Coastal Zone Surface Stress with Stable Stratification

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ABSTRACT

Summertime eddy correlation measurements from an offshore tower are analyzed to investigate the dependence of the friction velocity \( u_* \) for stable conditions on the mean wind speed \( V \), air–sea difference of virtual potential temperature \( \delta \theta_v \), and nonstationary submeso motions. The quantity \( \delta \theta_v \) sometimes exceeds 3°C, usually because of the advection of warm air from land over cooler water at this site. Thin stable boundary layers result. Unexpectedly, \( u_* (V) \) does not depend systematically on the stratification \( \delta \theta_v \) even for weak winds. For weak winds, \( u_* (V) \) increases systematically with increasing submeso variations of the wind. The \( u_* (V) \) relationship for a given \( V \) is greater in nonstationary conditions. Additionally, this study examines \( u_* (V) \) as a function of wind direction. The \( u_* (V) \) relationship appears to be affected by swell direction for weak winds and advection from land for short fetches.

1. Introduction

Over the sea, much of the variation of the surface stress can be described by the wind vector alone. Grachev et al. (2003) found a strong relationship of the surface stress to the mean wind speed \( V \), while Rieder and Smith (1998) found a strong relationship of the stress to \( V^2 \). The \( V \) refers to the speed of the vector-averaged wind. Several recent papers have found new utility in relating the friction velocity \( u_* \) to \( V \) (Foreman and Emeis 2010; Edson et al. 2013). In fact, competitive models of the sea surface stress can be formulated in terms of only \( V \) (Andreas et al. 2012; Vickers et al. 2015).

Andreas et al. (2012) and Edson et al. (2013) found that \( u_* (V) \) can be described in terms of a weak wind regime, where \( u_* (V) \) increases slowly with increasing \( V \), and a stronger wind regime, where \( u_* \) increases more rapidly with increasing \( V \). Both regimes are approximately linear, and the transition between them occurs at 8–10 m s\(^{-1}\). The transition is not as sharp as that found in the nocturnal boundary layer over land (Sun et al. 2012). The transition wind speed over land was on the order of 1 m s\(^{-1}\) at 1-m height and increased with height to about 3.5 m s\(^{-1}\) at 10 m.

a. Stratification

Stable boundary layers over the sea differ from clear-sky, nocturnal boundary layers over land where stable stratification is generated by strong radiative cooling at
the ground surface. Over the sea, stable stratification is generated by warm-air advection, surface evaporation, modest surface radiative cooling of the sea surface (Fairall et al. 1996), and downward transport of warmer air toward a cooler sea surface. Over the sea, large warm-air advection requires significant wind speed, which precludes the very stable conditions that generally occur only in weak winds (large Richardson number).

On the other hand, even modest stable stratification can restrict downward mixing of momentum, which in turn limits the development of the wind-driven sea. The reduced surface stress may still lead to large vertical divergence of the stress because the boundary layer is thinner over colder water with stable stratification as detailed by Samelson et al. (2006). The resulting very small surface roughness and $u_*$ leads to only limited coupling between the atmosphere and sea surface (Högström and Smedman-Högström 1984; Smedman et al. 1997a,b; Donelan 1990; Mahrt et al. 2001; Skillingstad et al. 2005). Some turbulence may be maintained by the downward transport of turbulence energy (Mahrt et al. 2001; Fairall et al. 2006; Vickers et al. 2013). However, rigorous examination of such conditions requires fluxes very close to the surface because the boundary layer is quite thin. This precludes use of most aircraft and ships (Fairall et al. 2006) for these conditions and is even a difficulty for fixed masts because the instruments must be above the maximum expected wave height.

