{\theta}}{g \bar{\Delta \theta}} \right]^{1/2}
\]  

where \( \bar{\Delta \theta} \) is the air-surface temperature difference before the growth of the boundary layer.

In this study, Eq.2.21 is used to evaluate the boundary layer growth rate based on the magnitude of \( c \). Different values of \( c \) are related to the observed boundary layer condition. In the warm to cold condition where an SBL is likely to form, Eq.2.22 is used here to compare observations with theoretical values.
2.2 Observation Background

CLIVAR Mode Water Dynamic Experiment (CLIMODE)

The CLIVAR Mode Water Dynamic Experiment (CLIMODE) is a field campaign conducted in 2006 and 2007 in the region of the Gulf Stream, ranging from 36 to 39° N, 62 to 70° W. It aims to observe the convection and re-stratification processes over the North Atlantic Sub-Tropical Mode Water (STMW). High spatial-resolution data was collected with mooring, floats, drifters and shipboard measurements (Marshall et al. 2009). There were two cruises conducted in the winter of 2006 and 2007 to study the wintertime intense mixing processes. This study focuses on observations from these two cruises and a description of their overall conditions is provided in the following sections.

CLIMODE Pilot Cruise

The cruise between Jan. 18 and Jan. 31, 2006 was conducted on R/V Atlantis and was referred as the Pilot cruise. A map of the ship track with an underlying SST field is shown in Figure 2-1. The ship track was designed to cross the SST front in a direction nearly perpendicular to the front to yield better observation across the front. Shipboard measurements included a full flux package that provided both direct-covariance flux measurements and bulk flux estimates with the COARE algorithm using the approach described by Edson et al. (1998). Along the ship track, 30 Rawinsonde launches were made to monitor the evolution of the atmospheric boundary layer. The overall atmospheric conditions during this cruise are described in Figure 2-2, 2-3 and 2-4. SST and specific humidity showed significant variance as ship crossing the front, while air temperature and humidity overall responded with a delay, with an exception on Jan. 22, when air temperature dropped as the ship crossed into the warm side of the front. A strong difference between SST and air temperature reaching 15°C was observed on the warm side of the Gulf Stream, while on the cold side the air temperature was still lower but much closer to underlying SST. Air pressure was much less sensitive to the local SST change and adjusted on a more synoptic scale. Strong winds of 15 m/s were
common during the cruise, and an extreme of 25 m/s wind was observed, while surface stress reflected the surface wind. Wintertime cold, dry air in this region induced strong net heat loss from the ocean, leading to overall positive buoyancy and latent heat fluxes. On Jan.27 a typical strong cold-air outbreak occurred, characterized by the intrusion of much colder and drier air on the warm side of the front, and a sharp increase in latent heat flux of about 500 W/m² was observed, together with buoyancy flux reaching 400 W/m². On Jan.22, another outbreak was observed, with a similar increase in latent heat flux but a smaller increase in buoyancy flux. The directions of the background wind during the cruise varied, but in general blew from the cold side of the front to the warm, or parallel to the front.

One major advantage of this data set is the relatively static shape of the SST front during the experiment and ship’s almost perpendicular crossings, which makes it less ambiguous when calculating the relative distance between the SST front and the measurements as discussed in the Methods section.

![Figure 2-1: Schematic of the Pilot cruise ship track (red) between Jan.18 and Jan.31, 2006, with a time-averaged SST map obtained from GHRSSST-L4. The color scale for the SST ranges from 15 °C (dark blue) to 25 °C (yellow). The cruise started at the north-westernmost point. Rawinsonde launches are marked as blue.](image-url)
Figure 2-2: (a) Time series of shipboard air temperature $T_{air}$ and near-surface ocean temperature $T_{sea}$. Yday represents the day of year. (b) Sea level air pressure time series measured from R/V Atlantis. (c) Time series of 5-min averaged buoyancy flux from ship measurements. Bulk estimates and direct covariance(DC) of surface fluxes are shown in blue and orange. Positive flux represents cooling of the sea surface.
Figure 2-3: (a) Time series of surface wind measurements from R/V Atlantis. (b) Time series of specific humidity of the sea surface $Q_{sea}$ and air above, $Q_{air}$, from R/V Atlantis. (c) Time series of 5-min averaged latent heat flux from ship measurements.
CLIMODE Knorr Cruise

Two legs of the CLIMODE main experiment, conducted on R/V Knorr, were between Feb.7–25, and Mar.3–20, 2007, respectively. Bulk and direct covariance flux packages provided similar surface flux estimates as during the Pilot cruise. Overall, the wind-ship alignments were less ideal than during the Pilot cruise and resulted in a number of gaps and noise in the surface flux time series. 143 Rawinsondes launches were made during these two legs, with most launched near the SST front during front-crossing events. Note that the SST front was much less static compared to the Pilot case. Meanders of the front moved on a time scale on the order of one day resulting in ambiguous estimates of the distance from the front. Over the duration of these two legs, SST varied considerably during the cruise, with the lowest SST reaching about 5°C on the cold side of the front. Air temperature varied more and was less correlated with underlying SST compared to the Pilot case. A strong cold-air outbreak was observed on day 65 (Mar.6) and sub-zero air temperature was observed, in phase with an increase in local air pressure. The sizeable difference between SST and air induced a sharp increase in the surface buoyancy flux exceeding 600 W/m², a typical value observed in wintertime (Marshall et al. 2009).

General wind conditions were strong, with 15 m/s winds encountered frequently on either side of the front, and a maximum wind speed exceeding 25 m/s near the surface was also observed. Background wind directions were diverse, with most blowing from cold to warm or parallel, and a few cases where warm-to-cold wind was observed. The specific humidity of the air showed a similar level of coherence with the sea surface level humidity. Again, note the effect of the cold-air outbreak on the intrusion of drier air and increased latent heat flux.

This data set comprises the largest number of shipboard measurements and sounding launches among the three experiments, however the more complicated front dynamics and the less ideal front-crossing angles induce a higher level of noise in the analysis. Methods to compensate for these disadvantages will be discussed in the next section.
Figure 2-4: Schematic of two legs of Knorr cruise ship track (red and light blue) in February and March 2007, with averaged SST map obtained from GHR SST-L4. Rawinsonde launches are marked as dark blue. Colorscale unit in °C.
Figure 2-5: (a) Time series of shipboard air temperature $T_{\text{air}}$ and near-surface ocean temperature $T_{\text{sea}}$. (b) Air pressure time series measured from R/V Knorr. (c) Time series of 5-min averaged buoyancy flux from ship measurements. Positive flux represents cooling of the sea surface.
Processes Driving Exchange at Cape Hatteras (PEACH)

The PEACH project was focused on the processes govern the exchange of waters between the US eastern shelves and the open ocean. Major forcings in this region, including air-sea interactions and the influence of the Gulf Stream on shelf/ocean exchange, were investigated during this project with extensive use of ship measurements, buoys moorings and numerical models (Seim et al, 2022). The primary dataset this study focuses on is from the last cruise of the PEACH project, conducted between Nov. 16 and Nov. 28, 2018, on R/V Neil Armstrong. During this cruise, flux packages similar to those in CLIMODE were used to provide surface heat flux measurements, and a total of 10
soundings were launched during an almost-front-perpendicular (80 degrees) crossing event on Nov. 24 (J. Zambon, personal communication, 2020). While fewer soundings were launched compared to CLIMODE, the background wind condition was fairly stable throughout the observing window, consistently blowing from the warm side of the front to the cold side, providing a rare opportunity to observe the evolution of the MABL under such conditions, as cold-to-warm wind is more prevalent in this region due to the overall geostrophic wind condition. The overall wind condition was strong, with on average 12 m/s wind throughout the cruise. A 30 m/s maximum was observed on the night of Nov. 24, which is assumed to be due to a cyclogenesis event near the coast. The continuously dropping air pressure during the cruise showed the effect of a low-pressure system passing through the study region. Note that the sharp increase in the SST time series was caused by the ship re-entering the warm side for a short period, and a significant change in buoyancy flux and latent heat flux can be seen in Figure 2-8 (c) and 2-9 (c). Outside the observing window with sounding launches, there was a major cold-air outbreak event during Nov. 22-23, marked by a sharp decrease in the air temperature, with buoyancy flux and latent heat flux reaching 400 W/m² and 900 W/m², respectively.

While the observation window was shorter, this dataset provides a comprehensive observation of the formation of a stable boundary layer (SBL) on the cold side of the front when the wind was blowing from warm to cold, which was not seen in the CLIMODE data.
Figure 2-7: Schematic of sounding launch locations (red) on Nov.24, 2018, with underlying SST map from GHRSSST-L4. Color scale unit in °C.
Figure 2.8: (a) Time series of shipboard air temperature $T_{\text{air}}$ and near-surface ocean temperature $T_{\text{sea}}$. (b) Air pressure time series measured from R/V Neil Armstrong. (c) Time series of 5-min averaged buoyancy flux from ship measurements. Positive flux represents cooling of the sea surface.
2.3 Methods

SST map and front identification

24-hour averaged SST maps are generated from Group for High Resolution Sea Surface Temperature (GHRSSST) L4 sea surface temperature analysis, with a spatial resolution of 0.01 degree. While shipboard measurements provide high-resolution time series of SST, ship tracks in all three cruises were not designed to follow the SST front closely, and thus...
SST measurements from the ship cannot provide comprehensive insights on the spatial distribution of the SST front. With satellite-based observations providing a spatial context for cruise transects, the overall SST front structure in the region of interest can be identified (Figure 2-10 (a)). The shipboard SST measurements are used to study specific front-crossing transect, where higher temporal resolution is needed.

To identify the SST front, satellite-based SST maps are filtered first with an SST threshold of 16 °C, which was the common temperature of cold shelf water during all three cruises, to distinguish roughly the cold water from the Gulf Stream. Filtered maps (e.g., Figure 2-10b) then are processed to identify the region where the sharpest SST gradient exists. These regions (e.g., the outline of the lighter region in Figure 2-10b) are then connected to form a continuous line which represents the defined SST front (Figure 2-10 (b)).

For the CLIMODE Pilot cruise, the SST front structure was almost static during the cruise. Very few major meanders formed, and the front shape varied insignificantly (e.g., rings of the meander moving more than 50km per day). For this reason, a 14-day average SST map is used in the analysis. For the PEACH cruise, the SST front was also relatively stable (general shape of the SST didn’t not change by more than 10%), and thus a 12-day average SST map is used.

During the CLIMODE Knorr cruise, the SST front was far more energetic, showing a significant movement (>50 km) toward the east on a daily basis. In addition, numerous meanders were formed over the course of the cruise, which further complicated the structure of the front. As a result, daily averaged SST maps are used for this cruise. Larger errors in defining the SST front are also likely induced in this case due to poorly defined boundaries caused by meanders.
Figure 2-10: (a) 14-day averaged SST map from GHR SST-L4 for CLIMODE Pilot cruise, resolution 0.01° × 0.01°, Colorscale unit in °C. (b) Calculated SST front (light blue) location based on (a).

**Distance from the front**

The distance from the front serves as an important attribute in this study and is determined by combining the shipboard GPS data and the defined SST front map described in the previous section. This value overall, by definition, shows a significant correlation with SST, comparing Figure 2-11 and Figure 2-2 (a). With a continuous line of SST front, the distance from the front is defined as the shortest distance between the ship location, and the line. Positive values indicate that the ship was on the warm side, and negative value the cold side of the front. In most cases, this distance accurately represents the amount of influence that a local observation was receiving from the SST.
front. In cases where the front shape is straight, this distance is analogous to the fetch of the cross-frontal wind. In cases where the point is surrounded by a U-shape front or near a meander, the physical implication of this distance as defined above is less obvious is considered on a case-by-case basis.

This distance, as described in the Introduction section, is used in several quantitative analyses. In the cold-to-warm cases, the estimation of the entrainment rate (Eq.2.15) and fetch-dependence boundary layer height growth (Eq.2.21) both require a properly defined distance from the front. In the warm-to-cold case, the SBL growth rate can be estimated with this distance and Eq.2.22. In the parallel wind case, this value serves to qualitatively indicate the amount of frontal influence on the observation.

**Background wind direction and magnitude**

For the observations from each cruise, the background wind direction is determined by identifying the angle between the drifting direction of the launched sounding balloon over the first 1000 m and the SST line (Figure 2-12). When the angle is greater than 45 degrees, the wind is identified as a cross-frontal wind, and with an angle smaller than 45 degrees as a parallel wind. The major advantage of using the balloon-based direction instead of surface ship measurement is that it better represents the overall background geostrophic wind within the boundary layer and is less noisy. The magnitude of the background wind is estimated by averaging the vertical wind velocity profiles from soundings.
Figure 2-11: The calculated distance from the front time series for CLIMODE Pilot Cruise (blue line), gray reference line represents the SST front.
Categorize soundings into groups

One major characteristic of the background wind in this region is the high temporal variability in wind direction, based on the sounding measurements retrieved from CLIMODE and PEACH. Based on observation, it is rare for the background wind to blow consistently in one direction for more than one day. Thus, it is necessary to separate soundings into groups, based on the consistency of the background wind, to examine characteristics of the MABL under different wind regimes. Continuous sounding launches in the vicinity of the front are first grouped into the same launch series. The background wind direction is then used to filter the group further. If the wind is blowing in one direction consistently, all soundings in the group are kept; if wind changed direction at the beginning or the end of the group, that specific sounding is dropped; in

Figure 2-12: The launch location (orange dot) and trajectory during the first 5km (blue) of ascension of sounding 15 from CLIMODE Pilot. Based on its relative angle with the SST front (light blue), the background wind in this case is identified as blowing parallel to the front.
cases where the wind varied significantly throughout out the series of launches, the entire
group of soundings is not considered, as the changing wind regime further complicates
the evolution of the boundary layer and is beyond the scope of this study.

By dividing soundings into groups, launches from three cruises are filtered and
categorized into seven transects, with three cold-to-warm cases, three parallel wind cases
and one warm-to-cold case. Analysis of each individual transect is addressed in later
sections, respectively.

**Boundary layer height estimation**

The height of the boundary layer is determined by identifying the inversion height if the
MABL is overall convective. The height of the surface stable layer is used instead if a
stable boundary layer is studied. The top boundary of a convective MABL is marked by a
sharp transition from vertically well-mixed potential virtual temperature to a profile in
which potential virtual temperature increases with height. This inversion location is often
coupled with a sharp decrease in specific humidity, as the upper layer typically contains
drier air. Vertical profiles of humidity and temperature from sounding launches are
examined to visually identify the inversion point, as shown in Figure 2-13.

In the case where wind is blowing from warm-to-cold and a SBL is formed, the spatial
resolution of the soundings is insufficient to capture the surface stable layer with enough
vertical resolution. The ascension speed of balloons was roughly 5 m/s, with instruments’
sampling rate at 1 Hz. While a newly formed SBL can have heights of 20-30 m, the
resolution of the vertical temperature profile may lead to considerable errors (over 50%) in
the estimation of the SBL height. The error can be smaller in a more-developed SBL,
where the top of the surface layer can reach 200 m. It is advised that in future studies,
temperature profiling with higher vertical spatial resolution is needed for SBL cases.
**Synoptic scale system**

ERA5, the fifth generation of the European Centre for Medium-Range Weather Forecasts (ECMWF) atmospheric reanalyses of the global climate (Hersbach, Hans, et al. 2020), is used here to provide information on the synoptic pressure system during each cruise. This data set has a 31 km spatial resolution and a one-hour temporal resolution. Spatially averaged sea-level-pressure (SLP) is used to provide the temporal evolution of the synoptic pressure system, e.g., atmospheric tide, and 24-hour averaged SLP is used to provide the spatial distribution of the synoptic pressure system. Shipboard air pressure measurements are compared here with reanalyses to study the local SST’s effect on air pressure.

Figure 2-13: From left to right: relative humidity, specific humidity, virtual potential temperature and wind speed profile of the first 5000 m from sounding 15 from CLIMODE Pilot. Note the sharp decrease in humidity and the inversion in virtual temperature, which mark the top of the boundary layer at 450 m.
2.4 Results and Discussion

2.4.1 Cold to Warm Case

When the dominant background wind is blowing from the cold side of the SST front to the warm side, cooler air is advected over a relatively warmer surface. The significant air-sea temperature difference leads to increased surface heat flux. Intense heating and evaporation enhance the heat and moisture exchange between the air-sea interface. A highly turbulent surface boundary layer provides energy for deeper vertical mixing, mainly supported by raising thermals. In this case a significant MABL height growth is often observed. Surface winds often accelerate, due to the mixing-down of high momentum air from above (O’Neil 2010). At the top of the boundary layer, strato-cumulus is often formed, and provides feedback to the ocean surface by adjusting the albedo. The radiative cooling at the cloud top can also be important in controlling the entrainment rate, sometimes contributes as much as half of the surface flux (de Szoede 2005).

On a longer time scale of O(1 month), the pressure system induced by SST heterogeneity is also found to affect the cross-frontal flow and adjust the wind profile inside the boundary layer (Plagge 2016). However, whether the thermally induced pressure gradient is able to affect the local flow on shorter time scales of one day is less clear. In this section, results from CLIMODE Pilot and Knorr are discussed to examine the response of the MABL to the SST front on such a time scale.

**Transects from the CLIMODE Knorr cruise**

Within the CLIMODE main campaign during February and March 2007, two transects were identified as cases where the background wind was continuously blowing from the cold side of the front to the warm side. These two transects, referred as KnCW1 and KnCW2, were 21 hours long (starting at 2AM UTC, Mar.16) and 17 hours long (starting at 12PM, Mar.12), respectively. During these two transects, the wind direction estimated from the sounding launches was overall towards the south, which led to a strong cross-
frontal advection, given that the overall SST front was aligned east-west. Observed boundary layer depth-averaged wind speed was consistently around 9-11 m/s during KnCW1, while during KnCW2 a maximum of 16 m/s was observed during the first half of the transect, and wind speed gradually decreased over time to a minimum of 6 m/s at the end of the transect. The shipboard wind measurements were noisy during KnCW2 due to poor ship-wind alignment, and buoyancy flux estimation was not available during most of the time.

The overall boundary layer conditions in these two transects were as expected: cold air advected to the warm region induced a significant air-sea temperature difference and thus induced strong surface buoyancy flux, over 100 W/m² once the ship crossed the front in KnCW1, and near 300 W/m² in KnCW2 at where the largest air-sea temperature difference was observed. Stability parameter $z/L$ changed sign from overall stable to unstable on the warm side, consistent with what was often observed in previous studies and model simulations (Friehe 1991; Rouault 2000; Skyllingstad 2007). Significant boundary layer height growth was observed in both cases, with over 400 m in KnCW1 and over 600 m in KnCW2. It is worth noting that in KnCW1, a shallow surface stable boundary layer was observed on the cold side. While the overall wind was blowing from cold to warm, given the observed air-sea temperature difference, we believe this SBL was caused by the relatively warm air advected from a warm meander in the north, crossing a filament of cooler sea surface. The formation of such a SBL in response to a small-scale SST feature is discussed in more detail in a later section 2.4.3, where we consider the cases of warm air blowing over cooler sea.
Figure 2-14: Ship track (blue) and SST front (red) superimposed on the synoptic surface pressure retrieved from ERA5 for KnCW1. Colorscale unit in mbar.
Figure 2-15: Same as Figure 2-14 but for KnCW2.
Figure 2-16: (a) MABL height estimated from sounding profiles (b) MABL height vs fetch using Eq. 2.21. (c), (d), (e) $T_{sea}$, $T_{air}$ and air-sea temperature difference, respectively, each retrieved from shipboard measurements.
Figure 2-17: (a) Shipboard surface wind measurements. (b) Surface stress estimated with COARE3.5 algorithm. (c) Direct covariance buoyancy flux from ship measurements. (d) Stability $z/L$ parameter at 10 m estimated with COARE3.5 algorithm. (e) Shipboard surface pressure perturbation $P_{air}$. 
In both transects, an increase in wind speed was observed in the vicinity of the warm side of the front (0-20 km in Figure 2-17 (a)). An increase of 2.5 m/s was seen in KnCW1, while a weaker increase of 0.5 m/s in KnCW2 was observed. The increase in wind speed in a convective boundary layer has been frequently observed and discussed in previous studies (Hayes 1989, Chelton 2001, Mahrt 2004, Spall 2007, Small 2008, O’Neil 2010, 2012, Kilpatrick 2014). Two major mechanisms discussed in the introduction section are the pressure adjustment and vertical mixing. These previous studies have shown good agreement on the importance of these two mechanisms on a time scale of 15 days to one month. Yet, as Spall (2007) and Kilpatrick (2014) argued, the adjustment time scales for these two mechanisms are different, depending on the cross-frontal wind strength and spatial scale. They proposed that:

\[
L_M = \frac{UD_M^2}{K_M}, \quad L_P = \frac{UD_P^2}{K_P},
\]

where \( U \) is the cross-frontal wind velocity, \( L_M \) and \( L_P \) are the adjustment length scales for mixing and pressure, \( D_M, D_P \) are the depth scales, and \( K_M, K_P \) are the vertical mixing coefficients, respectively. They argued that the mixing coefficients are of the same order, and the major difference between the two length scales is due to the cross-front wind related depth scale. In a strong cross-frontal wind case, where \( D_M \ll D_P \) and \( L_M \ll L_P \), an internal boundary layer in which the vertical mixing mechanism is dominant is likely to form. In contrast, for a weak wind case, \( L_M \sim L_P \), the SST induced pressure gradient can have comparable effect on the boundary layer. In the following section, this hypothesis is tested with observation and model analysis, focusing mainly on comparing the effect of these mechanism on the boundary growth and wind adjustment.