b. Wave effects

Wave effects may significantly influence the $u_*-V$ relationship, especially for weak winds. See Högström et al. (2013) for a recent example. Wave effects are thought to become small for measurements made at levels above one significant wave height (Hristov and Ruiz-Plancarte 2014). Rieder and Smith (1998) and Grachev et al. (2003) found that removing the wave-correlated atmospheric fluctuations led to a better defined dependence of the stress on the wind speed and led to closer agreement between the wind and stress direction. A number of studies have found that $u_*(V)$ is greater for flow against the swell, as in Drennan et al. (1999). Vickers and Mahrt (1997) and Grachev et al. (2011) found that the wind stress for offshore flow against the swell is significantly larger than the theoretical prediction of $u_*$. The stress due to swell does not vanish with vanishing wind speed because there is still a relative velocity difference between the swell and the air. With vanishing wind speed, swell can generate upward momentum flux (Smedman et al. 1994; Grachev and Fairall 2001; Sullivan et al. 2008).

c. Nonstationary submeso motions

Rieder and Smith (1998) noted the potential impact of the nonstationarity of the wind on the drag coefficient. For weak large-scale flow, nonstationary submesos winds can intermittently enhance the turbulence (Mahrt et al. 2013), sometimes because of the flow profile distortion, transient near-surface wind maxima, wind directional shear, and apparent inflection point instability. The nonstationarity of the wind is associated with submeso motions nominally on time scales of minutes or tens of minutes for stable stratification. Sometimes submeso motions are alternately defined as nonturbulent motions on spatial scales less than 2 km. Such air motions include numerous wavelike perturbations and more complex structures.

The atmospheric turbulence appears to decay more slowly with decelerating flow compared to the rapid generation of turbulence with accelerating flow; these effects collectively enhance the average $u_*$ for a given $V$ (Mahrt et al. 2013). Over a water surface, the nonstationarity of the wind leads to alternating wind-driven generation and decay of short waves on the water surface (Uz et al. 2002). Nonstationary wave enhancement of $u_*$ for a given $V$ results from the expectation that short surface waves decay more slowly with decelerating atmospheric flow compared to the wind-driven development of short surface waves with accelerating flow (Uz et al. 2002). Vickers and Mahrt (1997) found increased $u_*$ with broader spectra of the wave field (no dominant wave mode) often associated with light winds and changing wind direction.

d. Analysis plan

Our goal is to examine the relationship of $u_*(V)$ to the air–sea temperature difference, nonstationary submeso motions, fetch, and wave state. After describing the data, we determine the averaging strategy (section 3) and a method for quantifying time changes of the wind as an aid for quantifying nonstationary submeso motions (section 4). We then investigate the vertical structure of the thin boundary layer in short-fetch conditions (section 5) using near-surface flights by the LongEZ to assist us in interpreting the measurements from the air–sea interaction tower (ASIT) south of Martha’s Vineyard. Section 6 examines the relationship between $V$ and the air–sea temperature difference for a fixed level on the tower and then examines the relationship between $u_*$ and $V$ to the air–sea temperature difference and fetch. Section 7 inferentially studies the potential effect of swell. Finally, we examine the important relationship of $u_*$ to the nonstationary submeso motions (section 8). The collective relationship of $u_*$ to $V$, the air–sea
temperature difference, submeso motions, offshore advection, and swell are considered in the conclusions with the concession that completely isolating these interrelated influences requires more extensive measurements.

2. Data and analysis

a. LongEZ data

We analyze data collected by the LongEZ aircraft operated by Tim Crawford for the pilot program of the Coupled Boundary Layers and Air Sea Transfer (CBLAST) experiment (CBLAST Weak Wind) conducted over the Atlantic Ocean south of Martha’s Vineyard, Massachusetts, during July–August 2001 (Fig. 1). The data were collected at a rate of 50 samples per second with a flight airspeed of about 50 m s\(^{-1}\), which corresponds to a horizontal interval of about 1 m between data points. Winds were measured using the Best Atmospheric Turbulence Probe (BAT; Crawford and Dobosy 1992) positioned 2 m in front of the nose and five wing widths ahead of the canard. Fast-response temperature was measured using a 0.13-mm microbead thermistor mounted inside the design stagnation point port on the BAT hemisphere. The height above the sea surface was measured by a NovAtel GPS sensor and calibrated with a Riegl laser (LD90–3100 VHS). The LongEZ instrumentation is described further in Sun et al. (2001) and Mahrt et al. (2014). We analyze the 13 flights with sufficient measurements below 12 m.

Based on the dependence of the fluxes on the width of the averaging window, we choose 100 m as the width of the averaging window. This is the smallest averaging length that included all of the momentum flux. Using larger averaging windows potentially includes non-turbulent motions as part of the perturbation flow. This inclusion can cause the local random flux error to be converted to systematic error in the computation of \(u_\text{w}\) (Mahrt 2010). For our study of stable conditions, we require that the air–sea virtual temperature difference \(\delta \theta_v\) is greater than 0.1 K and require that the flight level be below 12 m.

b. ASIT data

We also analyze data from the ASIT collected during the CBLAST experiment in late summer of 2003 (Edson et al. 2007). The offshore tower is located 3 km south of Martha’s Vineyard in 15 m of water (Fig. 1). The 20-Hz turbulence measurements were collected using 3D sonic anemometer (CSAT3; Campbell Scientific, Inc.). Here, we analyze data from the CSAT3 measurements approximately 6 m above the mean sea surface to calculate eddy correlation fluxes of momentum and virtual heat flux. The exact height above the sea varied with the tidal cycle and occasionally with storm tides. No corrections were made for transducer shadowing (Horst et al. 2015). Slow-response measurements of temperature and moisture (Vaisala HMP 35A sensors) at 7 m are used for the mean temperature and specific humidity of the atmosphere. The sea surface temperature (SST) was measured by a Heitronics KT15 radiometer using an Eppley precision infrared radiometer (PIR) to correct for sky reflections. In addition to the nominal quality control that eliminated meteorologically impossible values, data with wind direction between 0° and 150° were eliminated to reduce the effects of flow distortion by the tower.

For physical understanding, estimating the stratification within the atmosphere based on the temperature difference between two atmospheric levels on the tower would be preferable to using the air–sea temperature difference because it would include short-term variations of the stratification that are not necessarily captured using the SST, particularly for stable conditions when the communication between the air and SST can be slow. However, because the lowest temperature level is 7 m, only the atmospheric stratification above 7 m can be estimated. Temperature gradients measured on the tower are in the middle of the boundary layer for the short-fetch cases, as in the example in section 5. In addition, the temperature differences between two levels on the tower are relatively small and vulnerable to instrument errors.

Our study, therefore, estimates the stratification as the difference between the 9-m virtual potential temperature and the surface virtual potential temperature, symbolized here as \(\delta \theta_v\). The surface virtual potential temperature is simply computed in terms of the saturation-specific humidity for the sea surface temperature. We implicitly assume that stratification of the
air near the surface is related to $\delta \theta_w$. The impact of water vapor is included because the turbulence kinetic energy is converted to potential energy through the buoyancy flux. Our use of the term “stratification” will refer to $\delta \theta_v$. Although $\delta \theta_v$ can be significantly different from $\delta \theta$ for individual data points, these differences have little influence on the averages used in our study.

We have also analyzed wave data from a subsurface node 1.5 km offshore in the Atlantic recorded by Martha’s Vineyard Coastal Observatory (MVCO; http://www.whoi.edu/mvco). During the field program, the swell propagation direction was almost always out of the south with phase speeds between 6 and 8 m s$^{-1}$. We use this information in interpreting our analyses of the tower measurements. However, we are uncertain if the small variations of swell propagation speed and direction are within the accuracy of the data and therefore do not include analysis of these small variations. There are not enough data with swell propagation from other directions to satisfy even lenient sampling criteria. Regrettably, the wave measurements are not collocated with the ASIT tower and therefore do not allow us to isolate the wave-coherent velocity fluctuations.

The fetch from land for the ASIT tower can be as short as 3 km with northerly flow. The fetch is less than 5 km for about 20% of the measurements. For these short fetches, $\delta \theta_v$ averages about double that for fetches longer than 5 km. The $\delta \theta_v$ is often still positive with the longer fetch flow from the south or southwest, possibly because of flow from warmer water over cooler water. The horizontal variation of sea surface temperature is sometimes concentrated in microfronts with changes of several degrees kelvin (Vickers and Mahrt 2006).