To evaluate the pressure induced effect, Eq.2.13, shown again below, is used:

\[
P_t = \frac{g}{\Theta} h \frac{\partial[\theta]}{\partial x},
\]

where \( h \) is the boundary layer height identified from sounding profiles, \([\theta] \) the depth averaged virtual potential temperature and \( \Theta \) a reference temperature. The thermally induced pressure gradient in both transects then is estimated and shown in Figure 2-18.
The magnitude of the pressure gradient term \((10^{-3} \text{ m/s}^2)\) is consistent with Mahrt (2004) and Kawai (2014), and the maximum magnitude is near the front where the largest air-sea temperature exists, which can lead to an increase in wind speed seen in Figure 2-17 (a). In Kawai (2014), it has been shown in their observation that the thermally induced pressure gradient tends to be negative (positive) where SST decreases (increases), which in general agrees with our \(P_t\) estimation. They suggested that such gradient can lead to the formation of sea-breeze wind, enhancing the strength of surface wind. Due to the limited sample size and spatial coverage of sounding launches, most terms in the momentum budget cannot be quantitatively estimated, but based on typical values from previous studies (e.g., Kawai 2014; O’Neil 2010; Kilpatrick 2014), it is reasonable to assume that the magnitude of estimated \(P_t\) shown here is comparable or greater than those of the Coriolis term \((O(10^{-3} \text{ m/s}^2))\) and advection terms \((O(10^{-4} \text{ m/s}^2))\) often seen in mid-latitude regions. If we assume that the pressure gradient was relatively static around the location where the observation was made, estimated \(P_t\) can contribute to observed wind change, especially near the SST front where the \(P_t\) is the greatest due to sharp SST change. Away from the front, where the thermally induced pressure gradient is smaller \((0.3-0.5 \times 10^{-3} \text{ m/s}^2)\), wind speed also varies less as expected for a region where the pressure forcing is small. It is worth noting that, while the sign of the pressure gradient generally changes in concert with that of the observed wind change, given the rather complicated dynamics inside the MABL, it is insufficient to quantitatively determine the direct correlation between wind change and estimated \(P_t\). In addition, observations in Kawai (2014) were mostly made during weak background wind (5 m/s), while in this case the background wind exceeded 10 m/s. Note that as discussed in Spall (2007) and Small (2008), effect of such pressure gradients is more prominent in a weak-wind scenario where the adjustment time scale for the pressure gradient is comparable to other adjustment dynamics. In this strong-wind case, however, the magnitude of \(P_t\) is significantly smaller than those in a weak-wind case (shown in later sections), which can be assumed to have less impact on the observed wind change. In conclusion, thermally induced pressure gradient in this case can be responsible for part of the wind adjustment, yet its strength in this case is not dominant.
Figure 2-18: Thermally induced pressure gradient at each sounding launch location. Positive means higher pressure on the warm side.

To examine the effect of the vertical mixing mechanism, the entrainment rate of the MABL is estimated with Eq.2.15, shown again below:

\[ U \frac{\partial z_i}{\partial x} = w_e + w_L. \]

The subsidence rate during the cruise is estimated with The Modern-Era Retrospective analysis for Research and Applications, version 2 (MERRA-2) pressure velocity at 850 hPa, and has a magnitude of around 1mm/s. The boundary layer height is retrieved from sounding observations, while cross-frontal wind velocity is derived from a depth-averaged sounding profile, and the entrainment rate is calculated as a residual. It is worth noting that while the wind was generally blowing from cold to warm, directly using the wind velocity would greatly overestimate the entrainment rate (over 50%). This is due to the fact that the wind direction showed considerable variation during the transect, and the true cross-frontal component of the wind was less than the observed wind speed. The
complicated shape of the front during the Knorr cruise further decreased the magnitude of the true average cross-frontal wind velocity. To estimate the observed entrainment correctly, depth-averaged wind velocity is multiplied by a factor of 0.4-0.7 to better estimate $U$ used in Eq.2.15, based on observed wind direction variance.

The contribution to entrainment from surface buoyancy flux is estimated with Eq.2.16, shown again below:

$$W_e = \frac{Aw^3}{h\Delta b}$$

de Szoeke (2005) used $A = 0.2$ and found the surface buoyancy flux contributed only a small fraction to the observed entrainment. Other studies (Pino et al. 2003, Conzemius et al. 2006) suggested that $A$ around 0.32 gives a better representation of the surface buoyancy flux force. In this study, we used $A = 0.25-0.35$ to estimate the surface buoyancy driven entrainment, and this range is reflected by the error bar in Figure 2-19. To take the shear-driven entrainment into account, Tennekes (1973) proposed that, in a purely shear-driven boundary layer, the entrainment rate can be estimated as:

$$\frac{dh}{dt} = \frac{2.5\Theta_0 u_*^3}{g\Delta \theta v h}$$  (2.24)

where $\Theta_0$ is the initial average temperature across the boundary layer. While during the observation the boundary layer is not purely shear-driven, this rate should still be a valid reference for shear contribution, and is shown in Figure 2-19 and 2-20 (orange).

Another important component of entrainment is the cloud top radiative cooling that induces boundary layer top heat flux, which can act in a manner similar to the surface flux. Previous studies (de Szoeke 2005, Ghate et al. 2014, Zheng et al. 2018) have shown the importance of cloud top radiative cooling in the total heat budget and entrainment. In this study, due to the lack of dedicated cloud measurements, a rough estimation of the cloud top radiative cooling flux based on overall conditions and values from previous studies is used. In future studies, it is suggested to estimate more accurate cloud top flux with combined measurements from soundings and satellite with the Rapid Radiative Transfer Model (RRTM). Of the ten soundings that were launched on the warm side
during the two transects, all of them observed boundary layer top cloud formations, where relative humidity reached over 90%, and a sharp decrease beyond inversion. With heavy cloud cover at the top of the MABL, a 30-40 W/m² heat flux is reasonable to be assumed based on previous studies (e.g., Zheng et al. 2018) and observations, and the entrainment cause by such a flux is estimated in Figure 2-19 as yellow points.

Figure 2-19: Surface buoyancy driven entrainment with Eq.2.16 (blue), shear- driven entrainment with Eq.2.24 (orange), cloud top radiative cooling induced entrainment with Eq.2.16 (yellow), and total summation of three (purple) for KnCW1. Dash lines mark the range of the total entrainment estimated from observation.
The cloud top cooling entrainment estimated in KnCW1 at 200 km is uncommonly high due to the relatively weak capping inversion (~1.5K) combined with the assumption of roughly the same amount of flux, the latter of which is a gross overestimation. In KnCW2, entrainment at the 40-140 km region cannot be estimated due to the absence of surface flux measurements. Based on the consistently high air-sea temperature difference observed in KnCW2, it is reasonable to assume that the high entrainment rate (14-17 mm/s) near the front would be lower but overall persistent in this region. Neglecting one outlier, we can see in both transects that the entrainment is greater near the front, consistent with regions where the highest buoyancy fluxes were observed due to the high air-sea temperature difference. The total entrainment is 5-10 mm/s in KnCW1, and 4-16 mm/s in KnCW2. Compared to entrainment rate estimated with Eq.2.15 from observed wind and boundary layer growth, which are 9-14 mm/s and 12-20 mm/s in KnCW1 and KnCW2, respectively, the surface buoyancy-shear-cloud top combined estimation of the entrainment is the primary contributor to the boundary layer growth.

Compared to the pressure gradient adjustment mechanism, the IBL (internal boundary layer)-like physics, which mainly reacts to surface buoyancy, shear and boundary layer...
top cooling, is more prominent in these two transects. If we assume that the vertical-mixing mechanism brings down high momentum wind from the upper atmosphere, as previous studies have found (e.g. Hayes 1989, Song 2004, Small 2008, O’Neil 2010), then the usual vertical adjustment time scale in a convective boundary layer similar to these two transects, which can be as short as 15 minutes (Stull 1988), can lead to a sharp increase in wind speed at the surface in the region where the entrainment is largest. This is what we observed in KnCW1 and KnCW2, where wind accelerated as the ship entered the warm region where cold air was rapidly advected over warm sea surface.

**Transect from the CLIMODE Pilot cruise**

One transect during the CLIMODE Pilot cruise is identified as a predominantly cold-to-warm background wind case, referred as PiCW1. This transect started on 12PM UTC Jan. 27 2006 and ended on 3PM UTC Jan. 28, with four sounding launches made along the course. An ASIS (Air–Sea Interaction Spar) buoy was also deployed to take simultaneous measurements of key air-sea quantities near the front. It is worth noting that this 1.2 day long transect took place right after a major cold air outbreak observed earlier on Jan. 27, and as a result, the major characteristics of the boundary layer were unique compared to cases during the Knorr cruise. The air was abnormally dry and cool on the warm side of the front, and led to surface flux reaching 400 W/m². Air-sea temperature difference was significantly high, continuously over 9 °C as going deeper into the warm region. Depth-averaged wind speed over the course of the transect was consistently greater than 16 m/s blowing from cold to warm. Such high surface buoyant forcing led to a significant increase in boundary layer height; MABL height was observed to increase by more than 1700 m. The highly turbulent environment led to significant cloud formation at the top of the boundary layer, with four soundings all showing high relative humidity at the inversion base region, a typical sign for stratocumulus.
Figure 2-21: Same as Figure 2-14 but for PiCW1.
Figure 2-22: Same as Figure 2-16 but for PiCW1.
Figure 2-23: Same as Figure 2-17 but for PiCW1, the temporal air pressure trend was removed with ASIS pressure measurements.
The observed surface wind velocity and wind stress were constantly high over the warm region, with a peak where the highest buoyancy flux was observed. They show very little covariation with stability parameter $z/L$, however, suggesting that the wind modification from stability induced surface stress change described in Eq.2.9 was weak in this case. With Eq.2.13, the thermally induced pressure gradient is shown in Figure 2-24. Compared to Knorr cases, the thermally induced pressure gradient was nearly one order of magnitude larger ($8 \times 10^{-3} \text{ m/s}^2$ vs. $10^{-3} \text{ m/s}^2$). If such a pressure gradient was purely spatial, then a wind velocity adjustment much larger than what is seen in Figure 2-23 (a) would be observed. It is not clear why such a high thermal pressure gradient is produced from Eq.2.13, yet there are two potential explanations. The first is that a temporal change, rather than a spatial change in air temperature, was observed in the shipboard data. In Figure 2-22 (d), the air temperature was recovering to normal value at the end of the transect, which was near the front, based on larger scale air temperature maps. Such a recovery after a major outbreak can be synoptic rather than directly affected by the sea temperature underneath, and thus the dependence on distance from the front would be invalid. The second possibility is that the boundary layer height dependence in Eq.2.16 does not apply in this case, where an extremely large increase was observed. In previous studies (Mahrt 2004, Kawai 2014), much smoother MABL increases were studied; in those cases, a 100 m increase in a 1000 m high boundary layer only contributed to a 10% difference in pressure. In PiCW1, however, with an overall 1700 m increase over four sounding launches and lowest boundary layer height observed at 600 m, the increase in $h$ would have a large relative influence on the magnitude of calculated thermally induced pressure. Small (2008) pointed out that a significant increase in boundary layer height would have a non-negligible impact on the pressure, as the integrated air mass would differ due to the distinct difference between boundary layer air and air above inversion. We believe that in this case factors other than the heterogeneity of SST dominate the air pressure change.
The components of entrainment calculated with Eq.2.16 and 21 are shown in Figure 2-25. Given the prevalent cloud coverage observed from sounding profiles, we again assumed a cloud top radiative cool flux of 30-40 W/m². The sudden drop in entrainment rate at the second sounding is due to an inversion jump twice the average, but overall, the entrainment rate is consistent. Compared to the entrainment rate calculated from observed boundary layer height growth and cross-frontal wind speed, which is 68-136 mm/s depending on the factor used in Eq.2.15, the estimated entrainment due to IBL-like physics is again contributing the most to the observed entrainment. There are three major differences between the PiCW and KnCW cases. First, the contribution of surface buoyancy flux is much higher in PiCW1 due to the large air-sea temperature difference created by the earlier cold air outbreak. Surface buoyancy flux was only relatively high near the front in KnCW, and the air soon reached equilibrium, leading to a decreased entrainment and slower boundary layer growth further away from the front, while in PiCW the air was consistently cold even over warmer sea surface, causing a much higher buoyancy flux which led to heavy entrainment and significant boundary layer growth.
deep into the warm region. Secondly, the shear driven entrainment is more effective in PiCW1, contributing as much as 30% of the buoyancy driven entrainment. The consistent high wind (~16 m/s) observed during PiCW effectively increased the friction velocity $u_*$, providing wind shear at the surface to generate turbulent kinetic energy. Thirdly, cloud top radiative cooling contributed less in PiCW. This is likely due to an underestimation of the cloud top flux due to the rough approach we used. A 35 W/m$^2$ flux is comparable to the surface buoyancy flux of 100-250 W/m$^2$ in KnCW, but such flux would be small compared to observed 400 W/m$^2$ in PiCW. Previous studies (e.g., Ghate 2014) have reported over 100 W/m$^2$ cloud top flux in a strongly convective boundary layer. Further analysis of the data with RRTM would help to determine if the cloud top flux needs to be reconsidered in PiCW.

![Figure 2-25: Same as Figure 2-19 but for PiCW1.](image)

Overall, the transect from PiCW, though under different synoptic condition, supports the findings in KnCW1 and KnCW 2. On a time scale of one day with strong cross frontal wind (8-16 m/s), the boundary layer is IBL-physics dominant, where the vertical mixing mechanism provides the majority of the boundary layer growth. The pressure adjustment
mechanism may have affected the wind field, yet it is unclear whether the observed wind modification is directly related to the pressure anomaly induced by the SST heterogeneity, as the validity of $P_t$ in a case where MABL height increases dramatically is questionable.

2.4.2 Parallel Wind Case

When the background wind is blowing predominantly parallel (defined as relative angle smaller than 45º) to the SST front, the cross-frontal component of the wind is weak compared to that considered in the previous cases. In this case, it is expected that air-sea temperature difference on either side of the SST front would be smaller due to the relatively weak cold to warm advection. In the vicinity of the front, since a small portion of air is transported across the front, the boundary layer can still be weakly convective depending on the magnitude of the cross-frontal transport. Farther away from the front, weak cross-frontal wind can no longer bring air into this region, and as a result, it is expected that the boundary layer would be quasi-stationary.

Another important difference compared to the strong cross-frontal wind case is expected in the observed adjustment scale for the vertical mixing mechanism and the pressure gradient mechanism. In a weak cross-frontal wind scenario, the adjustment length scale for pressure $L_p$ can be on the same order as the vertical mixing length scale $L_M$ (Spall 2007, Kilpatrick 2014). In this case the pressure adjustment mechanism can be important on a time scale of one day. Three transects from CLIMODE Knorr and the Pilot cruise are analyzed in the following sections to examine the contribution of these two mechanisms to the boundary layer evolution.
Transect from CLIMODE Knorr cruise

One transect (later referred as KnPL1) from the Knorr cruise has wind blowing predominantly parallel to the SST front. This transect started on 12PM Mar. 13 and ended on 3PM Mar. 14. During this transect, no major synoptic weather system other than the surface high pressure shown in Figure 2-26 was observed. It is worth noting that although this transect was identified as a parallel wind case, the observed wind direction was slightly towards to the warm side, with an angle no greater than 30 degrees. The cross-frontal wind component was 2-3 m/s, compared to the observed true depth-averaged wind speed of 8-11 m/s.

The overall boundary layer condition during KnPL in the vicinity of the SST front is similar to KnCW, as there was net cold-to-warm air transport near the front. However, the major difference is that the boundary layer growth was confined within the first 60km of the warm side, where in KnCW and PiCW the boundary layer height also increased in the region further away from the front. In KnPL, further away from the front, the boundary layer height was quasi static, consistent with our hypothesis that weak cross-frontal wind transport could not bring the influence of the SST front deeper into the warm region. The observed buoyancy flux reached its maximum of 200 W/m² at 25 km, where the largest wind speed and air-sea temperature difference were also observed. Wind speed was consistently high near the front, reaching a maximum of 13 m/s, and decreased by 3.5 m/s beyond 75 km into the warm side, possibly because the boundary layer reached equilibrium in this region. It is worth noting that the stability parameter was still positive in the region where wind decelerated, suggesting that the wind modification from stability induced surface stress change was relatively small. The observed air pressure anomaly is likely due to the spatial synoptic pressure system, as the ship track was along the pressure gradient shown in Figure 2-26.
Figure 2-26: Same as Figure 2-14 but for KnPL1
Figure 2-27: Same as Figure 2-16 but for KnPL1.
The thermally induced pressure gradient for KnPL1 is shown in Figure 2-29. Compared to KnCW, the magnitude of the pressure gradient is on average two times higher (1-2.5 \times 10^{-3} \text{ m/s}^2 versus 0.5-1 \times 10^{-3} \text{ m/s}^2), consistent with Spall 2007’s argument that weak background wind enables stronger thermal pressure adjustment. The major difference
between KnPL and KnCW is that, for a weak cross-frontal wind case, the pressure adjustment length scale $L_P$ is shorter, and the SST induced pressure gradient is more likely to have a more significant effect on the boundary layer. Note that the largest pressure gradient was observed where the wind changed significantly. With the magnitude of the estimated pressure gradient, and assuming that this gradient was persistent within 25 km from the location of the sounding, the wind modification from the pressure term can lead to a net change of more than 5 m/s in surface wind speed. This is, however, not a direct evidence of pressure gradient forcing driving wind change, as such a correlation is not seen everywhere during this transect, and the surface stress shown in Figure 2-28 (b) does not agree well with high $P_t$ in the region. Given the time gap between each sounding launch, the observed wind change can be a result of a combination of various forcing on different scales, and thus explaining the disagreement between dynamics. In this case, however, it is safe to argue that $P_t$ is quantitatively much stronger, despite that its correlation with observed wind change remain qualitative.

Figure 2-29: Same as Figure 2-16 but for KnPL1.
Based on sounding observations, stratocumulus was assumed to be prevalent during this transect on the warm side of the front, as the boundary layer was still overall convective. Thus, we still assume a typical 30-40 W/m² cloud top radiative cooling flux when calculating the entrainment. For the observed boundary layer entrainment, an alternative method, Eq. 2.25, is used to compare with values obtained from Eq.2.15:

\[
\frac{\partial z_i}{\partial t} = w_e + w_L. 
\]  

(2.25)

The major difference between Eq.2.25 and Eq.2.15 is that Eq.2.25 attempts to estimate boundary layer growth based on time elapsed between each observation at different locations, assuming that the change in boundary layer height is purely temporal and no advection is involved. Such an assumption is obviously not valid in a strong cross-frontal wind case. However, in KnPL1, it is reasonable to assume that advection is relatively weak due to the almost parallel wind. The entrainment estimated based on Eq.2.25 for KnPL is 9-11 mm/s, which is close to the value from Eq.2.15, which is 8-14 mm/s. The total entrainment from IBL-like physics, including surface buoyancy, shear, and cloud top cooling, is shown in Figure 2-30.

![Figure 2-30: Same as Figure 2-19 but for KnPL1.](image-url)
We can see that the total entrainment from the model is again close to the observed values. The entrainment is highest near the front, with the majority of the entrainment coming from surface buoyancy flux. This is consistent with the observed rapid boundary layer growth (over 300 m) within the first 60 km on the warm side of the front. Further away from the front, the entrainment rate decreased to around 5 mm/s. This value can also include some error from the overestimation of the cloud top cooling in the near equilibrium region. Together with the estimated subsidence rate of 1-2 mm/s, the estimated boundary layer height adjustment is close to the observed quasi-static MABL further away from the front in Figure 2-27 (a), (b). On the cold side of the front where surface buoyancy flux was small, the shear-driven entrainment and cloud top cooling together still provided a strong entrainment of 6-8 mm/s, which is consistent with the MABL height growth at -20 km to 20 km, implying the importance of shear-driven entrainment, which has been often neglected in many previous studies.

Compared to cold-to-warm cases, KnPL1 shares similar characteristics, including the significant growth in boundary layer and high surface flux, as the overall wind was slightly tilted towards the warm side. However, in this case, the pressure gradient term may have greater impact on the boundary layer. The behavior of the MABL further away from the front was also different: MABL was quasi-static in KnPL1, while in KnCW2 and PiCW1 further deepening was still observed at distances greater than 120 km. These distinct differences suggested the important role of cross-frontal wind on advecting air with different properties and leading to a significant boundary layer change: weak winds greatly limit the ability of the MABL to receive the influence from the front due to limited advection, and air temperature can adjust to underlying SST to produce greater thermally induced pressure gradient.