For the shortest fetches of 3 km in northerly flow from Martha’s Vineyard to the ASIT tower, the largest turbulent eddies from the heated land surface may not have completely decayed upon arrival at the tower site, as can be inferred from the results of Sahlée et al. (2014).

3. Averaging

The choice of averaging time and the analysis techniques, briefly presented below, are discussed in more detail in Mahrt and Thomas (2015). Turbulent perturbations are computed as the deviation from an average over a window width of $\tau$. Choosing an averaging time $\tau$ that is significantly greater than the time scale of the largest transporting turbulent eddies leads to perturbation quantities that include nonturbulent motions and omits some of the nonturbulent flow in the $V$ calculation. As a result, $V$ does not completely capture all of the motions generating the turbulence. A $\tau$ value that is too large includes significant nonturbulent flow as part of the perturbations and contaminates the computed relationship between the turbulence and the mean flow. The 1-min averaging window for the ASIT data includes almost all of the flux for weak and moderate wind speeds. The 1-min averaging begins to exclude detectable flux for stronger winds where transporting eddies become more elongated in the wind direction, as inferred from the aircraft data (not shown), and are advected past the tower more slowly. For the aircraft data, we choose an averaging length of $\tau = 100$ m, which includes almost all of the momentum flux.

The flow is then partitioned as

$$\phi = \phi' + \bar{\phi},$$

where $\phi$ is one of the velocity components, $\bar{\phi}$ is the average over the averaging time $\tau$, and is the deviation from such an average. The vertical flux of $\phi$ for a given averaging window is then

$$w'\phi'.$$

We use a Cartesian coordinate system where $v$ is directed toward the north. The surface friction velocity $u_*$ for the given averaging window is computed as

$$u_* = (w'u'^2 + w'v'^2)^{0.25}.$$

Therefore, we include both the along-wind and cross-wind stress. With the above nomenclature, the wind speed for each averaging window is computed as

$$V = (\bar{u}^2 + \bar{v}^2)^{0.5}.$$ 

For some of the calculations, additional averaging is applied to reduce scatter, as is explained in the next subsection.

Interval (bin) averaging

The small averaging time of 1 min, compared to more usual values of 5 min or more, reduces bias in the relationship between the turbulence and the mean flow but results in larger scatter. Therefore, for some of the analyses, we sorted the 1-min values of $u_*$ into different intervals of a mean variable, such as wind speed, and then averaged the values of $u_*$ within each of those intervals. We discarded intervals with fewer than 20 samples, although most intervals include a large number of samples, generally more than 1000.

The standard errors for each wind speed interval are generally small compared to the variation between intervals because of the large number of samples. However, the standard error might significantly
underestimate the uncertainty because the assumption of independence is sometimes strongly violated (many of the samples for a given interval of $V$ can come from adjacent averaging windows). Correcting the standard error estimate for this dependency (Leith 1973; Wilks 2006) is difficult because the required autocorrelation function is not well defined and highly variable for the weaker wind conditions.

Interpretation of the interval averaging implicitly assumes a cause and effect relationship. For example, plotting $u_*$ for different intervals of the wind speed assumes that the wind speed is an important forcing mechanism responsible for the generation of $u_*$. For nonstationary weaker winds, cause and effect relationships are more difficult to establish because the turbulence and the nonturbulent wind profile interact on time scales that are not large compared to the largest turbulent times scales. This contrasts with the more usual case where the mean shear is considered to be time independent and generates turbulence without rapid interaction with the turbulence.

Our analysis also includes bivariate bin averaging (Williams et al. 2013) such as an average of all of the values of $u_*$ that simultaneously occur for a specific interval of $du_y$ and a specific interval of $V$. The requirement of 20 samples per joint interval eliminates some of the coverage in $V$–$du_y$ space but provides new perspective.