**Transects from CLIMODE Pilot cruise**

Two transects from the Pilot cruise have primary wind blowing parallel to the SST front. The first transect, later referred as PiPL1, started on 1PM Jan. 20 and lasted 1.5 day. The second transect PiPL2 started on 2AM Jan. 24 and lasted 0.9 day. The major difference between these two transects and KnPL1 is that the background wind was more aligned
with the SST front, causing less cross-frontal air transport. While the depth-average wind speed was still high in both transects (9-16 m/s), the boundary layer further away from the front in these two cases is expected to receive less influence from the SST front due to weak cross-frontal advection. There was no major synoptic system observed during PiPL1, while there was a significant pressure system passing during PiPL2 (observed from ASIS buoy drifting in the vicinity of the SST front), causing the wind speed during the first half of the transect (ship in the warm side of front, right-hand side of Figure 2-34a) to be constantly high (15 m/s), and lower wind speed during the second half (ship closer to cold region, left-hand side of Figure 2-34a).

Figure 2-31: Same as Figure 2-14 but for PiPL1
Figure 2-32: Same as Figure 2-14 but for PiPL2
Figure 2-33: Same as Figure 2-16 but for PiPL.
Figure 2-34: Same as Figure 2-17 but for PiPL.
The overall boundary layer conditions in PiPL1 and 2 were similar. Boundary layer height was nearly constant during both transects, except for the observed formation of SBL at -40 km in PiPL1. This is consistent with the observed lower buoyancy flux (50-100 W/m²). Weaker cross-frontal advection caused air to reach equilibrium with underlying SST change more readily, leading to the observed smaller air-sea temperature difference compared to previous cases (Figure 2-31d). While the boundary layer can still be considered as weakly convective due to the positive buoyancy flux, the entrainment caused by such forcing would be much lower and can be on the same magnitude of regional subsidence, leading to the observed quasi-static boundary layer height. There was still an observed significant acceleration (5 m/s) in wind speed on the warm side of the front in PiPL1, while in PiPL2 the superposition of the synoptic system made it hard to estimate the spatial distribution of wind speed and surface wind stress. Consistent with previous cases, the change in stability parameter $z/L$ showed very weak or no correlation with the observed wind change in PiPL1, suggesting a weak contribution to wind modification from the stability induced surface stress change.

The thermally induced pressure gradients during PiPL1 and 2 are shown in Figure 2-35. In PiPL1, the magnitude of the pressure gradient is similar to KnPL1, higher on average (neglecting the negative outlier) compared to cold-to-warm cases. The sign of the pressure gradient was positive near the front, but later changed to negative further away from the front. If we assume such pressure gradient is nearly persistent, then the calculated value could contribute significantly to a wind acceleration of magnitude similar to the observed acceleration (5 m/s) in the region between 20 km and 50 km, as in this case the wind was blowing parallel to the front, and the sign of the pressure gradient would increase the observed surface wind speed and surface stress, acting like a sea-breeze, which agrees with result from Kawai 2014. In PiPL2, the thermally induced pressure gradient is weaker ($0.8 \times 10^{-3}$ m/s²), consistent with the observed low variation in wind speed in the 0-45 km region in Figure 2-34 (a). Due to the superposition of the synoptic pressure system induced wind acceleration, the pressure gradient induced wind modification was unclear away from the front in PiPL2. Overall, compared to the cold-to-warm case, thermally induced pressure gradient in the parallel case is about 50% higher, which is again consistent with our hypothesis that under weak cross-frontal advection, the
adjustment length scale for pressure $L_P$ would become relevant compared to the vertical mixing scale. There is a relatively weak correlation between observed wind speed and pressure gradient, which is still stronger than that for the cold-to-warm case. However, due to temporal and spatial trends over the transects and superposition of mechanisms on other scale such as atmospheric tide and synoptic pressure system, such correlation was not evident everywhere.

The cloud formation observed during PiPL1 was prevalent, due to the weak but still convective boundary layer. The average cloud depth was 150-300 m, less thick compared to strong convective cases, where decoupling due to deep cumulus formation is more common. In PiPL2, the sounding launched at 110 km did not observe strato-cumulus formation while the other two soundings observed the cloud at the top of the MABL around 500 m. Given the less rampant cloud formation, we assumed a cloud top radiative
cooling flux of around 10-15 W/m² compared to the 30-40 W/m² we assumed for KnCW and PiCW. The entrainment rate estimation based on Eq.2.15 and Eq.2.25 would be invalid in this case, as the boundary layer height was quasi-static, likely due to the balance between the small entrainment and subsidence. We still calculated the components of entrainment, and they are shown in Figure 2-36 and 2.37. In PiPL1 we can see that the total entrainment rate is 2-4 mm/s, and with estimated subsidence rate of 1-2mm/s, the combined boundary layer height adjustment is consistent with the observed quasi-static boundary layer height shown in Figure 2-33a. In PiPL2, the shear-driven entrainment at 110 km is significantly larger, comprising over 70% of total estimated entrainment, mainly due to the high surface wind speed induced by the passing pressure system. The reason we do not see corresponding rampant boundary layer height adjustment in Figure 2-33 (a) is likely that such entrainment is highly local due to weak cross-frontal advection. While with the high shear-driven entrainment the local MABL can rapidly grow, the next sounding observation was made 60 km away and 6 hours later, which can hardly resolve the local boundary layer adjustment. If soundings were launched at the same location, it would be more likely to observe the local MABL evolution. The surface buoyancy and cloud top flux in PiPL2 act more in concert with observed MABL adjustment, providing relatively small entrainment that can be balanced by subsidence. The correlation between entrainment and distance from the front is much weaker compared to the strong cross-frontal wind case, which is consistent with our initial expectation, as the MABL would be less likely to feel the presence of the SST front with smaller advection.
Overall, the parallel wind MABL showed different behaviors compared to the cold-to-warm case discussed in the previous section. Thermally induced pressure gradients were stronger in this case, as weak advection allowed more time for air to adjust to SST heterogeneity. The coupling between the pressure gradient and observed wind adjustment
was slightly stronger, yet the correlation was not always evident. The boundary layer height growth showed weak to no correlation with the distance from the front when the wind was more aligned with the SST front, implying the weak influence that MABL received from the front in this case due to weak advection. Buoyancy flux and cloud top cooling provided lower entrainment, as the air-sea temperature difference was smaller, and the overall height evolution observed was consistent with the estimation.
2.4.3 Warm to Cold Case

As discussed in the introduction section, when the background wind is blowing from the warm side of the SST front to the cold side, warmer air advected to a cooler sea surface will likely cause the formation of a stable boundary layer (SBL). Characteristics of such a layer are shallow height, stably stratified temperature profile, near zero or negative surface heat flux, and frequent formation of a super-geostrophic, fast-moving low-level jet at the top of the SBL (Stull 1988). The previously formed upper part of the boundary layer becomes the residual layer on the cold side, due to the decoupling induced by the surface SBL and its height is mainly modified by subsidence, as the surface flux can no longer support enough vertical turbulent mixing to affect the upper part of the MABL. Given these characteristics, a decrease in boundary layer height with distance on the cold side of the front is generally expected, along with a spatial and temporal deepening of the SBL height, described by Eq.2.20 and Eq.2.22.

**Case from the PEACH cruise**

While only 10 rawinsondes were launched during the PEACH cruise, station 2-9 (Nov. 23 2018, 1AM UTC through Nov. 23 11PM) provided a well-resolved observation of the boundary layer structure across the SST front when the background wind was strong (O(10m/s)) and blowing from warm to cold. The ship track is shown in Figure 2-38. It is clear that the advection by the background wind brought warmer air to the cold side and caused a reversed sign in air-sea temperature difference once the ship crossed the SST front, marked by the 0 distance in Figure 2-40 and 2-41. The formation of a stably stratified internal boundary layer in the first 150m above the sea surface was observed (Figure 2-39 and 2.40a). Note that the outlier at -21km was not abnormal but due to the deployment sequence of the soundings. Sounding 2 (the outlier) was first launched on the cold side, then the ship steamed into the warm region and came back to launch the rest of soundings. As a result, sounding 2 observed the internal boundary layer height at the beginning of the formation, while adjacent points on Figure 2-40 (a) observed the SBL height at much later times. In Figure 2-39, where SBL heights are plotted against time,
the temporal evolution of the SBL height is much clearer. It is also clear that with the flipped sign in air-sea temperature difference, the stability parameter \( z/L \) of the boundary layer quickly adjusted to a neutral near stable state from an unstable state on the warm side (Figure 2-41b). In concert, the surface buoyancy flux became negative once the ship entered the cold region, marking a sharp decrease (200W/m\(^2\) to negative, near zero). The change in surface buoyancy flux and stability clearly shut down the vertical turbulent mixing, decoupling the surface layer and the residual layer, marked by a decrease in residual layer height of O(200m) over the course of this transect (24 hours). Such a decrease (O(2mm/s)) was observed in the vertical temperature profile by identifying the inversion top of the residual layer and was assumed to be mainly caused by regional subsidence, which was of the same order of magnitude estimated from the regional reanalysis MERRA-2.

With Eq.2.22 and ship temperature measurements, an estimate of the boundary layer height is shown in Figure 2-40 (a). While there is a consistent under estimation over the first 25 km, which is assumed to be due to the fact that soundings (except 2) were launched much later in time and the SBL had adjusted further, overall, the estimate from Eq.2.22 is consistent with the observations, suggesting the importance of distance, background wind speed and temperature difference in SBL height modification.

The local air pressure, shown in Figure 2-41 (e), did not show a strong correlation with the SST gradient. Previous observations on a longer scale (15-90 days) and numerical model simulations suggested a possible correlation between SST gradient and thermally induced pressure gradient over longer time scale, but in this case, there was no evident local pressure difference on the cold side of the front. Air pressure did not differ significantly between the two sides of the front, except that the air pressure further dropped on the cooler side (-30 km). Based on the cruise report (Seim et al, 2018), cyclogenesis was reported in the evening of Nov. 23 in the region of interest, and it is assumed that the pressure drop was due to the passing of the synoptic low-pressure system, rather than local thermally induced pressure anomaly. The thermally induced pressure gradient cannot be calculated with Eq.2.13 as the surface layer was decoupled from the upper boundary layer and the hence analysis of the pressure mechanism can be
only qualitative. The result is similar to the cold-to-warm case, where the thermally induced pressure gradient is weak. This is consistent with the finding that on a shorter time scale (O(1 day)) and when cross-frontal wind is strong, the adjustment timescale of a local thermally induced pressure gradient could be too long to have a major impact on the boundary layer adjustment. In the PEACH case, the strong cross-frontal wind (O(10 m/s)) shortened the length scale for turbulent adjustment and the surface heat flux became the driving force in SBL formation.

The wind velocity, shown in Figure 2-39, is slowed by 2 m/s on average on the cold side compared to the warm side, if one neglects the sudden increase at ~40 km, which is likely due to the local pressure gradient induced by the cyclogenesis. The surface stress acted in coherence with the wind velocity. Such a decrease in wind velocity can be explained by two possible mechanisms: the stability induced surface stress change shown in Eq.2.9, and the lack of high-momentum air mixing down due to the decoupling caused by the formation of the SBL. Considering the usual vertical mixing timescale (O(30 min)), the decoupling could have enough time to modify the surface wind over a distance of 10 km (the bin width in Figure 2-40 and 2-41), and thus both mechanisms could play important roles in this case.

It is worth noting that a strong (O(25 m/s)) low-level jet was observed from the vertical wind profiles from the soundings. The height of the maximum wind speed, unlike the SBL height, was constantly around 140 m above the surface. We assume that this was a cyclone induced LLJ, rather an SBL inertial oscillation induced LLJ, as the wind speed reached a maximum of 50 m/s, which was too fast to be supported by a recently formed SBL.

Based on the vertical humidity profile, there was no significant cloud formation on the cold side during this transect, consistent with theoretical expectations; the formation of the SBL retarded the moisture exchange between the surface and the upper atmosphere, prohibiting the formation of stratocumulus that were often seen in a convective boundary layer. As the cloud cover was minimal, the radiative cooling at the top of the residual layer could be ignored and the boundary layer, in this case, can be treated as fully driven by the surface layer.
Figure 2-38: Ship track (blue line) during sounding launches on Nov.23, 2018. 24-hour averaged sea surface pressure from ERA5 reanalysis is shown as the background. SST front identified with GHRsst is shown as the red line.

Figure 2-39: Temporal SBL height evolution during sounding launches.
Figure 2.40: (a) Spatial SBL height evolution as a function of distance from the front (dots) and estimated SBL height based on Eq.2.22 (orange line). (b) SST distribution near the front (c) same as 2.40(b) but for air-sea temperature difference. (d) same as 2.40(b) but for air temperature. Negative distance indicates the cold side of the front.
Figure 2-41: (a)-(e) Bulk buoyancy flux, z/L, wind velocity, surface wind stress and air pressure plotted against distance from the front.
2.5 Summary

Meteorological observations, including sounding launches, in situ sea surface measurements and flux estimates from three field campaigns in the region of the Gulf Steams, are used to study the boundary layer adjustment near an SST front. Combined with satellite data and reanalysis, this study focuses on examining the relation between observed wind and boundary layer height modification and physical dynamics proposed in previous studies: stability induced surface stress change, vertical mixing mechanism due to internal boundary layer formation, and pressure gradient adjustment mechanism. To better understand the role of cross-frontal advection, this study divided transects from the campaigns into three categories: wind blowing from the cold side of the front to the warm side, the warm side to the cold side, and parallel wind along the SST front.

In the cold to warm case, the boundary layer height showed a significant increase over the warm region, where high surface buoyancy flux was observed due to the significant air-sea temperature difference. The stability induced surface stress change, however, only contributes to a small portion of the observed wind change. The correlation between the calculated thermally induced pressure gradient and the wind adjustment is also weak in this case. The estimated entrainment rate combining the contribution from surface flux, shear, and cloud top radiative cooling, is close to the observed entrainment, suggesting that the IBL physics is likely to be dominant in this case and the vertical mixing mechanism is acting to bring high momentum air from above to accelerate the surface wind.

In the parallel wind case, the boundary layer adjustment is confined within 50 km of the SST front due to the weak cross-frontal advection. Stability induced surface stress change again shows a weak correlation with the observed wind change. The magnitude of thermally induced pressure gradient is considerably larger than it is in the cold to warm case and shows a slightly stronger correlation with surface wind. This may be because the adjustment length scale for pressure gradient in weak cross-frontal wind case is comparable with the scale for turbulence (Spall 2007, Kilpatrick 2014). The entrainment
rate estimated from IBL dynamics is again close to the observed value, suggesting that in parallel wind case the vertical mixing is still the dominant mechanism.

In the warm to cold case, a stable boundary layer is formed over the cold side of the front due to the warm air advected above the cool sea surface. Due to the inadequate resolution of the vertical sounding profile, the height of the surface stable layer contains significant uncertainty, and the analysis is more qualitative in this case. Negative buoyancy flux was observed on the cold side together with decreasing air pressure. The estimated SBL height with Eq.2.22 is close to observed SBL height, suggesting the importance of the air-sea temperature difference in SBL formation.

The effect of synoptic system on the local observations in this study is not negligible. The superposition of a synoptic pressure system led to complicated air pressure and wind adjustment in several transects and was difficult to filter due to the rather long interval between sounding launches. Cold-air outbreaks, which are common in the wintertime Gulf Stream region, led to abnormally high surface fluxes exceeding 400 W/m², and the behavior of the boundary layer was unique during the event compared to normal cold-to-warm cases.

Overall, this study provided observational evidence for the dynamics proposed in previous studies (Small 2008, Kilpatrick 2014). The observed boundary layer evolution near an SST front can be explained by the two major mechanisms: pressure adjustment induced by SST, and vertical turbulent mixing. The stability induced surface stress change, as suggested in previous studies (e.g., Wai 1989), has less influence. It is worth noting that in previous studies (e.g., de Szoeke 2005) the entrainment induced by shear was often ignored as it is small compared to the surface buoyancy and cloud top cooling. In this study, we showed that the entrainment due to shear could contribute as much as 1/3 of the total observed entrainment when the surface wind was high.

Future studies will compare the observations with numerical model outputs to better quantify the effect of different dynamics without the superposition of synoptic systems and uncertainty in determining the cross-frontal wind component. The higher temporal resolution afforded by models can help to better track the effect of turbulent mixing, which is often on a scale of 15 minutes. During the field campaign, the ship could not
stay at one location to track the MABL growth consistently, while in a model being able to track the boundary layer change at a set location will help to further refine the relation between entrainment and underlying dynamics.
Chapter 3
WRF 2-D Simulations

Abstract

A series of Weather Research and Forecasting model (WRF) idealized 2-D simulations are carried out to study the dynamics governing the evolution of the boundary layer under the influence of a meso-scale sea surface temperature (SST) front. Model background wind is varied from 0 to 15 m/s to simulate the marine atmospheric boundary layer (MABL) under different cross-frontal wind regimes. Sensitivity tests are carried out with physical schemes switched on and off to differentiate the contribution from several proposed dynamics: surface latent heat, atmospheric moist process, and cloud formation. Vertical mixing and pressure adjustment mechanisms are identified and analyzed in each case to study their relative strength. Results show that moist processes, including both surface and atmosphere moisture exchange, alter the behavior of the MABL near the SST front significantly by providing extra vertical mixing strength via buoyancy fluxes. The thermally induced pressure gradient is also affected, as the boundary layer height modification is different when the model is initialized with moist data. Comparison between model runs with different wind strength also shows that the pressure adjustment mechanism is significantly stronger and becomes comparable to the vertical mixing mechanism when the cross-frontal wind is weak. The addition of cloud formation in the model not only provides extra entrainment but also generates variabilities in surface temperature distribution. An unexpected finding of the research is that a moister and colder boundary layer occurs with the cloud formation. The comparison between the 2-D model and its observational reference suggests that the inclusion of moist processes is necessary in order to yield model MABL structure close to observational data.

3.1 Introduction

It is widely accepted that over a warm ocean, the high humidity and sensible heat exchange at the air-sea interface will have a significant impact on the marine atmospheric boundary layer (MABL) (Stull 1988; Small et al., 2008). In the Gulf Stream region, wintertime cold air outbreaks can often lead to surface latent heat exceeding 300 W/m² (Sweet et al., 1981; Warner et al., 1990). This strong moisture exchange directly affects the strength of the surface buoyancy flux, which can further strongly modify the structure of the boundary layer (Venkatram et al., 1977; Stull 1988; Liu et al., 2007; Small 2008).
It is thus of importance to include moist processes in the model to better simulate the response of the atmosphere to the underlying warm, moist ocean surface.

In previous studies, however, such moist process are rarely considered with a significant focus. Idealized model runs that had similar 2-D cross frontal wind setups (e.g., Skyllingstad et al., 2007; Spall et al., 2007) rarely included buoyancy flux analysis, or just had completely dry settings (e.g., Kilpatrick et al., 2014, 2016). One reason for such a configuration can be the less important role of moisture in a pure sensible-heat-momentum analysis, which was a focus of said studies. These studies often had their models initialized with dry atmosphere and minimum surface moisture to reduce the complexity of the dynamics involved in models of situations where the role of the moisture flux was expected to be negligible, which could be true when the boundary layer was over a dry land, or the aim of the study was focused on aforementioned dry dynamics. These studies focused on the surface wind profile change due to an SST gradient, and the related vertical momentum mixing and pressure gradient adjustment, in which the dominant dynamics can be complete without the addition of buoyancy flux. In more realistic 3-D simulations (e.g., O’Neill, et al. 2010), moist processes could be included as part of the initialization, but the analysis was still heavily focused on the sensible heat flux change due to the SST and resultant vertical mixing, while the role of the buoyancy term remained understudied.

This approach has been proven to be effective based on previous results (e.g. Schneider and Qiu 2015), and we can also see in later sections that the overall boundary layer structure is reasonably realistic in our dry setup as well. It is the point of this study, however, to argue that when studying the MABL evolution over a moist sea surface near an SST front, it is important to include the moisture flux. The surface latent heat flux, specifically, acts as a key element of the dynamics, as it provides a non-negligible portion of vertical mixing, which has been shown in previous studies (Small 2007; Spall 2008; Kilpatrick 2014) to be the major contributor in MABL evolution near an SST front when a strong cross-front wind is present.

With such a goal, this study aims to investigate the role of both surface latent heat flux and atmospheric humidity in the evolution of the boundary layer. Compared to the
Kilpatrick (2014) setup, this study includes cases that initialize the model with a moist sounding from observation instead of a completely dry one, which sets up a more realistic PBL structure with innate internal boundary layer and humidity profile. Surface latent heat, microphysics and cumulus schemes are used to enable the formation of clouds from moisture. It is clear from primitive simulations that with added moist processes, the structure of the MABL and evolution time scale is significantly different compared to a dry setup. How these processes affect the MABL dynamics will be investigated with a series of analysis.

As mentioned previously, the absence of moist processes in simulations disabled the formation of cloud and led to the lack of cloud’s effect on the evolution of the MABL in these model simulations. Based on observational results, however, the presence of cloud at the top of the boundary layer in a convective case has been proven to have a significant impact on the vertical mixing and entrainment (e.g., de Szoeke et al., 2005). It is thus important to have cloud schemes in the simulation to have a reasonable estimate of the cloud-top radiative flux and its impact on the MABL. During the CLIMODE campaign, cloud top observations were not available, and cloud top cooling could only be roughly estimated based on previous studies. By having proper cloud configuration, simulation results provide a more reasonable estimate of the radiative cooling that can be used as a reference when compared to observations.

The primary effects of clouds on the MABL can be divided into two main categories: the radiative cooling due to the negative buoyancy flux, and the direct interaction with both solar and infrared radiation (Stull 1988). The first process creates upside-down “thermals” of cold air sinking from the cloud top and acts as a form of entrainment (Stull 1988). The entrainment velocity $w_{eCloud}$ when cloud-top radiative cooling dominates can be again approximated by:

$$w_{eCloud} \approx \frac{\Delta l^*}{\Delta_{EZ} \theta_v}, \quad (3.1)$$

where $\Delta l^*$ is the net long-wave radiative flux divergence near the cloud top and $\Delta_{EZ} \theta_v$ the virtual potential temperature difference across the entrainment zone (Stull 1988). Such processes were still understudied in previous studies on MABL, where surface heat
fluxes were thought to be dominant in driving vertical mixing in a convective layer. Older PBL schemes were also less focused on resolving this process, as its contribution to the total TKE budget was small. More recent studies (e.g., EPIC 2001 campaign) however have shown the importance of such mechanism in driving entrainment at the top of the boundary layer, which can contribute as much as 20% of the total entrainment observed (de Szoeke 2005). Given the rigorous formation of bands of stratocumulus cloud along the north wall of the Gulf Stream due to the large magnitude of latent heat flux over the warmer ocean surface, analysis with an adequate focus on such a process is necessary. The effect of the cloud top cooling can be quantitatively studied with the more recent version of MYNN (Mellor–Yamada–Nakanishi–Niino) PBL scheme, which allows the parametrization of the cloud induced negative buoyance at the top of MABL. Sensitivity tests are performed by comparing simulations with microphysics/cumulus schemes turned on and off, which directly control the formation of any clouds. The goal is to quantitatively relate the boundary layer growth to the magnitude of the cloud radiative flux, which is less seen in previous studies, due to the lack of cloud top measurements with high enough resolution, frequency, or coverage.

The second effect, shading caused by cloud cover, is, however, a much more complicated process to simulate. An accurate parameterization of such a process requires knowledge of liquid water content, cloud cover, solar zenith angle and many other factors (Welch and Wielicki, 1984; Schmetz and Beniston, 1986). Given the rather primitive cloud schemes used in the model, it is beyond the scope of this study to quantitatively investigate the role of cloud shading in MABL evolution. This process usually creates negative feedback over land, as fewer thermals can be generated due to reduced surface heating and the growth of the mixed layer will be slowed. However, over a warm sea surface, the reduction in the solar radiation often causes much less temperature variation than it does over land due to the relatively stable SST. As a result, shading caused by cloud cover often creates much less influential feedback on the boundary layer growth over the ocean. It is thus safe to neglect this process even with cloud schemes fully functioning in this particular study, where the SST is set to be constant.
As mentioned in the observational part of the study, four major possible mechanisms have been proposed from previous studies to drive the atmospheric response to an SST change. They are: a) the stability induced surface stress change due to the imposed air-sea temperature difference; b) the enhanced mixing due to stronger surface turbulence and the mixing down of air above (Hayes et al., 1989; Wallace et al., 1989); c) the thermally induced pressure gradient with secondary circulation (Lindzen and Nigam, 1987), and d) the relative motion between air and sea surface due to current (Kelly et al., 2001; Cornillon and Park, 2001). The contribution from pure stability induced surface stress change has been shown to be much less significant (see e.g. Wai and stage, 1989; Spall 2007). The ocean current effect on the MABL, on the other hand, has been shown to be significant in regions where strong currents are present (see e.g., Seo et al., 2016). While the current in the Gulf Stream region can be strong, in this particular study, the current effects are not considered due to the 2-D model setup. The vertical mixing and pressure gradient mechanisms, thus, are the two major dynamics this study focuses on.

In the observational part of the study, the vertical mixing mechanism and its impact on the MABL evolution were evaluated by calculating the strength of buoyancy induced turbulent mixing, shear induced mixing, and cloud top negative flux, and comparing the resulted entrainment rate to the observed MABL height growth. While results have shown a positive correlation between the vertical mixing strength and MABL growth, the relatively large spatial and temporal gap between sounding launches made the sample volume small and inconsistent. The non-uniform wind direction over a 24-hour observation window also complicated the MABL structure. The pressure adjustment analysis, which relied on the thermally induced pressure gradient calculated between soundings, is also affected by the similar issue. With the ideal model simulation, the analysis can be more quantitative with a much higher spatial resolution. Comparison of the MABL at different locations at a particular time is also made possible. While similar model simulation studies have been carried out before, the strength of this study is the combination of observational data and model simulation that provides high resolution informed by realistic parameters to set model configurations.
The main hypothesis of this study aims to solve is the contribution of said two mechanisms in various wind cases. As Spall (2007) and Kilpatrick (2014) argued, the adjustment time scales for these two mechanisms are different, depending on the cross-frontal wind strength and spatial scale. They proposed that:

\[ L_M = \frac{UD_M^2}{K_M}, \quad L_P = \frac{UD_P^2}{K_P}, \]  

where \( U \) the cross-frontal wind velocity, \( L_M, L_P \) the adjustment length scales for mixing and pressure, \( D_M, D_P \) the depth scales, and \( K_M, K_P \) the vertical mixing coefficients, respectively. They argued that the mixing coefficients are on the same order, and the major difference between the two length scales is due to the cross-front wind related depth scale. In a strong cross-frontal wind case, \( D_M \ll D_P \) and \( L_M \ll L_P \), an internal boundary layer where vertical mixing mechanism is dominant is likely to form, while in a weak wind case, \( L_M \sim L_P \), the SST induced pressure gradient can have a comparable effect on the boundary layer. With proper choices of initial wind and SST profile, the relative strength of these mechanisms are analyzed via a 2-D model configuration.

3.2 Model Setup

General 2-D setup

A two-dimensional domain in \( x-z \) direction is achieved in the Weather Research and Forecasting model (WRF) with the ideal case em_seabreeze2d_x. Compared to other ideal case setups in WRF, this case allows full-physics to better simulate the MABL environment, which is expected to be comparable with observations collected in previous CLIMODE field campaigns. Some of the settings are to recreate the Kilpatrick 2014 2-D model, which is the starting point for this study. Additional modifications are added to allow more realistic simulations and more sophisticated dynamics, and details on these modifications will be addressed in sections below.

The primary domain used in all three wind cases is the same, which comprises 202 grid points along the \( x \)-axis with grid spacing of 2 km. On the \( z \)-axis, 144 grid points are
unevenly spaced across the 20000 m domain, with the majority of them below 3000 m to better capture the evolution and dynamics inside the MABL, which is often lower than 2 km. The latitude of the domain is set to 36°N with the corresponding Coriolis parameter, which is similar to the region where the CLIMODE experiments were conducted.

A periodic boundary condition is used along the y-axis, while an open boundary condition is used at upwind and downwind boundaries to ensure the energy propagates out properly, as described in Klemp and Lilly (1978), Klemp and Wilhelmson (1978) and Kilpatrick (2014).

**SST setup**

In this study, the initial SST profile is an ideal product generated based on observations and is considered to be static during the simulation. Specific values of the SST, front width, and gradient can be adjusted based on the case being studied and the corresponding observations with which the simulation are to be compared.

In cold to warm and parallel wind cases, the SST profile is set such that the warm side occupies the majority of the domain to allow full development downwind. The SST front starts at 60 km from the upwind side of the domain, which is referred as the cold side hereafter. The front width and temperature gradient over the front are set to be close to observations from CLIMODE, which are 20 km and 6-10°C, respectively. The cold side SST is uniform, and increases linearly inside the front region, then becomes uniform again on the warm side. A sample SST profile for the cold to warm case is shown below.

![SST profile](image)

Figure 3-1: A sample of SST profile used in cold to warm case. 0 marks the start of the SST front. Negative distance represents the cold side.
Sounding Initialization

To achieve more realistic simulations, MABL measurements from Rawinsondes are used as initialization profiles for the model. Sounding profiles from the CLIMODE campaign are used due to the relatively stable wind condition and wind-front angle during the experiment.

For cold to warm and parallel wind cases, a single vertical profile averaged from all soundings in cold to warm condition is used. Input from this profile includes pressure, temperature, and mixing ratio. Once the temperature profile is determined, the underlying SST value is adjusted so that the sounding profile is always neutral/slightly stable on the cold side. A wind profile is generated based on the on the wind case we are interested in and is uniform across the entire vertical column. For cold to warm case, a U=15 m/s wind blowing along the x-axis (cross-front direction) is used, and a U = 3 m/s cross-frontal wind is used in the weak/parallel wind case.

Given that the 2-D WRF model can only take a single vertical column as the initial profile, a spin-up period is needed for the model to adjust the entire domain, which differs from more realistic simulations, where a domain-wide initialization can be achieved with a set of realistic data. As a result, it is necessary to assume that the first 4-6 hours of the simulation, based on initial tests, are not good representations of the actual dynamics. Kilpatrick (2014) using a similar one-column initialization, showed that the simulated MABL was more consistent after 24 hours, while in this study, the simulation shows a relatively stable state at 12 hours.

Surface Layer and PBL Schemes

An improved Mellor–Yamada (MY) turbulence closure model (MYNN model: Mellor–Yamada–Nakanishi–Niino model) is used for the surface layer and boundary layer scheme. The MYNN model is a second-order turbulence closure model that considers buoyancy and stability effects with constants determined from a LES data base (Nakanishi and Niino 2009). Compared to other model, MYNN’s strength in simulating PBL with strong convection and humidity makes it a preferred scheme to use over a warm ocean surface.
In MYNN, the 1-D equations for mean quantities are given by:

\[
\begin{align*}
\frac{\partial U}{\partial t} &= -\frac{\partial}{\partial z} \langle uw \rangle + f \left( V - V_g \right), \\
\frac{\partial V}{\partial t} &= -\frac{\partial}{\partial z} \langle vw \rangle - f \left( U - U_g \right), \\
\frac{\partial \Theta_l}{\partial t} &= -\frac{\partial}{\partial z} \langle w \Theta_l \rangle + \frac{f \theta_0}{g} \left( V \frac{\partial U_g}{\partial z} - U \frac{\partial V_g}{\partial z} \right), \\
\frac{\partial Q_w}{\partial t} &= -\frac{\partial}{\partial z} \langle w q_w \rangle,
\end{align*}
\]

where \( U, V, W \) are the velocities, \( U_g \) and \( V_g \) the geostrophic wind, \( q_w \) the total water content, \( \theta_l \) the liquid water potential temperature, \( f \) the Coriolis parameter, and \( g \) the gravitational acceleration. Capital letters denote ensemble-averaged variables, small letters turbulent variable, angle brackets for ensemble average and subscript 0 as a reference state. (Nakanishi and Niino 2009)

The unknown second-order turbulent fluxes MYNN solves are given by:

\[
\begin{align*}
\frac{\partial q^2}{\partial t} &= -\frac{\partial}{\partial z} \left( w \left( u^2 + v^2 + w^2 + \frac{2p}{\rho_0} \right) \right) \\
&\quad -2 \left( \langle uw \rangle \frac{\partial U}{\partial z} + \langle vw \rangle \frac{\partial V}{\partial z} \right) + 2 \frac{g}{\theta_0} \langle w \theta_V \rangle - 2 \varepsilon^2 \\
\frac{\partial (\Theta_l^2)}{\partial t} &= -\frac{\partial}{\partial z} \langle w \Theta_l^2 \rangle - 2 \langle w \Theta_l \rangle \frac{\partial \Theta_l}{\partial z} - 2 \varepsilon_{\Theta_l} \\
\frac{\partial (\Theta_l q_w)}{\partial t} &= -\frac{\partial}{\partial z} \langle w \Theta_l q_w \rangle - \langle w q_w \rangle \frac{\partial \Theta_l}{\partial z} - \langle w \Theta_l \rangle \frac{\partial Q_w}{\partial z} - 2 \varepsilon_{\Theta q} \\
\frac{\partial (q_w^2)}{\partial t} &= -\frac{\partial}{\partial z} \langle w q_w^2 \rangle - 2 \langle w q_w \rangle \frac{\partial Q_w}{\partial z} - 2 \varepsilon_{q_w}
\end{align*}
\]

where \( p \) is the pressure, \( \rho \) the air density, \( \theta_V \left( \equiv \theta \left( 1 + 0.61 q_v - q_l \right) \right) \) the virtual potential temperature, and \( \varepsilon, \varepsilon_{\Theta_l}, \varepsilon_{\Theta q}, \) and \( \varepsilon_{q_w} \) are the dissipation rates of \( \frac{\varepsilon q^2}{2}, \frac{\varepsilon \Theta_l^2}{2}, \frac{\varepsilon \Theta_l q_w}{2}, \) and \( \frac{\varepsilon q_w^2}{2} \), respectively. More detailed parameterizations and justifications have been thoroughly discussed in previous studies (e.g., Nakanishi and Niino 2001, 2004, 2006,
and 2009), and will not be discussed further in this study. The WRF specific MYNN configurations, however, will still be of importance to note as below.

MYNN computes the local component of the turbulent fluxes with an eddy-diffusivity approach. The fluxes are represented as a product of a local gradient of \( \phi \) and an eddy-diffusivity coefficient, given by:

\[
\overline{w'\phi'} = -K_{h,m} \left( \frac{\partial \phi}{\partial z} - \gamma \right),
\]

where \( \phi \) can be any scalar or momentum component and the counter-gradient term, \( \gamma \), a function of the higher order moments only used in the level-3 closure, \( K_h \) the eddy diffusivity coefficient for thermal and moisture variables, and \( K_m \) the eddy-viscosity coefficient for the horizontal velocity components (Olson, et al., 2019) MYNN generates \( K_{h,m} \) as a function of mixing length scale \( \ell \), and stability functions \( S_h \) and \( S_m \), shown as:

\[
K_{h,m} = lqS_{h,m}.
\]

The TKE equation in MYNN is given by:

\[
\frac{\partial q^2}{\partial t} = \frac{\partial}{\partial z} \left[ lqS_q \frac{\partial q}{\partial z} \right] + P_s + P_b + D,
\]

where \( P_s, P_b, \) and \( D \) refer to shear production, buoyancy production, and dissipation, respectively. The first term on the right-hand side refers to the vertical transport term. It is worth noting that the buoyancy production term, compared to original MYNN, has been modified to take cloud-top radiative cooling into account. As previous studies (e.g. Deardorff 1980; Duynkerke and Driedonks 1987; de Szoeken 2004) have shown, the negative buoyancy generated by stratocumulus clouds can drive convective turbulence when the surface buoyancy fluxes are relatively small. The cloud top convective velocity scale is then defined as

\[
w_l = \left[ \frac{g}{\theta} \left( w'\theta' \right) \right]^{1/3},
\]
which is similar to $w_*$ from surface heat flux but uses cloud radiative fluxes instead.

It is worth noting that, as MYNN-EDMF (eddy-diffusivity / mass-flux) defines TKE on mass points, it is possible to advect TKE, whereas other models usually cannot achieve such function (Olson et al. 2019). In this study, the numerical instabilities near the boundary are still significant, especially in the transport term, so the TKE advection option is still to be used with caution.

One noteworthy configuration used in this study is the choice of the mixing length. The original MYNN defines the mixing length as a harmonic average in the form of:

$$\frac{1}{l} = \frac{1}{l_s} + \frac{1}{l_t} + \frac{1}{l_b},$$

(3.9)

where, $l_s, l_t, l_b$ are surface-layer length, turbulent length and buoyancy length, respectively. A revised version that includes cloud mixing length can be found in Ito et al. (2015). The revised version of the mixing length is well-tested and shown to have better performance in stable boundary layers (Olson et al. 2019) and is used in this study.

While in situ surface observations collected in previous field campaigns are based on the COARE parametrization, the WRF simulation cannot use the COARE-based MM5 similarity surface scheme, as the MYNN PBL scheme requires the coupled MYNN surface scheme to function. Potential differences between schemes should be treated with caution, but are not a focus of this study.

**Radiation Schemes**

To make the simulation closer to observations, a set of primitive radiation schemes is used to provide necessary physics, such as absorption and scattering. The shortwave scheme used is the Dudhia scheme (Dudhia 1989). This scheme allows simple and efficient parameterization of the short-wave radiation with proper scattering, attenuation and absorption with cloud taken into account. For long wave radiation, the Rapid Radiative Transfer Model (RRTM) is implemented. RRTM uses a lookup table for efficient long-wave radiation parametrization, which utilizes correlated-k approach to calculate fluxes and heating rates (Mlawer et al., 1997).
In previous studies (e.g., Kilpatrick 2014, 2016), the radiation scheme was often turned off in 2-D simulations, especially when latent heating was turned off so cloud formation was prohibited. In this study, based on simulation results, the MABL is not significantly affected by the diurnal cycle or cloud-covering-induced radiation change within the 36-hour simulation window; it is, however, still necessary to add a functioning radiation scheme, as other schemes such as cumulus and microphysics depend on the radiation to run properly.

**Microphysics Scheme**

To enable basic cloud dynamics, the Kessler microphysics scheme (Kessler, 1969) is used. The Kessler microphysics scheme aims to parametrise the warm-rain process with simplicity and efficiency. This scheme does not include ice or snow, which is acceptable in this particular simulation, where such processes are not considered being significant in MABL evolution over a warm ocean surface. Being a one-moment model that ignores several important microphysics processes, the Kessler scheme significantly simplifies cloud dynamics, which can lead to unrealistic precipitation profiles compared to other models. In this study, however, as precipitation is not as impactful, the Kessler scheme should not produce a significant bias in simulation results, while its high reliability and efficiency surpasses other available schemes in our model test runs. More details regarding this model can be found in Kessler (1969).

**Cumulus Scheme**

Cumulus formation is an important process in MABL evolution, especially in a convective boundary layer where thermals created by surface buoyancy flux bring moist air up and create clouds at the top of the mixed layer (Stull 1988; Small 2008). Such process can not be ignored when the cloud top radiative flux is taken into account, which can generate significant impact in the form of negative buoyancy flux at the top of MABL (e.g., de Szoeke 2005). Thus, it is crucial to properly parametrise the cumulus formation process in this study. The Kain-Fritsch (KF, Kain and Fritsch, 1993)
convective parameterization scheme (CPS) is used to allow deep and shallow convection. This scheme is based on the fundamental closure assumption as the Fritsch-Chappell (FC) (1980) scheme, which assumes the convective effects to remove convective available potential energy in a grid element within an advective time period (Kain and Fritsch, 1993). It modulates the two-way exchange of mass between cloud and environment as a function of the buoyancy characteristics of various mixtures of clear and cloudy air, while also assuring the conservation of mass, thermal energy, total moisture and momentum (Kain and Fritsch, 1993; Kain 2004).

With microphysics and cumulus schemes, the model has been able to generate reasonable clouds at the level where the top of the MABL is assumed to be, as shown in Figure 3-2. The distribution of clouds is qualitatively consistent with previous observations (e.g., Carson, 1950; Sublette and Young 1996; Small 2008;), where cumulus clouds were formed over the warm region.

Figure 3-2: The cloud fraction profile in a cold to warm simulation, \( U \) (from left to right) =15m/s, cloud fraction is colored code with range 0 to 1 as blue to yellow.
3.3 Methods

**Boundary Layer Height**

An accurate planetary boundary layer height (PBLH, or often referred as $z_i$) estimation is of crucial importance in MABL studies, as it provides a fundamental description of the structure of the boundary layer and is used as a base parameter in various flux and turbulence parameterizations. While there are numerous ways to determine the PBLH, including thermal profile, humidity profile, capping inversion height, and critical Richardson number, the commonly used and proved-to-be-accurate method in a convective boundary layer is to find the top end of the well-mixed virtual temperature profile.

In the original WRF configuration of MYNN, PBLH is determined by the 1.5-theta-increase method, where the top of the mixed layer is set to be the first level where the virtual temperature is higher than the minimum virtual potential temperature inside the boundary layer by 1.5 K. While this method is widely used and tested and has shown relatively unbiased results compared to observational data (e.g., Nielsen-Gammon et al. 2008), the 1.5 K critical value is constantly overestimating the PBLH in this 2-D study. Based on test results, lowering the critical value to 0.2 K yields diagnostic PBLH that captures the top of the mixed layer most accurately. While this value is not widely tested, it provides a standard that is similar to the criteria used in the observational data that is to be compared with the simulation results. In CLIMODE sounding profiles, the top of the boundary layer is manually picked by visually identifying the capping point in the virtual potential temperature profile, and the 0.2K method would best agree with the manually determined values, which keeps the consistency between the model diagnostic PBLH and real observations.