4. Submeso velocity variations

To examine the potential impact of submeso motions, we relate $u_*$ to submeso changes of the “mean” wind on time scales of minutes, discussed in more detail in Mahrt and Thomas (2015). Recall that the time series is partitioned into averaging windows of a width of 1 min, and $V$ and $u_*$ are computed as the magnitude of the vector-averaged components for each 1-min window. Two-point differences of $V$ are computed centered about the $u_*$ calculation for the $i$th averaging window (Fig. 2) such that

$$\delta V(i) = V(i + k) - V(i - k);$$

that is, the velocity difference centered on the $i$th averaging window is computed between the $k$th averaging windows before and after the $i$th averaging window. Our study chooses $k = 3$, corresponding to a time separation of 6 min between the centers of the two averaging windows. Our goal is to assess the nonstationarity on time scales just larger than the largest turbulent eddies. We do not choose $k = 1$ because the same structures could simultaneously contribute to both the fluctuations on scales less than 1 min and to $\delta V$ on scales greater than 1 min, leading to a form of artificial correlation. The choice of $k = 3$ is a compromise. The relationship between $u_*$ and $\delta V$ decreases only gradually with increasing $k$. Our study only briefly considers the sign of Eq. (5). Therefore, in the remainder of the paper, $\delta V(i)$ will refer to the absolute value of Eq. (5) unless otherwise noted.

5. Aircraft measurements of the vertical structure

The atmospheric boundary layer in the summertime coastal zone is sometimes too thin to study with aircraft. As a result, aircraft can substantially underestimate surface fluxes. We briefly examine this issue for short-fetch, north-northwest flow using data from the LongEZ aircraft, which was able to fly close to the surface.

The example in Fig. 3 represents a thin boundary layer measured by eight flight legs located 6–10 m above the sea surface and three legs around 20 m above the surface. The value of $u_*$ decreases with height and appears to reach a very small residual value of about 0.02 m s$^{-1}$ around 20 m, perhaps zero within the flux uncertainty. This vertical structure is consistent with aircraft soundings shown in Figs. 3c and 4c of Mahrt et al. (2014).

Subjective extrapolation of the vertical structure in Fig. 3 to the surface indicates that even measurements at 5 m above the sea sometimes miss significant surface momentum flux. More typical flight levels of 30 m or higher would miss the boundary layer entirely for this
short-fetch case. For this reason, only LongEZ flights legs below 12 m are used in our analysis, although even flights this low may miss significant flux in thin boundary layers. Based on concurrent aircraft soundings for this flight (not shown), the wind speed decreased sharply above 25 m implying a low-level jet. Variability of \( u^* \) between legs in Fig. 3 is partly due to random errors that can be substantial, as can be inferred from Martin and Bange (2014).

6. The \( u_a(V) \) relationship

We now explore the relationship between \( u_a(V) \) and the air–sea temperature difference with the recognition that both the aircraft and the 6-m tower level can underestimate \( u_a \) in the thin short-fetch boundary layers. But first, we examine the relationship between \( \delta \theta_v \) and \( V \).

a. The \( \delta \theta_v-V \) interdependence

The stratification is forced at least partly by warm-air advection from land over cooler water or by warm-air advection from warmer water to the south. Important stratification generally requires significant advecting wind speed. For the weakest winds, \( \delta \theta_v \) is generally, but not always, less than 1.5 K. Stronger winds lead to greater mixing and potentially reduces \( \delta \theta_v \). However, for the tower data, \( \delta \theta_v \) increases gradually with increasing \( V \) (Fig. 4). This result suggests that the effect of larger warm-air advection with stronger winds exceeds the reduction of \( \delta \theta_v \) by the increased mixing. The interplay between these two processes prevents large Richardson number.

b. The relationship between \( u_a(V) \) and stratification: Spatial averages

We examine the dependence of \( u_a \) on \( V \) by averaging the 100-m averages of wind and stress components over the entire flight and then computing \( u_a \) from these flight-averaged components (Fig. 5a). Alternatively, the 13 flight averages are constructed by directly averaging the 100-m values of \( u_a \) and \( V \) over the entire flight (Fig. 5b). The values of \( u_a \) and \( V \) in Fig. 5a are smaller than those in Fig. 5b because averaging the components for meandering wind and stress vectors leads to some cancellation due to the sign changes of the components within the averaging. However, the differences between the two calculations are relatively small except for the flight with the weakest wind speed (far-left red point). These “flight averages” produce only 13 values of \( u_a \) and \( V \), but each value is based on a large number of measurements.