Note that in a cloudy boundary layer, due to the much more complicated vertical virtual potential temperature profile, the default diagnostic PBLH method based on temperature is no longer valid, as it assumes a vertically uniform temperature profile. To estimate a more accurate PBLH with cloud in presence, the height where there is a cap in specific
humidity is used instead. Such a method has been tested in no-cloud cases and has shown a good agreement with the default method.

![Virtual Potential Temperature Profile](image)

Figure 3-3: A virtual potential temperature profile with diagnostic Zi from the model with 0.2K method. The gray line shows the level where the top of the mixed layer is. Location is 140km into the warm side, with cold to warm wind U=15m/s. Zi is 987m.

### Entrainment Calculation

The relationship between entrainment rate and boundary layer height growth in a free convective boundary layer can be approximated by:

\[
\frac{dz_l}{dt} = w_e + w_L, \tag{3.10}
\]

where \(w_e\) is the entrainment rate and \(w_L\) the subsidence rate. The subsidence rate in the model is set to be on order of 1 mm/s, similar to MERRA2 year-round estimate in the
The entrainment rate calculation is broken down into three components, similar to the formulation in the observational part of this study.

The first component of the entrainment rate is the direct effect of surface buoyancy flux, which is dominated by the convective velocity scale, defined as:

$$w_* = \left[ \frac{g z_l}{\overline{\theta_v'}} w' \overline{\theta_v'} s \right]^{1/3}, \quad (3.11)$$

The “s” subscript denotes the buoyancy flux at the surface. The entrainment rate then can be expressed as:

$$w_e = \frac{Aw_*^3}{\Delta b}, \quad (3.12)$$

where $\Delta b = g \Delta \theta_v / \theta_v$ is the buoyancy jump capping the inversion and $A$ the fraction of surface flux contributing to entrainment. In de Szoeke (2005) they used $A=0.2$ and found the surface buoyancy flux contributed only a small fraction to the observed entrainment. Other studies (Pino et al 2003, Conzemius et al 2006) pointed that $A$ around 0.32 is a better representation of the surface buoyancy flux force. In this study we used $A=0.25-0.35$ to estimate the surface buoyancy driven entrainment. The buoyancy jump is determined by taking the temperature of the model level 100m above the diagnostic PBLH and subtract it by the mixed layer temperature.

The second component comes from the shear-induced entrainment. Tennekes (1973) proposed that, in a pure shear-driven boundary layer, the entrainment rate can be estimated as:

$$\frac{dh}{dt} = \frac{2.5 \theta_0 u_*^3}{g \Delta \theta_v h}, \quad (3.13)$$

where $\theta_0$ is the initial average temperature across the boundary layer, $h$ the boundary layer height. The frictional velocity $u_*$ is a direct output from the model, while the buoyancy jump is calculated following the same steps in the previous section.

The third component, the entrainment due to cloud top radiative cooling, can be approximated by Eq.3.1. In this study, however, a more efficient way to calculate the
entrainment due to cloud is achieved by the built-in function MYNN PBL scheme that computes the cloud convective velocity scale, shown previously as Eq.3.8. This velocity scale is similar to \( w_* \), and thus can be directly used to calculate the entrainment rate due to the negative buoyancy at the top of the mixed layer.

**Thermally induced pressure gradient**

The pressure perturbation induced thermally by a surface heating due to SST can be calculated in the form:

\[
P_t = \frac{g}{\Theta} h \frac{\partial[\Theta]}{\partial x},
\]

(3.14)

where \([\Theta]\) is the vertically integrated potential temperature from the surface up to the boundary layer height \( h \), and \( \Theta \) is a scale value of the potential temperature (Mahrt et al. 2004). Note as mentioned in Chapter 1, we use potential temperature here to better compare model results with previous studies that are without moisture (e.g., Kilpatrick). In their study, they pointed out the inadequacy in the vertical structure observation and thus could only qualitatively compare the local pressure gradient with observed flow acceleration. In this study, the thermally induced pressure gradient is calculated in the same way, but the analysis will be more quantitative with much higher temporal and spatial resolutions.
3.4 Results and Discussion

3.4.1 Base Model Performance

Similar to model outputs from Kilpatrick 2014, the 2-D model stabilizes 4-6 hours after the initialization and shows no significant change in MABL structure after 12 hours, likely due to reaching quasi-equilibrium. It’s noteworthy that though the MABL still deepens slowly beyond 12 hours, the increase is slow and non-significant, thus is treated similarly as the trivial deepening after 24 hours shown in Kilpatrick’s work. The 2-D structure of the boundary layer is saved on a 30-min interval, and Figure 3-4 blue line shows the 12th hour snapshot in a cold to warm case run, with no cloud formation, microphysics, or latent heat flux, which is a basic setup close to Kilpatrick’s 2-D cold to warm simulation, where major moisture exchange from the sea surface was not allowed.

In this setup, the model is initialized with a uniform, constant, cold to warm geostrophic wind at a speed of 15m/s. The initial boundary layer is set with an averaged cold side sounding profile from a cold to warm transect during the CLIMODE Knorr cruise, hereafter referred as KnCW1. The width and gradient of the SST front in the model is set to be close to observations made during the transect. Note that, in the observation, the complex shape of the front led to a continuously changing wind-front angle, while in the model the wind is always crossing the front perpendicularly. This leads to a difference in the magnitude of the cross-front component of the wind between the observation and the model, which will be discussed in later sections.

Shown in Figure 3-4 as blue, it is evident that the MABL responds to the 8-degree SST front and adjusts its structure rapidly within the first 50km into the front. Sensible heat flux rises quickly from stable conditions to a 100W/m^2 maximum at around 25 km into the warm side of the front, then slowly decreases further into the warm region. Surface latent heat flux is all zero, as prescribed in the model setup. The wind field shows a structure similar to Kilpatrick 2014: a dome-like cell within the boundary layer with intensified surface wind, shown in Figure 3-4 (a) and (b), where a significant increase in the 10-m wind speed U10 can be seen on the warm side of the front. This is consistent
with previous studies (e.g., Chelton 2004, Spall 2008, O’Neil 2010), where wind acceleration on the warm side is expected due to several possible mechanisms, including vertical mixing and pressure gradient forcing. The layered structure of both cross-front and along-front wind components $U$ and $V$ on the cold side get rapidly mixed with enhanced vertical mixing on the warm side. The potential temperature profile shows a well-mixed structure inside the boundary layer, with its top matching the diagnostic height of the mixed-layer. The specific humidity profile is in a uniform dry state, due to the lack of moisture mixing from the sea surface. Note that in later simulations, the model is initialized with a sounding with moisture, but in this particular case a completely dry initialization is used to better emphasize the lack of moist processes.

With the model behaving similarly to previous studies (e.g. O’Neil 2010, Kilpatrick 2014), it is reasonable to assume that this base setup can represent a reliable primitive simulation of the boundary layer near the SST front.

Figure 3-4 (on next page): 2-D Model outputs from three cold to warm cases ($U=15m/s$), from top to bottom, (a) to (f): SST (all three cases have the same value), surface sensible heat flux, surface latent heat flux, surface buoyancy flux, 10m wind velocity $U_{10}$, and model diagnostic boundary layer height $PBLH$.

Blue, red and yellow lines represent the dry sounding with no moisture setup, the dry sounding with moisture setup and the wet sounding with moisture setup, respectively. The wet sounding has a moist initial humidity profile with surface air at 4.7g/kg, compared to the vertically uniform 0.0014g/kg in the dry sounding used in Kilpatrick 2014. This snapshot is taken at 11.5hours after the model start.
Figure 3-5: Bottom 2000m of model outputs from the base no moisture dry sounding case at 11.5 hours. From top to bottom, (a) to (d): U, V, potential temperature, specific humidity. Corresponds to blue in Figure 3-4.
Figure 3-6: Same as Figure 3-5 but for the dry sounding moist setup, corresponding to red in Figure 3-4.
Figure 3-7: Same as Figure 3-6, but for the wet sounding moist setup, corresponding to yellow in Figure 3-4.
3.4.2 No-Cloud Cases

**Cold to warm**

To examine the influence of the surface moisture flux on the MABL, two similar setups with surface latent heat flux enabled are used for comparison, shown in red and yellow in Figure 3-4. Note that in addition to the difference between no surface flux and full moisture flux (blue vs. red, blue vs. yellow), differences in multiple profiles between dry sounding initialization and wet sounding initialization (red vs. yellow) are also non-negligible. As mentioned in the Method section, the initialization sounding used in most cases (including yellow) in this study is based on observations made during the CLIMODE campaigns and includes observed vertically varying humidity profiles, while the dry initialization used in particular base setups (blue and red) is close to a completely dry state (with vertically uniform specific humidity 0.0014 g/kg), to resemble Kilpatrick (2014)’s dry setup. The role of the moisture in the evolution of the boundary layer, including both surface flux and atmosphere humidity, will be compared and discussed in detail below. To compare the model results with observations, the data from KnCW1 is used as a reference. Note that due to the limited dynamics and initialization options in 2-D WRF cases, as well as the lack of synoptic systems, the model results are not matching its observational counterparts perfectly. The difference is more prominent, especially in actual boundary layer height. We conclude that, however, the model results should still agree with observations relatively well in entrainment rate, pressure gradient and relative PBLH growth trend, as these dynamics are less dependent on larger scale processes and are better simulated in the 2-D WRF setup. The justification for such a conclusion will be addressed in the later 3-D model chapter.

The surface sensible heat flux (Figure 3-4 (b)) in completely dry case (blue) and completely wet case (yellow) behaves similarly with almost identical magnitudes and trends. The surface heat flux reaches the peak at 25km into the warm side, where cooler air from upwind-side brought by strong cold to warm advection creates the largest air-sea temperature difference. Further away from the SST front, the surface flux gradually decreases, as air-sea temperature difference decreases when the boundary layer receives
less influence from the SST front and reaches equilibrium. Note that the dry sounding with moisture flux case (red) has a similar overall trend but a much higher magnitude in both sensible and latent fluxes. As there is no cloud or other dynamics in these three cases, the difference can only be due to dynamics inside the boundary layer. Panel (d) in Figure 3-5, 6, and 7 show the difference in the vertical structure of the potential temperature. It is clear that the red case has a significantly cooler boundary layer (1-2 K lower) compared to the other two cases. The reason for such a difference is the much more vigorous vertical mixing on the cold side of the front. As shown in Figure 3-4 (f), the boundary layer height on the cold side is much higher. The enhanced deepening in red is mainly due to the high air-sea humidity difference induced by the dry sounding initialization with surface moisture flux enabled. In Figure 3-4 (d), we can see that red is the only case where a positive buoyancy flux is present in the cold region. While after 12 hrs of simulation, the humidity difference has been reduced, such a difference fuels the vertical mixing over the cold region during the spin-up phase by generating a significant surface buoyancy flux, creating an unstable boundary layer over the cool sea surface. The resulting temperature profile reflects the deeper upward mixing of the cool air that has a temperature close to cold side SST (285K). This column of cooler air is then horizontally advected by the strong cross-frontal wind, causing the stronger air-sea temperature difference inside the boundary layer over the warm region, ultimately leading to the shown higher sensible and latent heat flux, as the influx of cooler air is felt by the surface via enhanced vertical mixing over the warmer sea surface. Such mixing on the cold side is not seen in the other two cases, as in the blue case (dry sounding, no surface latent heat), the lack of moisture exchange prohibits vertical mixing, as the cold side is naturally thermally stable. In the yellow case (wet sounding with surface latent heat), the humidity difference is neutralized by using the sounding profile averaged from observations, which has a humidity profile already in quasi-equilibrium with underlying cool sea surface, and thus limited vertical mixing can be generated on the stable side. This stable side mixing behavior implies the importance of moisture in boundary layer studies, as previous studies (e.g. Stull 1988; Kilpatrick 2014; Schneider and Qiu 2015) that focus heavily on sensible heat budget and dry dynamics often use a simplified dry atmosphere and no surface moisture, as the sensible heat budget is their main focus, and the lack of moisture
is considered to have negligible effect. In this study, however, it is shown that when simulating a marine atmospheric boundary layer, the addition of moisture, while the sensible heat budget is not directly affected, can lead to a significant change in overall boundary layer structure that can indirectly affect the dry dynamics (e.g. the change in sensible heat flux in the red case).

Wind profiles in blue and yellow (Figure 3-5 and 7) show a strong zonal (cross-frontal) acceleration over the first 25 km into the warm side of the front, and a deceleration in meridional (along-frontal) component, indicating an anti-cyclonically turning wind, consistent with O’Neill (2010) and Kilpatrick (2014). On the cold side, meridional wind develops due to rotation and friction (Kilpatrick, 2014). In the red case, the cold side wind adjustment reaches as far as 100 km into the warm side, due to the higher boundary layer on the cold side and overall strong background advection.

The model diagnostic PBLH reflects the increased entrainment due to the addition of moisture. An increase greater than 18% can be seen in the red case, while the completely wet case shows a further deepening over the warm region. These increases in boundary layer height are considered as main consequences of vertical mixing, as the extra latent heat flux from the sea surface produces a significant amount of surface buoyancy flux, which directly contributes to the entrainment rate at the top of the boundary layer, described by Eq.3.11 demonstrated again below:

$$w_* = \left[ \frac{g z_i}{\partial \theta_v} \frac{w' \theta'_s}{\partial s} \right]^{1/3}. \quad (3.11)$$

The strength of this mechanism can be seen in Figure 3-8, where the entrainment rates for all three cases are shown. On the cold side, all three cases have close to zero entrainment, despite that the red case has a slightly higher rate due to the aforementioned air-sea humidity difference. The positive entrainment rate over the cold region in this case agrees with the much higher MABL and the well-mixed structure. Note that as the capping inversion at 12 hrs is strong in this case (over 5 K), the entrainment is minimal (1mm/s). During the spin-up phase, however, as the capping inversion is much weaker, entrainment over 12 hrs on the cold side can support the much higher PBLH shown in this case. On the warm side, it is evident that the addition of surface moisture greatly
increases the $\overline{w'\theta'_{v_s}}$ term and thus provides a significantly higher entrainment (yellow vs. blue). While in Figure 3-4 (d), it is shown that the red case has the highest buoyancy flux $\overline{w'\theta'_{v_s}}$, which in turn should yield the strongest vertical mixing, the entrainment rate in this case has a different trend compared to the other two. This again is due to the cooler boundary layer in red, which leads to a stronger capping inversion $\overline{\theta_y}$. The top of boundary layer temperature difference in red, shown in Figure 3-6 (e), is above 5 degrees on the cold side, and gradually decreases to lower than 2 degrees deeper into the warm side, while in blue and yellow case, the magnitude of the capping inversion is much smaller (1 – 2 K), due to the overall warmer boundary layer. This difference leads to the different trends in the entrainment rate in Figure 3-8, where the entrainment rate in red actually increases in the warmer region due to the weaker capping inversion, despite that the surface buoyancy flux gets weaker as the surface temperature difference reaches equilibrium. In blue and yellow cases the boundary layer reaches a state with similar temperature profiles after spin-up, and the resulting entrainment rates reflect the difference in surface buoyancy flux. This difference is also coherent with model diagnostic PBLH, where yellow shows the largest growth, despite that its surface buoyancy is lower than the red case. Compared to the observed entrainment from KnCW1, which is 9-14 mm/s (gray dash lines in Figure 3-8), the yellow line is closest to observations, justifying the inclusion of both surface moisture flux and atmospheric moisture. While the red case is also in range, the entrainment rate from observations decreases gradually deeper into the warm side, which conflicts with the red trend. Note that the cloud-top-induced entrainment is excluded from these three cases, while the observation value from KnCW1 includes both cloud contribution and shear contribution. The contribution from cloud on entrainment will be discussed in later sections, while the shear-induced entrainment is on the same order in all three cases due to their similar friction velocity $u_*$ and not to be discussed here.

Thermally induced pressure gradients in these cases are shown in Figure 3-9. It is important to see that while the thermally induced pressured gradient described by Eq.3.16 is often considered a dry process (e.g. Mahrt 1982, 2004; Kilpatrick 2014), the addition of various moist processes alters the behaviors of such a mechanism. The difference
between blue and red is mainly due to the larger overall $\partial [\theta] / \partial x$ in red caused by cooler air advection from up-wind. Note that in the 0-50 km region the $\partial [\theta] / \partial x$ is actually close to zero in red due to the highly horizontally mixed boundary layer, hence the smaller $P_t$, despite that $h$ there in red is much higher. In the warmer region, the larger temperature gradient contributes to over 70% of the seen difference in $P_t$ between red and blue, and red and yellow, while the contribution from difference in $h$ becomes even smaller as PBLH grows rampantly in yellow. Note that the PBLH $h$ plays a more important role in complete dry vs complete wet: the main difference between blue and yellow can only be caused by $h$ rather than $\partial [\theta] / \partial x$, as they share a temperature profile with a similar gradient. The difference in $h$ kicks in at 25-50 km, where the PBLH in yellow starts to outgrow blue significantly, and a larger difference in $P_t$ is seen. Deeper into the warm region, the magnitude of $P_t$ decreases as $\partial [\theta] / \partial x$ decreases in a quasi-equilibrium state boundary layer. $h$ is mainly modified by vertical mixing, which has been shown to be highly sensitive to the addition of moisture. While the relative difference in magnitude between wet and dry cases is significant (over 30%), the absolute difference ($0.2 \times 10^{-3} \text{ m/s}^2$) is small compared to common values near an SST front (e.g., Mahrt 2004; Kawai et al. 2014) and observed values from KnCW1 ($0.5 - 1 \times 10^{-3} \text{ m/s}^2$). This still, however, demonstrates the importance of moist processes in boundary layer studies, as its contribution to boundary layer evolution can also indirectly affect dynamics that are considered to be dry.
Figure 3-8: Surface buoyancy flux-induced entrainment rate for cold to warm cases (U=15m/s), snapshot taken at 11.5 h. Blue is the dry sounding no latent flux setup, red the dry sounding with latent flux setup, and yellow the fully wet setup.

Figure 3-9: Thermally induced pressure gradient $P_t$ for cold to warm cases (U=15m/s), snapshot taken at 11.5 h.
Parallel/Weak Wind

The three setups in the last section were also run in a weak wind setup. Due to the 2-D configuration of the model domain, the meridional (along-frontal) wind component has limited variability compared to common 3-D simulations, these cases still represent a valid weak wind scenario where the cross-frontal wind component is 3 m/s. To compare the model results with observations, another transect from CLIMODE Knorr is used as a reference, and hereafter referred as KnPL1. KnPL1 has a similar 8-degrees SST front, with 2-3m/s cross-frontal wind. Note that KnPL1 has a relatively strong (8-9m/s) along-front wind component, while the simulation has only a cross-frontal wind. This difference does not change how the model agrees with the observation in general, but does lead to a slightly higher entrainment in KnPL1 due to stronger shear production, which will be discussed later.

Like in strong wind cases, the dry sounding with surface-latent-heat case (red) again shows a unique stable side high (250 m) boundary layer, in concert with the positive buoyancy flux shown in Figure 3-10 (d). This deepening on the cold side creates a vertically mixed column of air, in contrast with the typical shallow stable boundary layer shown in the other two cases. As discussed in the previous section, this layer, with background wind advection, causes a cooler boundary layer to pass the SST front and reaching the warm side, ultimately leading to a higher inversion cap and stronger $\partial[\theta]/\partial x$, which affects both vertical mixing and pressure gradient force. Compared to its strong wind counterpart, in this case the cooler air cell extends only 50 km into the warm side, due to the weaker background advection. Its height is also lower (250 m vs over 500 m), indicating the weaker entrainment due to reduced surface wind stress. The differences in completely dry (blue) and completely wet (yellow) is similar to those in strong wind cases, mainly induced by the difference in surface latent heat flux.

Wind profiles in blue and yellow are similar to strong wind cases, with the difference being that beyond 150 km, the horizontal wind speed gradient no longer exists, as weaker advection (3 m/s) can only bring air to this point at 11.5 h. The increase in surface wind can be clearly seen in Figure 3-10 (e), where a 2m/s increase is present in both blue and
yellow case. It is noteworthy noting that such an increase is not evident in the red case. One possible explanation will be discussed in the later pressure gradient section.

The 150 km cut-off is also seen in PBLH (Figure 3-10 (f)), where the boundary layer height does not grow with distance beyond this point, as it no longer receives the influence from the front, and all the MABL deepening is caused by local buoyancy flux. In KnPL1, a similar cut-off is also seen at 100km. Note that again, due to the stronger inversion cap induced by the cooler boundary layer, the entrainment rate in red is much lower compared to yellow (Figure 3-14). The difference between dry (blue) and moist (yellow) setups leads to an over 40% increase in PBLH due to the stronger buoyancy flux during the spin-up phase. Compared to the strong wind case, these weak wind cases have a significantly lower entrainment rate from the smaller surface buoyancy flux due to the reduced wind velocity at the surface. The stronger advection in strong wind cases allows boundary layer deepening beyond 150 km (Figure 3-4 (f)), and while the entrainment rate becomes smaller with distance, it still modifies the top of the boundary layer far away from the front. In the weak wind case, the limited cross-frontal wind component causes the boundary layer modification to be within the first 150 km, and the entrainment rate becomes flat beyond that point, as the influence of the SST front is minimal. With the subsidence rate of 1-2 mm/s in the model, it is reasonable to believe that beyond the cut-off point, the boundary layer height is quasi-stationary. Compared to observed rates from KnPL1 (9-11 m/s), the entrainment rates in all these cases are lower. As mentioned previously, however, the along-front wind in KnPL1 creates a higher friction velocity that enhances the shear-induced entrainment, which is not in the model setup. A rough estimate of pure surface-buoyancy induced entrainment from KnPL1 is made to better compare with model results, shown as gray dash lines in Figure 3-14. Again, the completely wet case (4-6 mm/s) best agrees with the observations (5-7 mm/s).

In terms of thermally induced pressure gradient, weak wind cases in general have significantly higher magnitude, and their influence is confined within the first 150 km into the warm region. The substantially different behavior in pressure gradient is consistent with Eq.3.2, which argues that the weaker wind allows longer adjustment time for thermally induced pressure gradient to have a greater impact on boundary layer
evolution. The effect of cross-front wind strength can be seen in Figure 3-16, where results from 6 completely wet cases with different background wind speeds are shown. It is clear that as the cross-front wind component gets slower and advection gets weaker, the cooler air mass being transported gets more time to react to the underlying warmer sea surface, and thus creating a more stacked, less spread-out peak of $P_t$ on the warm side. In the $U = 0$ m/s case, thermally induced pressure is almost completely local at the SST front, while in the strong wind cases, $P_t$ is evened out as cooler air mass gets carried further into the warm region and its temperature gets closer to underlying SST. As shown in Figure 3-15, the magnitude of $P_t$ is much stronger within the first 150 km. The magnitude is consistent with KnPL1 (1-2·10^{-3} m/s²) and common values from previous studies in weak wind regions (e.g. Kawai 2014). The observed wind acceleration (Figure 3-10 (e)) in yellow and blue is in concert with the high $P_t$ there, indicating that $P_t$ can be at least partially responsible for an acceleration of 2 m/s, given that the vertical mixing in this region is less rampart compared to stronger wind cases. Note that despite the greater $\partial[\theta]/\partial x$ in red and the resulting high $P_t$, the surface wind speed in red does not shown significant increase crossing the SST front. The reason for this is less clear, but we assume that it can be due to the much better mixed layer in red, which consists of a more vertically uniform wind profile that is less affected by the surface SST. This can also explain the weaker increase shown in the strong wind case (Figure 3-4 (e)), yet this assumption should be taken with caution.

In general, weaker wind cases show similar differences between dry and wet setup, compared to strong wind cases. The addition of both surface and atmospheric moist process makes the simulation closer to observations. Compared to strong wind cases, the thermally induced pressure gradient is much stronger. Its horizontal distribution is also highly tied to advection strength, which supports the previous argument that $P_t$ can be more responsible for boundary layer adjustment in a weak wind case, given that adjustment time scales for pressure gradient and vertical mixing become comparable.
Figure 3-10: Same as Figure 3-4 but for weak wind cases (U=3m/s).
Figure 3-11: Same as Figure 3-5 but for weak wind cases (U=3 m/s).
Figure 3-12: Same as Figure 3-6 but for weak wind cases (U=3m/s).
Figure 3-13: Same as Figure 3-7 but for weak wind cases (U=3 m/s).
Figure 3-14: Same as Figure 3-8 but for weak wind cases (U=3 m/s).

Figure 3-15: Same as Figure 3-9 but for weak wind cases (U=3 m/s).
Figure 3-16: Thermally induced pressure gradient for different background wind strength. (U=0, 3, 6, 9, 12, and 15 m/s, respectively)

3.4.3 Cloud Case

Another important mechanism enabled by the addition of moist process in the atmosphere, the formation of clouds, has been shown in previous studies (e.g. de Szoeke 2005) to have a non-negligible impact on boundary layer evolution, mainly by cloud top radiative cooling, which acts similarly to surface buoyancy flux. As shown in previous studies (e.g., Carson, 1950; Sublette and Young 1996; Young and Sikora 2003; Small 2008), in a highly convective boundary layer, the formation of stratocumulus due to enhanced mixing is prevalent over the warm sea surface (Stull 1988). During the reference transect KnCW1, bands of stratocumulus and cumulus are identified from vertical sounding profiles over the warm side of the Gulf Stream, consistent with common satellite images of the Gulf Stream region that show similar a cloud pattern (e.g. Young and Sikora 2003). In this study, a model run with microphysics and cumulus schemes enabled is compared with the completely wet, cold to warm case in the previous section to analyze the role of cloud in a convective boundary layer.
As shown in Figure 3-17 (e), a significant amount of cloud formation is prevalent on the warm side of the front, with the cloud thickness increasing with distance from the front. Note that the formation of temperature cells is prevalent under clouds, as shown in 3-17 (c). The inclusion of the microphysics scheme enables atmospheric moisture exchange and phase transition, which is assumed to cause the much more complex potential temperature structure.

Compared to the no-cloud case, these temperature cells directly alter the surface air temperature distribution and ultimately cause the surface sensible heat flux to change accordingly. The maximum sensible heat flux is co-located with the cloud topped boundary layer, consistent with observations made in the EPIC (East Pacific Investigation of Climate) section across the Equatorial Front, as discussed in Small (2008). The latent heat flux, on the other hand, shows much smaller variations, as the boundary layer humidity is well mixed, as shown in (d). Note that the boundary layer becomes colder (shown from increased surface sensible heat flux) and moister (shown from decreased surface latent heat flux) with the formation of clouds, which is counterintuitive. It is common for cloud topped boundary layer to cool over land, as the cloud shading will reduce the incoming sunlight and cause the surface to cool, but in this case as the SST is prescribed to be constant over the model window, so cloud shading should not cause such a cooling effect. One possible explanation for the cooler MABL is the downdraft cool air from the free atmosphere, brought down by the cloud mixing. While it is beyond the scope of this thesis to quantify this process, the aforementioned temperature cells between cloud structures shown in Figure 3-18 (c) do cause the surface to be cooler at certain locations. The slightly moister surface layer, while the increase in humidity is small, is also less expected. The formation of clouds should drain moisture from the surface and thus dry the surroundings, yet in this case such an effect is not prominent. The possible explanation for moister surface layer is some potential precipitation events under the heavy cloud region, where the relative humidity is over 100%, yet this process needs further investigation in future studies.

As another product of temperature cells, the wind profile in the cloud-rich region is also less homogeneous, with $U$ and $V$ show a frequent $\sim 1\text{m/s}$ change in magnitude. The
thermally induced pressure gradient is assumed to be the main cause of such variations and is addressed in detail in the later discussion. The change in wind speed also leads to fluctuations in sensible and latent heat flux, as the friction velocity is varying. It is clear that the change in surface wind (Figure 3-17 (e)) is in phase with surface fluxes.

As mentioned in the Methods section, the PBLH estimation is based on humidity profile instead in cloud cases, shown in Figure 3-17 (f). It is clear that the boundary layer growth is significantly higher in the cloud case, making a over 10% higher PBLH in the cloudy region. The vertical mixing mechanism is again considered to the main contributing dynamic. In Figure 3-19, the surface-buoyancy-induced entrainment in the cloud case (red) provides an extra 1-3 mm/s of growth compared to the no-cloud case (blue). Note that this difference is mainly due to the surface buoyancy flux differences shown in Figure 3-17 (d), which ultimately is the product of the enhanced sensible heat flux caused by cooler surface air, rather than cloud top radiative flux. The cloud top radiative cooling is plotted separately as yellow in (d), which is a linear fit of the model diagnostic cloud-top buoyancy flux data. This flux starts to have an impact on the boundary layer at 50km, where a thin layer of cloud is formed, and reaches a maximum of 40 W/m² in the warm region where thick cumulus bands are present. This value is consistent with our estimation of 30-40 W/m² in KnCW1 and common values seen in previous studies (e.g., Zheng et al. 2018). The total combined entrainment in the cloud case is shown in Figure 3-19 as yellow. It is clear that the additional cloud top radiative cooling can support up to 3 mm/s more of entrainment, consistent with de Szoeke (2005), where the author claimed that this process can make up as much as 25% of observed entrainment. Note that while the rate in yellow has reached the upper bound of the 9-14 mm/s estimated entrainment in KnCW1, it is still safe to assume that the extra entrainment from cloud top is valid and necessary, as in KnCW1, it is more often to underestimate the total entrainment rate due to the inconsistent and conserved cross-frontal wind strength estimation.

In Figure 3-20, we can see that in the cloud case the thermally induced pressure gradient is much more complex, due to the aforementioned temperature cells induced by microphysics. The pressure gradient changes sign rapidly in the cloud-topped region, with a magnitude of $10^{-3} \text{ m/s}^2$, due to the horizontal heterogeneity of potential
temperature. We believe such a change is the main cause of the wind variation seen in this region, as it is in phase with the location of temperature cells and wind changes of 1 m/s, shown in Figure 3-17. Such a variation is not seen in the less cloudy region (0 - 150 km), where the horizontal temperature gradient is also smoother. While in the previous section, we argued that in the strong wind case, the pressure gradient is less important in wind change, in this case, however, it is still reasonable to assume that the pressure gradient is dominating the small local wind variability in the cloudy region, as vertical mixing does not provide a valid forcing that can change the wind as shown in Figure 3-17.

In general, the inclusion of cloud processes has been shown to be important. The phase change in the atmosphere and formation of cloud not only provide extra vertical mixing strength via cloud top radiative cooling, but also change the temperature profile inside the MABL, which can lead to changes in both buoyancy flux (by modifying surface sensible heat flux) and pressure gradient (by creating horizontal temperature gradient). The overall agreement with KnCW1 demonstrates the importance of microphysics and cumulus schemes in simulating the boundary layer evolution.
Figure 3-17: Same as Figure 3-4 but for the cloud case. Note that in (d) the cloud-top buoyancy flux from the cloud case is shown in yellow. The straight line represents the linear fit of the model’s diagnostic cloud top buoyancy flux, rather than the raw value.
Figure 3-18: Same as Figure 3-5 but for the cloud case.
Figure 3-19: Same as Figure 3-8 but for the cloud case. The yellow line represents the sum of red and estimated entrainment from cloud top buoyancy.

Figure 3-20: Same as Figure 3-9 but for the cloud case.
3.5 Summary

In this study, a series of idealized 2-D WRF simulations are conducted to study the impact of several mechanisms on the boundary layer evolution near an SST front including stability induced surface stress change, vertical mixing, and thermally induced pressure gradient. Moist processes are turned on and off to separate and examine the contribution of surface latent heat, atmospheric moisture, and cloud formation to boundary layer growth in both strong and weak background wind cases.

In the strong cross-frontal cold-to-warm wind case, the inclusion of surface latent heat flux and atmospheric moisture has been shown to have a significant impact (over 25%) on the boundary layer growth. In this case, the total estimated entrainment rate is closest to our reference observations KnCW1 when the model is initialized with a realistic sounding containing a moisture profile and with surface moisture flux enabled. The moist process also alters the boundary layer height, which indirectly changes the behavior of the thermally induced pressure gradient. The dry sounding with surface latent heat flux case shows a significant boundary layer growth on the thermally stable cold side, indicating the important role of air-sea humidity in producing vertical mixing. The wind change due to pressure gradient mechanism in a strong wind case is less detectable, as vertical mixing provides the majority of momentum gain inside the boundary layer through cold, high momentum air entrainment.

In the weak wind case, moist processes have also been proven to be important in simulating the boundary layer growth accurately. In this case the entrainment is weaker due to the reduced surface buoyancy flux, which entrains less high-momentum air from the free atmosphere. The pressure gradient is significantly higher and concentrated near the front due to the much weaker horizontal advection. Its corelation with wind change is also stronger. This is consistent with Spall (2007) and Kilpatrick (2014), where they argued that the adjustment time scales for weaker wind cases allow a stronger adjustment from pressure gradient force.

In the cloud case, the formation of cumulus and stratocumulus leads to temperature heterogeneity inside the boundary layer, which alters the local pressure gradient and can
be responsible for the more complex wind profile. The formation of clouds also enhances the vertical mixing by providing extra cloud top radiative cooling. A novel result of this research is the counterintuitive result that the boundary layer becomes colder and moister when clouds are formed. Effects of cloud shading on SST can not be a cause of this effect, given the prescribed SST. Possible causes of the cooling are increased cold air downdrafts, while the effects of rainfall evaporation into the boundary layer may increase the moisture. This aspect of the boundary layer response could be a fruitful avenue for further examination.
Chapter 4

WRF 3-D Simulation

Abstract

A realistic 3-D simulation in the Weather Research and Forecast (WRF) model is carried out to study the boundary layer response to underlying sea surface temperature front under a strong (12 m/s) cross-frontal background wind. The model is initialized with data from ERA5, the fifth generation of the European Centre for Medium-Range Weather Forecasts (ECMWF) atmospheric reanalysis of the global climate. A comparison between the model output and observational data in the same region is conducted to test the validity of the model formulation and setup. The overall agreement between data and model suggests that the 3-D model can reasonably represent the boundary layer structure near the Gulf Stream SST front. Key properties, including surface flux, cloud formation, boundary layer growth and related physical dynamics, are consistent in both cases despite that some discrepancies exist. The model output is also compared against a 2-D idealized WRF run, and the results suggest that simplified initialization and model schemes in the 2-D case are still able to re-create the dynamical processes identified in previous studies properly. The inclusion of moisture is shown again to be crucial to achieve a higher level of agreement between idealized runs and realistic simulations. Two main mechanisms, vertical mixing and pressure adjustment, show consistent behavior across all three cases, suggesting that previous analysis on their effects is valid.

4.1 Introduction

As discussed in the previous two chapters, the primary dynamics governing the evolution of the MABL near an SST front have been identified and investigated. In Chapter 2, observations made during the CLIMODE campaign (Marshall 2009) and the PEACH campaign (Seim et al. 2022) have been categorized and analyzed to study the change in boundary layer under different background wind direction and strength, as suggested by Friese (1991). In this study, a large volume of ship-board surface flux measurements and Rawinsonde launches provided a comprehensive in situ dataset that is rarely seen in previous studies. Divided into observation windows with length of one day, these ship transects near the SST front offer invaluable opportunities to study the MABL change on a time scale of 1 day or less, which is much closer to the MABL adjustment time scale of several hours, compared to previous studies that often rely on satellite data or reanalysis.
with coarse temporal and spatial resolution. The four main mechanisms proposed in Small (2008), namely stability induced surface stress change (Wai and Stage 1989), vertical mixing (e.g., Wai and Stage 1988; Spall 2007), pressure gradient adjustment (Lindzen and Nigam 1987), and the surface current effect (e.g., Kelly 2001), have been studied quantitatively to determine their contribution to observed changes in the MABL in the vicinity of the Gulf Stream SST front. Results have shown that, while the stability induced surface stress change and surface current effect contribute less, the vertical mixing inside the boundary layer due to enhanced surface flux and cloud formation over warmer ocean surface, and the pressure gradient and resulting circulation induced by thermal adjustment, can be dominant depending on the background wind strength (e.g., Spall 2007; O’Neil 2010; Kilpatrick 2014; Schneider 2015).

In Chapter 3, a series of WRF idealized 2-D simulations are studied to further separate and analyze the role of each proposed mechanism. Building on a simplified 2-D model in Kilpatrick 2014, this study aims to expand the 2-D setup by including moist processes in the MABL. The role of moisture (surface latent heat flux, atmospheric moisture content, cloud formation) in boundary layer evolution is studied via a series of sensitivity tests with model setups containing different physics schemes. With adjustable background wind strength, the idealized 2-D model provides valuable information on MABL evolution under the effect of a prescribed SST front. Compared to Kilpatrick’s dry model setup, this study focuses on examining the effect of moisture by comparing model results with observations. It has been shown that the vertical mixing mechanism is significantly enhanced with the presence of surface latent heat flux and cloud formation (e.g., Wai and Stage 1989; de Szoeke 2005). Model runs under different cross-frontal wind strengths have also shown that the pressure gradient forcing is much stronger in a weak wind case, consistent with Spall 2007 and Schneider 2015, 2020.

It is still necessary, however, to investigate further the dynamics governing changes in the MABL near an SST front. Due to the limited spatial coverage of the CLIMODE ship track and the time gap between sounding launches, the observational data in Chapter 2 lacks high spatial and temporal continuity. Several assumptions (e.g., a continuous trend of the PBLH) are required to account for the absence of particular data. On the other
hand, the idealized 2-D model offers high temporal-spatial resolution and data availability, and the strength of different dynamics can be differentiated with sensitivity test, but its validity is still to be tested and its initialization is simplified. It is therefore reasonable to study the MABL dynamics in a way that can fill in this gap in the observations as well as justify and validate the idealized 2-D setup.

This chapter thus aims to connect the real observational data and the idealized model results with a WRF Real 3-D simulation. This simulation is initialized with ERA5 data to simulate one of the ship transects in Chapter 2. The main objectives are to validate assumptions made in observations, provide snapshots of the boundary layer during the observation window with a greater coverage, provide extra evidence for the conclusions made in previous chapters, and justify the model scheme by comparing model outputs to both observations and idealized 2-D results. In section 2, the model setup is introduced, with key dynamics and schemes described briefly. In section 3, analysis methods and assumptions made are provided. Section 4 is the results and discussion comparing the 3-D model to both observation and 2-D idealized model. Lastly, section 5 provides a summary of the work.

4.2 Model Setup

A 3-D nested domain is built in the Weather Research and Forecasting (WRF) model, as shown in Figure 4-1. The grid resolutions for the domains are 18 km, 12 km, and 2 km, respectively. Only model fields in the smallest nest are examined in this study. The region of the nest is set to be the same as one of the CLIMODE Knorr Cruise transects (hereafter referred as KnCW1, shown in Figure 4-2) in the Gulf Stream region, for which we have observational meteorological data that can be compared with the model outputs. The observation window for KnCW1 is 21 hours duration, starting on 2AM Mar. 16, 2007. The model thus starts at 12PM Mar. 15 to allow a 12 hr spin-up period. The model is initialized with ERA5, the fifth generation of the European Centre for Medium-Range Weather Forecasts (ECMWF) atmospheric reanalyses of the global climate (Hersbach, Hans, et al. 2020).
The model uses a stretched vertical grid with 80 levels, and 48 of them are below 2000 m to better capture to the vertical structure of the MABL, which typically has height of 1000-2000 m (Stull 1988). The top of the model is at 30 km, while its lowest level is at 12 m. Note that the one of the key differences between this model and its 2-D counterpart discussed in Chapter 2, is that with ERA5 reanalyses, the vertical column is initialized with local data profiles, instead of a single sounding that is repeated for the whole domain. This takes into account the initial difference in the atmosphere at different locations on both sides of the front, and potential MABL structures that are present before the model start. The geostrophic wind in ERA5 vertical profiles shows a general direction of north to west and speed of 7-15 m/s during the observation window, consistent with Rawinsonde-observed wind profile acquired during KnCW1, thus making this simulation a strong cold to warm wind regime.

The SST field in the model uses a daily-updated field from ERA5, with a resolution of 30 km. The resulting SST front in the model thus is wider, which has an average width of 50-75 km for a gradient of 8K. Note that this is larger than the width of observed SST front in KnCW1, which has a width of 25 km at the ship crossing. In 2-D simulations, the SST field is set to a gradient with width similar to KnCW1. The potential difference induced by this will be discussed in later sections.

The physics schemes activated in the model are similar to the model’s 2-D counterpart to ensure they can be compared effectively. The model uses an improved Mellor–Yamada (MY) turbulence closure model (MYNN model: Mellor–Yamada–Nakanishi–Niino model) for the atmospheric side surface layer and boundary layer scheme. The MYNN model runs as a second-order turbulence closure model that considers buoyancy and stability effects with constants determined from a LES data base (Nakanishi and Niino 2009). 1-D equations for mean quantities in MYNN are given by:
\[
\frac{\partial U}{\partial t} = - \frac{\partial}{\partial z} \langle uw \rangle + f (V - V_g), \\
\frac{\partial V}{\partial t} = - \frac{\partial}{\partial z} \langle vw \rangle - f (U - U_g), \\
\frac{\partial \theta_l}{\partial t} = - \frac{\partial}{\partial z} \langle w \theta_l \rangle + \frac{f \theta_0}{g} \left( V \frac{\partial u_g}{\partial z} - U \frac{\partial v_g}{\partial z} \right), \\
\frac{\partial Q_w}{\partial t} = - \frac{\partial}{\partial z} \langle w q_w \rangle,
\]

(4.1)

where \( u, v, w \) the velocities, \( U_g \) and \( V \) the geostrophic wind, \( q_w \) the total water content, \( \theta_l \) the liquid water potential temperature, \( f \) the Coriolis parameter, and \( g \) the gravitational acceleration. Capital letters denote ensemble-averaged variables, small letters turbulent variable, angle brackets for ensemble average and subscript 0 as a reference state (Nakanishi and Niino 2009). The unknown second-order turbulent fluxes are then given by:

\[
\frac{\partial q^2}{\partial t} = - \frac{\partial}{\partial z} \left( w \left( u^2 + v^2 + w^2 + \frac{2p}{\rho_0} \right) \right) \\
- 2 \left( \langle uw \rangle \frac{\partial u}{\partial z} + \langle vw \rangle \frac{\partial v}{\partial z} \right) \\
+ 2 \frac{g}{\theta_0} \langle w \theta_v \rangle - 2 \epsilon^2
\]

\[
\frac{\partial (\theta_l^2)}{\partial t} = - \frac{\partial}{\partial z} \langle w \theta_l^2 \rangle - 2 \langle w \theta_l \rangle \frac{\partial \theta_l}{\partial z} - 2 \epsilon_{\theta l}
\]

\[
\frac{\partial (\theta_l q_w)}{\partial t} = - \frac{\partial}{\partial z} \langle w \theta_l q_w \rangle - \langle w q_w \rangle \frac{\partial \theta_l}{\partial z} \\
- \langle w \theta_l \rangle \frac{\partial q_w}{\partial z} - 2 \epsilon_{\theta q}
\]

\[
\frac{\partial (q_w^2)}{\partial t} = - \frac{\partial}{\partial z} \langle w q_w^2 \rangle - 2 \langle w q_w \rangle \frac{\partial q_w}{\partial z} - 2 \epsilon_{q_w}
\]

(4.2)

where \( p \) is the pressure, \( \rho \) the air density, \( \theta_v \left( \equiv \theta (1 + 0.61 q_v - q_l) \right) \) the virtual potential temperature, and \( \epsilon, \epsilon_{\theta l}, \epsilon_{\theta q}, \text{and} \ \epsilon_{q_w} \) are the dissipation rates of \( \frac{q^2}{2}, \frac{\theta_l^2}{2}, \frac{\theta_l q_w}{2}, \text{and} \ \frac{q_w^2}{2} \), respectively. More detailed parameterizations and can be found in previous studies (e.g., Nakanishi and Niino 2001, 2004, 2006, and 2009).

The radiation schemes implemented in this simulation are Dudhia (Dudhia 1989) for short wave radiations and Rapid Radiative Transfer Model (RRTM) for long wave radiations. Note that while the Dudhia scheme is considered simplistic, it has been shown...
in previous studies (e.g. Stull 1988) that over an ocean surface the diurnal cycle has a lesser impact on MABL evolution due to the large heat capacity of seawater. It is thus safe to argue that the rather simple short wave model should not have a significant impact on simulation results, and can be used as it is in 2-D simulations. For microphysics, the Purdue Lin scheme (Lin, Farley and Orville 1983) is used. The Lin scheme is a 5-class scheme that includes ice and graupel processes. Note that in 2-D simulations, a simpler Kessler scheme is used for reliability and efficiency, but for the 3-D simulation, a more sophisticated scheme is required for a large domain with high resolution. A separate series of runs with Lin scheme was performed in 2-D to compare model outputs with Kessler scheme, and the results showed no significant difference, as in the model region ice process was not prevalent and precipitation was not our focus in this study. The direct comparison between 3-D simulation and 2-D is thus still valid. The cumulus scheme used in this simulation is the Kain-Fritsch (KF) convective parameterization scheme (CPS). A scale-aware version of KF is used to provide more accuracy in the region where the horizontal resolution is high.

Unlike the 2-D simulations, where physics scheme are turned on and off to separate the effects of moist processes, in this simulation all physics schemes are always on to generate a simulation closest to real observations.
Figure 4-1: Nested model domain. The smallest domain corresponds to the domain shown in Figure 4-2. Ocean surface is shown as white while land is colored in heatmap.
Figure 4-2: Ship track for KnCW1 (red), with the SST front identified from GHRSSST (blue) and model slice (black). Cold side is on the north, while the warm side is on the south. Synoptic sea level pressure retrieved from ERA5 is plotted as the colored background. Background wind was blowing predominantly from north to south.
4.3 Methods

Boundary Layer Structure

Key properties regarding the vertical structure of the boundary layer, including potential temperature, wind profile, humidity, are direct outputs of the WRF model. The virtual potential temperature $\theta_v$ we use to describe the MABL state, is defined as:

$$\theta_v = \theta (1 + 0.61r - r_L)$$

where $\theta$ is the potential temperature, $r$ the mixing ratio of water vapor and $r_L$ the mixing ratio of liquid water in the air. Note that while the MYNN scheme provides native diagnostic boundary layer height estimate based on a potential temperature difference between surface level and MABL top, its accuracy is limited on the warm side of the Gulf Stream region, where rampant cloud formation adds extra variabilities at the top of the mixed layer. To obtain a better PBLH estimation, the height where a strong humidity cap exists is picked instead. This takes the cloud structure into account and the estimated PBLH is at the top of the cumulus, while the default MYNN method would pick a level inside the boundary layer as the vertical potential temperature profile is complex and the temperature threshold is less valid. Diagnostic PBLH estimated with the humidity cap method is consistent with our criteria for defining the top of the PBLH in CLIMODE sounding data, which is described below.

In CLIMODE observational data, the top of the MABL is defined as the height where the potential virtual temperature is significantly less mixed, and the humidity of air decreases rapidly to form a cap. In KnCW1, the PBLH estimation is made based on seven Rawinsonde launched during the observation window. A sample sounding profile is shown in Figure 4-3. Due to a lack of dedicated cloud observations during KnCW1, we estimate the cloud structure roughly based on the relative humidity profile.

Note that, while the model output is temporally and spatially continuous, there was a 2-4-hour gap between each sounding launch in KnCW1. In this study, we assume that the general boundary layer structure did not change significantly over this period and changes in MABL structure observed were mostly spatial, as we assumed in Chapter 2.
Note that this assumption is less valid when comparing observation to a model snapshot, which shows the entire MABL structure at a particular time. Extra caution is needed when comparing observations that contain a certain degree of temporal variations to model fields that only contain spatial variations.

As sounding launches provided vertical profiles of key properties in the mixed layer, surface observations were made with ship-board measurements, including bulk and direct covariance estimates of heat flux, surface wind and stress, sea surface temperatures, air temperature, pressure and other common meteorological quantities. These data are continuously available most of the time during the transect, and a 10-minute average is analyzed in this study.

![Graph](image)

**Figure 4-3:** Relative humidity, specific humidity, virtual potential temperature and wind speed profile retrieved from sounding 119, launched during KnCW1. The launch location is inside the warm region. The estimated PBLH is 1000 m in this case, based on the inversion cap level. Relative humidity profile suggests that the cloud base is at 500 m, and the cloud top at 1000 m, where RH reaches over 80%.
Vertical Mixing

The strength of vertical mixing in this study is quantified by the boundary layer growth and entrainment rate, as shown below:

\[ U \frac{\partial z_i}{\partial x} = w_e + w_L, \tag{4.4} \]

where \( U \) is the averaged cross-frontal wind advecting the inversion, \( z_i \) the boundary layer height, \( x \) the distance from the SST front, \( w_e \) the entrainment rate and \( w_L \) the subsidence rate (de Szoeke 2005). The subsidence rate during the observation window is estimated with The Modern-Era Retrospective analysis for Research and Applications, version 2 (MERRA-2), which has a magnitude of around 1 mm/s. In a convective boundary layer, a convective velocity scale can be defined as:

\[ w_* = \left[ \frac{g z_i}{\overline{w' \theta'_v}} \right]^{1/3} \tag{4.5} \]

where \( \overline{w' \theta'_v} \) the buoyancy flux at the surface \( \overline{\theta_v} \) the depth-average virtual potential temperature. The surface heat flux forced entrainment then can be estimated by:

\[ w_e = \frac{A w_*^3}{h \Delta b} \tag{4.6} \]

where \( \Delta b = g \Delta \theta_v / \theta_v \) is the buoyancy jump capping the inversion in virtual potential temperature, \( A \) the faction of surface flux contributing to entrainment and, \( h \) the boundary layer height. The range of \( A \) is 0.25 to 0.35 based on previous studies (e.g., Pino et al. 2003; de Szoeke 2005; Conzemius et al. 2006).

For shear-induced entrainment, Tennekes (1973) proposed that, in a pure shear-driven boundary layer, the entrainment rate can be estimated as:

\[ \frac{dh}{dt} = \frac{2.5 \Theta_0 u_*^3}{g \Delta \theta_v h} \tag{4.7} \]

where \( \Theta_0 \) is the initial average temperature across the boundary layer. Note that while
the MABL in KnCW1 was not purely shear-driven, this value should still be a valid rough estimation for shear-inducedentrainment, as it functions similarly to Eq.4.6, but scales with $u_3^3$ instead of $w_3^3$.

Another important component of entrainment is the cloud top radiative cooling (Stull 1988, de Szoeke 2005), which has the form of:

$$e_{Cloud} \cong \frac{\Delta I^*}{\Delta E_Z \theta_v} \quad (4.8)$$

where $\Delta I^*$ is the net long-wave radiative flux divergence near the cloud top and $\Delta E_Z \theta_v$ the virtual potential temperature difference across the entrainment zone (Stull 1988). In KnCW1, direct cloud measurement was not available, and the radiative flux is estimated based on typical values in mid-latitude regions in previous studies (e.g., Zheng et al. 2019). In the WRF model, this flux is obtained with built-in MYNN cloud top flux estimation:

$$\omega_l = \left[ \frac{g}{\theta} \left( W' \theta' \right) z_l z_l \right]^{1/3} \quad (4.9)$$

which is similar to $w_3$ from surface heat flux but uses cloud radiative fluxes instead.

With all components combined, the sum of calculated entrainment is then compared to the observed PBLH change $U \partial z_i / \partial x$ to quantify the contribution from vertical mixing.

**Thermally Induced Pressure Gradient**

The thermally induced pressure gradient near the SST front can be approximated by:

$$P_t = \frac{g}{\Theta} \ h \ \frac{[\theta]}{\partial x}, \quad (4.10)$$

where $[\Theta]$ is the vertically integrated potential temperature from the surface up to the boundary layer height $h$, and $\Theta$ is a scale value of the potential temperature (Mahrt et al. 1982, 2004). In KnCW1, this gradient is calculated for every adjunct sounding launch, which has an average $\partial x$ of 40km. In the WRF model, the pressure gradient is
calculated as a 20km horizontal moving average based on corresponding vertical temperature profiles.

**Comparison with Observation and 2-D Model**

To compare with the observational data that are sorted as 2-D function of distance from the front, a slice of the model field is selected, as shown in Figure 4-2 (black line). While it is possible to follow the ship track in the model, the coarse resolution of the SST field (30 km) in the model cannot resolve the 25km wide 8-degree SST gradient we observed in KnCW1. The meander of the Gulf Stream creates a local thin filament of cold water near the front, while in ERA5 this feature is smoothed out and only a 4-degree difference in SST is present at this location. Thus, a direct comparison between the observation and model along the KnCW1 track is invalid as the initial SST condition is significantly different. The chosen model field slice along 54 °W, on the other hand, shares a wider but overall similar SST gradient with KnCW1. The model background wind direction and strength are roughly the same at these two locations. ERA5 reanalysis shows no significant synoptic difference between the two tracks either. We thus assume that this model slice should be a qualitatively similar representation of the MABL structure along KnCW1. The distance from the front for the model slice is calculated following the same procedure described in the Methods section in Chapter 2.

The same model slice is also used to compare with the full physics 2-D simulation, as the 2-D run is set up with KnCW1 SST profile as its reference initial condition. Note that in the 2-D simulation described in Chapter 3, the cross-frontal wind component is in positive U (westerly), while in the 3-D simulation, the cross-frontal wind is in overall negative V (northerly) with some variations. To avoid confusion, in later parts of this study, wind directions will be addressed as cross-frontal/along-frontal.
4.4 Results and Discussion

4.4.1 Comparison with Observations

The general structure of the boundary layer in both 3-D model simulation and in situ observation is shown in Figure 4-4. Compared to the observation, the SST profile in the model has a similar but more gradual 8-degree gradient over 50km, while the observed SST gradient from ship measurements is narrower (25km wide). The cold side of observed SST shows a much cooler region from -25 km to 0 km, which represents a cool water filament induced by a nearby meander. The MABL in this particular region is overall stable due to the negative air-sea temperature difference and resulting negative sensible and buoyancy flux, shown in Figure 4-4 (b), (c), and (e). This local stable boundary layer is induced by the warmer air from the north by the background wind. In the model slice, such a warm over cold regime is not present as the SST profile is smoothed out due to the 30km resolution of SST in ERA5, and thus the focus of the comparison will be on the warm side of the front.

The largest difference between the model and the observation is apparent at the air-sea temperature difference shown in Figure 4-4 (b), where the observed temperature difference reaches a maximum of 6 K at 25km, where the SST front ends, and gradually decreases to zero further away from the front as the air temperature reaches equilibrium with the sea surface, while the model difference is more persistent, still above 5 degrees at 160 km. This marks an overall 2 to 3 degrees higher \( T_{\text{sea}} - T_{\text{air}} \) in the model. While it is reasonable to argue that this discrepancy is due to that the model slice is showing the MABL structure at locations other than the ship track, as being discussed in the Methods section, we argue that the overall similar SST distribution and initial air temperature should not yield such a difference as the initial condition in both cases is similar. Thus, we believe that the discrepancy is more due to the temporal change in MABL over the observation window. As discussed in Chapter 2, the KnCW1 track lasted 21 hours, and while we assume that most variations observed were due to change in space, the air temperature reaching equilibrium on the warm side over time is inevitable. The 4 hours lapsed as the ship moved from 0km to 50km can be responsible for the observed lower
air-sea temperature difference. In Chapter 2, due to limited measurement availability, we assume this process to be secondary. But in this case, this process leads to the major difference we see in Figure 4-4 (b). This is not to argue that the comparison is invalid, however, as observations regarding the MABL on a time scale of 1 day or less are often based on qualitative assumptions (e.g., Mahrt 2004) due to the highly turbulent nature of the mixed layer (Stull 1988). In this case, the dynamics behind observed and simulated MABL variations should still be consistent and are the focus of the analysis.

The sensible heat flux, shown in Figure 4-4 (c), shows a significant increase in phase with underlying SST, implying the contribution from the air-sea temperature difference. The observed sensible heat flux is slightly higher than the model value at the front, despite the fact that the air-sea temperature difference is smaller in KnCW1. This is mainly due to the higher surface wind in the observation, which is 5 to 8 m/s near the front, compared to the model value of 4 to 5.5 m/s, shown in Figure 4-4 (e). Due to poor ship-wind alignment, surface heat flux measurements were not available beyond 120 km and only one data point is available at 190km, but the overall trend in sensible heat flux is in phase with the decreasing air-sea temperature difference. The wind variation beyond 50 km is small in the observation, indicating that the observed heat flux change is mainly due to changes in temperature. In the model, however, the sensible heat flux does not decrease with distance significantly while the air-sea temperature difference is decreasing, as the surface wind is increasing with distance. The latent heat flux shows a higher level of agreement between the observation and the model, which is consistently high (200W/m²) over the warm side. The more persistent high latent flux in the observation during KnCW1 can be evidence showing that the air-sea humidity difference takes longer for the ocean surface to adjust, compared to the fast-decreasing air-sea temperature difference. The latent heat flux in both observation and model shows a high positive correlation with SST after the variation in surface wind being removed, suggesting that this is an ocean-forcing-atmosphere scenario, consistent with Small (2008), which argues that the ocean drives the atmosphere on a smaller spatial scale. The difference in buoyancy heat flux (panel (e)) at the surface is mainly due to the difference in sensible heat flux. In the observation, the buoyancy flux at the front (25 km) is higher due to higher friction velocity $u_*$ and stronger air-sea temperature difference, reaching a
maximum of 150 W/m². It then decreases rapidly with distance, following the trend in sensible heat flux, and reaches near zero at 200 km, where the remaining buoyancy flux is largely due to the air-sea humidity difference. The model buoyancy flux, on the other hand, increases linearly with SST at the front, and remains at a high level of 150 W/m² beyond the vicinity of the SST front (>100 km), mostly due to the persistent large air-sea temperature difference (>5 degrees), while the increasing surface wind speed plays a secondary role.

The difference in surface heat flux ultimately leads to different levels of vertical mixing strength, which is quantified here by the entrainment rate. The surface buoyancy flux induced entrainment rates for both observation and model are shown as blue in Figure 4-5 and 4-6. It is clear that in the observation, the initially high buoyancy flux provides the majority of entrainment, hence the rapid PBLH growth in the 0-50 km region. The more stable PBLH height beyond 100 km is consistent with the lowered entrainment rate due to weaker surface buoyancy flux. In the model, the buoyancy flux also contributes to over 75% of the entrainment, but its magnitude remains significant farther into the warm region. Apart from the surface flux driven entrainment, the shear driven entrainment, estimated with Eq.4.7, is much less impactful in both the observation and the model, contributing less than 2 mm/s. In the model, the shear driven entrainment is low near the front due to the slower surface wind but becomes comparable to the observed value of 1-2 mm/s in the warmer region, where the surface wind speed is similar in both cases. The cloud top radiative cooling, compared to shear driven entrainment, is much more significant. In KnCW1 we assume a 30-40 W/m² of cloud top flux, and based on the sounding data, the resulting cloud top induced entrainment gradually increases with distance, as the capping inversion becomes weaker when the MABL erodes into the free atmosphere (Stull 1988), shown in Table 4-1. This increase is consistent with the model, where the linear fit of the MYNN diagnostic cloud flux induced entrainment shows a similar trend. The cause of the increase in the model is also mainly due to the weaker inversion, which is on average 4 degrees near the front, and gradually decreases to 2 degrees at 150 km. Note that at 200 km in the observation, the capping inversion is particularly weak, leading to a much higher cloud induced entrainment. The relatively stable boundary layer height at this point suggests that this is likely an overestimation.
Table 4-1: Capping inversion observed from sounding launches during KnCW1. Soundings 1-5 correspond to orange in Figure 4-5, from left to right, respectively.

<table>
<thead>
<tr>
<th>Sounding 1</th>
<th>Sounding 2</th>
<th>Sounding 3</th>
<th>Sounding 4</th>
<th>Sounding 5</th>
</tr>
</thead>
<tbody>
<tr>
<td>Capping inversion (K)</td>
<td>3.60</td>
<td>3.73</td>
<td>3.14</td>
<td>2.91</td>
</tr>
</tbody>
</table>

The combined entrainment, shown in purple in Figure 4-5 and 4-6, is within the range of the estimated entrainment based on Eq.4.4 (dashed lines), which implies that the vertical mixing is the main contributor to observed MABL deepening in both the observation and the model. In the observation, excluding the likely overestimation of cloud flux at 200km, the decreasing entrainment rate with distance is consistent with the near-stable boundary layer height beyond 100km in Figure 4-4 (g). The highest entrainment is also co-located with the fastest deepening at 50km. In the model, the steady increase in PBLH at 0-150km agrees with the high total entrainment in the region. Note that while in Figure 4-4 (g) the PBLH in the model does not show significant growth at 100-150km, in an extended figure (Figure 4-7), the PBLH is still increasing, consistent with the high buoyancy flux and calculated entrainment rate in this region. It is worth noting that the distance from the front calculation in Figure 4-7 beyond 200km is less valid due to a large wind-front angle, and thus is only for qualitative reference.

While the calculated total entrainment is still slightly less than the observed PBLH growth, which can be due to dynamics on different scales that add additional entrainment, it is reasonable to argue that the majority of the deepening is from the vertical mixing mechanism, as the combined entrainment rate matches the growth of the PBLH. Following Small (2008) and Schneider (2015), we can thus assume that this entrainment brings high momentum air from the free atmosphere, which ultimately increases the surface wind speed. In Figure 4-4 (f), the surface wind speed change is in phase with the observed PBLH growth in (g). In the observation, the strongest increase occurs at 0-25km, where the entrainment and PBLH growth are also the most rampant. The surface wind is less variable further into the warm region, in concert with the steadier boundary layer height and reduced entrainment rate. In the model, the surface wind increases gradually starting from the front at 0 km, and the consistent increase is in phase with the
aforementioned steady increase in boundary layer height. The agreement between the observation and model indicates that the vertical mixing mechanism is highly responsible for the observed wind change at the surface.

The thermally included pressure gradient $P_t$ is shown in Figure 4-8. A high level of variation is shown in the model, while the observation made during KnCW1 provides limited information due to the limited amount of sounding measurements. The order of magnitude in both cases is consistent with common values at mid-latitude in previous studies (e.g., Mahrt, 2004; Kawai 2014), yet the correlation with surface wind change is highly limited in both cases. Following Spall (2007) and Kilpatrick (2014), we assume that under such a strong cross-frontal wind, the contribution of thermally induced pressure gradient is limited as the air does not have sufficient time to adjust to underlying SST. As a comparison, the pressure gradient from weak wind cases (3 m/s) in Chapter 2 and Chapter 3 is about 2 to 3 times higher than the value shown in Figure 4-8. The main difference between the model and the observation, being that the model $P_t$ is persistent beyond the vicinity of the front while in the observation $P_t$ falls to near zero, can be due to the aforementioned temporal adjustment in the observation. The column of air reaches equilibrium during the 3-4 hour long observation gap, leading to a reduced horizontal heterogeneity in temperature, and the resulting $P_t$ is lower, while in the model, the horizontal temperature difference at a specific time is directly associated with the underlying SST and resulting air-sea temperature difference.

In general, while the model has some non-negligible differences compared to the in situ measurement from KnCW1, the key dynamics and resulting MABL evolution are consistent between the 3-D model and the observation. Surface fluxes are realistic compared to the observation after considering the difference in initial conditions. The vertical mixing induced entrainment is comparable in both cases, with its effect on PBLH growth and surface wind change being similar. The pressure gradient forcing in the model agrees with the observation and typical values. These agreements with the reference KnCW1 indicate that the 3-D WRF simulation is a valid and realistic representation of the MABL evolution near the SST.
Figure 4-4 SST, air-sea temperature difference, sensible heat flux, latent heat flux, buoyancy heat flux, surface wind velocity and PBLH, for model (blue) and KnCW1 observations (orange), from top to bottom, respectively. KnCW1 observations were made with ship-board measurements and seven sounding launches during the 21 h long observation window. The ship track started on the cold side (left) and moved into the warm region (right). Negative distance represents the cold side of the front.

Figure 4-5 Surface buoyancy flux driven entrainment (blue), shear driven entrainment (orange), cloud top cooling driven entrainment (yellow) and total combined entrainment (purple) for KnCW1. The estimated total entrainment based on Eq.4.4 is shown as a range between dashed lines (9-14mm/s).
Figure 4-6 Same as Figure 4-5 but for the 3-D simulation.

Figure 4-7 Extended boundary layer height profile for the 3-D simulation. Distance beyond 200 km is for reference only.
4.4.2 Comparison with 2-D Simulation

The 2-D simulation to be compared with the 3-D case has an initial SST profile similar to KnCW1, with a background wind of 12m/s blowing from cold to warm, comparable to the 7-15 m/s wind observed in KnCW1. As shown in Figure 4-9 (a), the SST front in the 2-D simulation is narrower (25km vs 50km), while the change in SST is comparable (8 degrees). The air-sea temperature difference, however, is considerably different in these two cases. The 2-D $T_{sea} - T_{air}$ is close to zero on the cold side, as the sounding used to initialize the atmosphere is prescribed to be neutral to cold side SST, while in the 3-D model where the atmosphere column is initialized with local ERA5 data, the air is slightly unstable, marking a 20 W/m² sensible heat flux over the cold ocean surface. On the warm side, both 2-D and 3-D show a persistent air-sea temperature difference, consistent with our expectation that the strong cross-frontal wind would bring cold air over warm sea surface continuously. This again indicates that the fast-recovering $T_{sea} -$
\( T_{\text{air}} \) in KnCW1 is more likely due to temporal changes. Note that the 3-D \( T_{\text{sea}} - T_{\text{air}} \) is on average 2-3 K higher than 2-D in the 25-150km region. The reason why the 3-D simulation has higher \( T_{\text{sea}} - T_{\text{air}} \) compared to the 2-D simulation can be due to several reasons. Firstly, the air column in the shown 2-D slice could be more adjusted to the underlying SST at the time the snapshot is taken (model time 12 hours after initialization). Snapshots at earlier time do show higher \( T_{\text{sea}} - T_{\text{air}} \), but for the consistency in comparison, 12 hours model time is chosen for all cases, as the MABL structure at this point is more developed, shown in Chapter 3. Secondly, while the 3-D model slice is also taken after 12 hours, continuous perturbations on different scales that are only present in the WRF Real simulation can make the cross-frontal air transport more variable, thus creating a different \( T_{\text{sea}} - T_{\text{air}} \). Lastly, as the 2-D case is initialized with one sounding, the initial temperature profile, while it is sampled from KnCW1 sounding profiles, is likely to be different from the ERA5 profiles used to initialize the 3-D case.

The magnitude of the sensible heat flux at the surface in both cases is comparable, despite the difference in air-sea temperature difference. The overall higher surface wind speed in the 2-D case provides enhanced friction velocity that contributes to the surface heat flux during the first 100km on the warm side. Sensible heat flux in the 3-D surpasses the 2-D value again in the region where the surface wind speed is similar, and \( T_{\text{sea}} - T_{\text{air}} \) dominates the difference in sensible heat flux. The latent heat flux, like in the previous section, is more consistent between two cases, suggesting that the air-sea humidity difference is more resilient and less affected by spatial and temporal perturbations in the vicinity of the SST front. The 2-D case has a higher latent heat flux reaching 300 W/m\(^2\) in the 0-50km region, mainly due to the enhanced surface wind, but the 3-D value becomes close as the wind speed increases. The overall buoyancy flux at the surface, combining the sensible and latent heat flux, is close in these two cases, which allows us to further compare its effect on entrainment and PBLH growth.

The structure of the bottom 2000 m in both models is shown in Figure 4-10 and 4-11. The overall wind structure (panel (a), (b)) is similar in both cases, despite the fact that the 3-D simulation has more variability. A dome-like structure outlining the MABL is seen in
cross-frontal and along-frontal wind. A high (low) speed cell near the surface on the cold side can be seen in the along-frontal (cross-frontal) wind profile due to rotation and frictional force, consistent with model results seen in Kilpatrick 2014. The surface wind starts to turn anti-cyclonically as it crosses the front, indicating the proper inclusion of the Coriolis effect (O’Neil 2010). The wind speed is higher above the MABL dome, supporting the vertical mixing mechanism that requires the mix-down of high momentum air from the free atmosphere. The virtual potential temperature profile (panel (c)) is more complex in the 3-D simulation, due to the formation of clouds. The inversion cap is also stronger in the 3-D case (3-4 degrees) compared to the 2-D value (2 degrees on average). The weaker capping inversion enables a stronger surface flux driven entrainment in the 2-D case, shown in Figure 4-12, despite that the surface buoyancy flux is close in both cases. The stronger virtual potential cap is likely induced by the much stronger cumulus formation (panel (e)), which creates a stronger gradient at the top of the boundary layer (Stull 1988). Note that while both cases show clear bands of cloud formation at the top of MABL over the warm side, the 3-D case shows cumulus with a higher fraction and a lower base. This is likely due to that ERA5 initializations contain cloud information, while the 2-D sounding is sampled on the stable, cold side without pre-existing clouds. During the 12-hour spin up, the 3-D simulation has cloud formation based on pre-existing clouds, while the 2-D simulation generates less thick cumulus despite having similar latent heat flux from the surface. Sounding profiles from KnCW1 (Figure 4-7) also suggest that the low-base, tall cumulus in 3-D simulation is closer to real world observations. Despite the difference in cloud fraction, both simulations manage to capture the feature of bands of stratocumulus/cumulus formation that is prevalent over the warm side of the Gulf Stream (e.g. Young and Sikora 2003).

In terms of the vertical mixing strength, the entrainment for surface heat flux is stronger in the 2-D model (Figure 4-12), despite that both cases have similar buoyancy flux at the surface (100 -150 W/m²). The main cause of the difference is the weaker capping inversion discussed in the previous section. The shear driven entrainment is also higher than the 3-D value (2 mm/s vs sub 1 mm/s) due to the same reason, especially in the 20-40km region where the capping inversion is the weakest, despite that the model estimate of friction velocity $u_*$ is comparable in both cases (0.2–0.6 m/s). Note that near the front,
the more rapid wind acceleration (Figure 4-9 (f)) also contributes to the higher shear driven entrainment there. The cloud top radiative cooling, due to the less rampant cloud formation, is much smaller in the 2-D simulation. In the 3-D simulation, the thicker cloud bands in the 40-100km region provide as much as 50W/m² of buoyancy flux, which contributes to 2 mm/s of entrainment with a capping inversion of 3 K on average. In the same region in the 2-D model, only a thin layer of stratocumulus is present and while the capping inversion of 2 degrees is weaker, the resulting entrainment from the much smaller radiative flux (10 W/m²) is only of O(0.1 mm/s). The cloud induced entrainment gets stronger in both cases as the cloud fraction increases over the warmer sea surface, and the capping inversion gets weaker as the erosion of MABL goes deeper into the free atmosphere. Despite that the strength of each component differs in two cases, the total combined entrainment is within the range of the rate estimated with Eq.4.4, shown in Figure 4-6 and 4-12. The continuous high total entrainment is consistent with the trend in PBLH growth shown in Figure 4-9 (g) for both 2-D and 3-D simulations, suggesting that the vertical mixing is the main contributor of MABL growth. It is different, however, from the previous comparison between the observation and the 3-D simulation, where the surface wind increases with PBLH growth. In the 2-D case, the wind change is only significant near the front, and beyond 50 km the surface wind speed becomes stable, despite that the PBLH is continuously growing and more high momentum air is mixed down. This can be due to that the initial surface wind is higher in the 2-D case (7-8 m/s on the stable side), and the momentum of free atmosphere wind (12 m/s) is not much higher than the surface wind on the warm side (10 m/s) after spin-up. The high momentum air being mixed down can only contribute a limited amount of acceleration due to the smaller difference, while in the 3-D simulation the difference in surface wind and free atmosphere wind is larger (over 7m/s), and the mixing downward creates a much more continuous acceleration in concert with MABL deepening.

The thermally induced pressure gradient is on the same order of magnitude (0.5 × 10⁻³ m/s²) in both cases, while the 2-D case has less variability due to the simpler model setup. The low magnitude and variation are consistent with the stable surface wind in the 2-D case, suggesting that in a strong cross-frontal wind scenario, the effect of pressure gradient adjustment is limited. The 2-D and 3-D simulations both show a higher $P_t$ in the
cloud-topped region (beyond 140 km in 2-D, beyond 40 km in 3-D), suggesting that the thermal budget within the boundary layer is more complex due to the horizontal heterogeneity induced by atmospheric moist process, as discussed in Chapter 3. Those variations in $P_t$ are co-located with surface wind variations, suggesting that $P_t$ can be responsible for some shown surface wind adjustments, but the contribution is still much less significant compared to vertical mixing.

In general, the 2-D and 3-D case show considerable consistency in key dynamics. The two major dynamics, vertical mixing and thermally induced pressure gradient, do not differ significantly (order of magnitude differs less than 25%) in both cases. The model PBLH in two simulations agrees with our combined entrainment estimation, which implies that our arguments and conclusions made in Chapter 3 in a 2-D setup are valid in a more realistic 3-D setup. The main discrepancies main come from the different initialization methods, but major physical processes do not show major disagreement.
Figure 4-9 Same as Figure 4-4 but for the 2-D and 3-D comparison.
Figure 4-10 Cross-frontal, along-frontal wind, virtual potential temperature, specific humidity and cloud fraction profile for the 3-D model field under 2000m.
Figure 4-11 Same as Figure 4-10 but for the 2-D simulation.
Figure 4-12 Same as Figure 4-5 but for the 2-D simulation.

Figure 4-13 Same as Figure 4-8 but for the 2-D and 3-D comparison.
4.5 Summary

A WRF Real 3-D simulation is carried out to study the behavior of the boundary layer in the vicinity of the Gulf Stream SST front. A slice of the model field is selected for comparison with a reference ship transect KnCW1 from the CLIMODE Knorr cruise conducted in March 2007. The same slice is also compared with a 2-D WRF simulation based on the initial condition observed in KnCW1.

Compared to the observation, the 3-D simulation shows a considerable consistency in key dynamics and MABL responses. While discrepancies exist due to the differences in sampling methods in the model and observations, the evolution of the boundary layer follows key mechanisms discussed in previous chapters. Key features, including cumulus formation, increased wind stress, stable boundary layer on the cold side, are shown in the 3-D simulation in concert with the observation. Components of the total entrainment are comparable in both cases given the similar initial condition and developed PBLH, suggesting that the vertical mixing is the main contributor to observed boundary layer changes. The surface wind shows a significant correlation with the PBLH growth, indicating the importance of the mixing down of high momentum air. The thermally induced pressure gradient’s effect is less prominent in both cases, consistent with our expectation that in a strong cross-frontal wind scenario, the pressure gradient adjustment is secondary.

Compared to the 2-D simulation, the more realistic initialization with ERA5 data creates a much more realistic model field with added synoptic and large-scale variabilities. The cloud formation in the 3-D simulation is more prevalent due to the pre-existing cloud in local sounding profiles. MABL’s response to enhanced surface buoyancy flux on the warm side is similar in both cases, with the mean difference being the strength of the capping inversion, which is assumed to be caused by the difference in initialization. Physical dynamics in both models drive the MABL evolution in a consistent way, and the resulting boundary layer structure is comparable in both cases, suggesting that the key dynamics discussed in Chapter 3 are presented in the 2-D simulation with validity.
The overall agreement between the 3-D simulation, the observation and the 2-D simulation indicates that the WRF model setup is a valid representation of the boundary layer and its behavior near the Gulf Stream SST front. This study thus reinforces the previous conclusions on the linkage between MABL change and suggested dynamics, including moist process, cloud formation, vertical mixing, and thermally induced pressure gradient.
Chapter 5

Conclusions

5.1 Thesis summary

This thesis aims to study the marine atmospheric boundary layer in the vicinity of an SST front, particularly in the region of the Gulf Stream. Vertical mixing and thermally induced pressure gradient are the two processes proposed to be the major contributors to boundary layer change. To investigate their contribution quantitatively, both in situ observations made in the Gulf Stream region and idealized model simulations in WRF are analyzed.

In Chapter 2, observations made during the CLIMODE campaign and the PEACH cruise are studied. Ship transects crossing the SST front are categorized into three groups depending on their background wind direction: strong cold to warm, parallel/weak, and warm to cold. Both shipboard data and metrological data retrieved from Rawinsonde launches are analyzed to identify the boundary layer changes due to the SST front. In the cold to warm wind case, boundary layer height and surface wind both show a significant increase over the warm sea surface. Enhanced surface buoyancy flux fuels vertical mixing significantly, resulting in a major entrainment at the top of the boundary layer. It is also shown that the cloud top radiative cooling and shear induced entrainment also contribute to the observed PBLH change in a non-trivial way. The thermally induced pressure gradient is relatively small in this case, consistent with previous studies (e.g. Mahrt 2004, Kilpatrick 2014), likely due to the shorter adjustment time in a strong advection scenario. The observed surface wind change does not show a significant correlation with the estimated pressure gradient. Note that in a specific transect, the passing synoptic system (a cold-air outbreak) alters the boundary layer dramatically, resulting in completely different boundary changes compared to other transects, which notes the importance of larger scale dynamics in modifying local boundary layer near an
SST front. In the parallel/weak wind case, the vertical mixing is weakened due to smaller surface fluxes, and the magnitude of thermally induced pressure gradient is significantly larger (over 70%) compared to the strong wind case, indicating that the adjustment timescale for these two mechanisms can be comparable due weaker advection (Spall 2007). Note that while the correlation between pressure gradient and surface wind change is stronger, such a relation is not evident everywhere due to the temporal and spatial gap between observations. Overall, the boundary layer growth estimation from each dynamic combined is reasonably consistent with the observed value, indicating the dominant role of said mechanisms. In the warm to cold case, only qualitative analysis can be done due to the limited vertical resolution inside the stable boundary layer, but the overall dynamics and changes are consistent with our expectation and previous studies (e.g., Garratt 1987).

In Chapter 3, a series of 2-D WRF simulations are studied to further examine the role of the aforementioned dynamics in boundary layer evolution in a more controlled, idealized setup. Moist processes are turned on and off to identify their impact on the MABL. In all wind cases, the addition of surface latent flux and atmospheric moisture exchange corrects the estimated entrainment and makes it closer to the reference value retrieved from observations. The formation of clouds enhances vertical mixing further by providing cloud top radiative cooling. The model diagnostic PBLH is significantly higher (over 40%) when all moist processes are enabled. The strength of vertical mixing is significant in both strong and weak wind cases, but the thermally induced pressure gradient shows completely different behaviors depending on cross-frontal wind strength. In the stronger wind case, the pressure gradient is more spread-out and has a smooth trend, while in the weak wind case the gradient is much stronger near the front, but decays rapidly deeper into the front, due to the much smaller horizontal air transport.

In Chapter 4, a 3-D simulation with realistic initialization is conducted to compare with the 2-D cases and observations. The overall boundary layer structure in 3-D agrees well with 2-D, as well as reference observations. Small discrepancies exist, but don’t change the key dynamics that this thesis focuses on.
The key insight of this thesis arises when the results from Chapter 2, 3 and 4 are combined. Observations from Chapter 2 alone are realistic but lack high temporal and spatial resolution due to the nature of field experiments. The boundary layer profile contains valuable information, but it is hard to distinguish the contribution of each underlying dynamic. The superposition of larger scale dynamics and varying SST front location also complicate the observational data. Model results from Chapter 3 and 4, on the other hand, have much higher resolutions and availability of data, but are idealized products of a simplified model. The comparison between these data, however, enables a much more thorough analysis of the boundary layer evolution. Using the observation as a reference, this thesis has shown the non-negligible contribution from moisture, which was less of a focus in previous studies. The enhanced vertical mixing due to surface latent heat, atmospheric moisture exchange and cloud formation is seen in both observations and model results. The impact of the thermal pressure gradient is also shown in model results to be directly related to background advection and indirectly related to moisture (by PBLH modification). While in observation such a relation between pressure gradient and background wind is less clear, in model simulations with sensitivity tests on background wind strength, this relation is obvious. The 3-D simulation bridges the idealized 2-D case and reference observations, justifies and validates arguments this thesis makes. This thesis as a whole provides non-trivial advances in understanding the boundary layer evolution and its underlying dynamics on a time scale of 1 day or less, with an emphasis on differentiating the contribution of each potential governing mechanism, also with an extensive discussion on the inclusion of moist processes, which is less seen in previous studies.

5.2 Future Work

While this thesis provides advances and insights on the boundary layer evolution near an SST front, there is a significant amount of future work can be done to expand the scope of this study.
As mentioned previously, the analysis of the warm to cold wind case is limited due to the lack of SBL observations. In Chapter 3, WRF simulations do not cover warm to cold case either, as the model resolution is not enough to resolve the fine structure inside a usually 100m high SBL. Future efforts thus should focus on providing a much more comprehensive and quantitative analysis on this case, potentially by either conducting high-resolution observations (e.g., lidar measurements) or numerical modeling capable of resolve SBL structure (e.g. LES). Note that dominating mechanisms (vertical mixing and pressure gradient) discussed in cold to warm and parallel/weak wind case act differently in a SBL. Entrainment rate, for example, is entirely different in a stable layer. Thus, it is important to develop a different but consistent set of methodologies to describe and analyze the boundary layer evolution over a cool sea surface.

Another direction for future work is the 3-D simulation in WRF. While in this thesis the 3-D simulation acts only as a validation and justification for 2-D cases, the potential of it is great. With proper setups, 3-D WRF simulations can reproduce 2-D cases in Chapter 3 with much more realistic dynamics. The addition of synoptic systems and other large-scale dynamics that are not included in 2-D cases can also be implemented. The cold air outbreak in Chapter 2, for example, can be better understood with a 3-D realistic setup. One major issue with observations in this thesis is the location and time of observations were both changing as the ship moved to the next observation site, which makes data points to be much less spatial-temporal continuous. The 3-D model solves this issue by providing the boundary layer structure across the whole domain at any given time, which provides a much more rigid estimation of entrainment. It is also possible to follow the ship track in the model, to better compare the observation and model results.

More 2-D cases can be also tested for a more comprehensive coverage of wind regimes. While in this thesis, Chapter 3 has covered a series of cases, they are mostly based on two reference transects during the Knorr cruise. As shown in Chapter 2, several other transects in the Pilot cruise also have their unique characteristics. Ship tracks and variability in the SST front were also different in these two cruises. While the overall governing dynamics should not change, it is of interest to run simulations based on these transects to provide more insights on the MABL evolution.
Lastly, this thesis mostly follows the methodology in Chapter 2, which is to compare the estimated values of proposed dynamics with observed values. While this approach is valid and robust, and fits the observational data we are studying, a more theoretical analysis (e.g., Kilpatrick 2014) based on thermal, momentum and TKE budget can provide more quantitative insights into each dynamic. While in observations it is harder to achieve due to the lack of certain types of measurements, in 2-D and 3-D models, such an approach should be executable.
References