The value of \( u_a \) based on these flight averages does not exhibit a clear dependence on \( V \) for \( V < 5 \text{ m s}^{-1} \) (Fig. 5), suggesting asymptotic nonzero values of the stress for vanishing wind speed. For \( V \) greater than 4 \text{ m s}^{-1}, \( u_a \) increases systematically with increasing \( V \) collectively for the near-neutral (cyan) and stable (black points) flights.

c. Behavior of \( u_a(V) \) based on the time averages

All analyses for the remainder of this study are based on the ASIT tower data (Fig. 1) where the sample size is much larger than that for the aircraft data. To examine
using the tower data, we partitioned the measurements into two wind direction groups and three groups of $\delta \theta_u$ (Fig. 6). Offshore northerly flow is defined as wind directions between 330° and 360°. Northeasterly flow is contaminated by flow through the tower and excluded from analysis. Onshore southerly flow is defined as wind directions between 150° and 225° and corresponds to long fetches of several hundred kilometers or more, depending on the curvature of the flow trajectory. This flow will be thought of as open-ocean conditions subject to possible shoaling. For $V < 4 \text{ m s}^{-1}$, $u_a(V)$ at the 6-m level on the tower is on average significantly greater for flow with a northerly flow component compared to flow with a southerly component (cf. the dashed lines with the solid lines in Fig. 6). The $u_a$ might be augmented by the propagation of swell mainly from the south against the northerly flow and reduced $u_a$ in wind following swell from the south. We return to this potential influence of swell in section 7. Advection of turbulence from land might also augment $u_a$ for offshore flow, although this augmentation disappears for stronger flow when advection of turbulence should be stronger. It is not known if the emergence of the dependence of $u_a(V)$ on the stratification for stronger offshore flow is due to advection from land. The impact of swell and advection of turbulence cannot be isolated with the current data.

For the northerly flow, the slope transition in $u_a(V)$ (see the introduction) might be about 1.5 m s$^{-1}$ (Fig. 6, dashed lines) if it is possible to determine such a transition with only one averaged value on the weak wind side. For southerly flow (solid lines), $u_a(V)$ shows a slope transition between 2 and 3 m s$^{-1}$, smaller than that for the aircraft measurements that included all wind directions and heights up to 12 m. Our work below concentrates on wind speeds weaker than the transition value.

The influence of $\delta \theta_u$ (Fig. 6) on $u_a$ is detectable only for northerly flow with stronger winds when $u_a$ for the subclass of smallest $\delta \theta_u$ (green dashed, Fig. 6) is 10%–15% larger than $u_a$ for the subclass of largest $\delta \theta_u$ (red dashed). The direct impact of stratification is unimportant for weaker winds.

Although the overall impact of $\delta \theta_u$ on $u_a(V)$ is undetectable for weaker winds in the present dataset, $z/L$ is greater than 0.5 for about 15% of the observations with stable stratification, where $L$ is the Obukhov length. The value $z/L$ becomes relatively large partly due to small $u_a$ associated with small surface roughness over the sea. The variation of $z/L$ between 0 and 0.8 accounts for a factor of 4 increase of the non-dimensional shear $\phi_m$ for this dataset (Edson et al. 2013). Why does this calculation indicate an important impact of stability and yet the influence of $\delta \theta_u$ on $u_a(V)$ is generally not evident? The $z/L$ is an internal stability parameter that depends on the turbulence itself. The variation of $u_a$ tends to dominate variation of both $\phi_m$ and $z/L$. Evidently, $\phi_m$ can depend significantly on $z/L$.
even though the direct dependence of \( u_0 \) on the air–sea temperature difference is weak.

The relationship between \( u_0 \) and \( \delta \theta_v \) cannot be necessarily thought of as a one-way cause and effect. The primary effect of stratification may be indirect through reduced downward momentum flux from higher levels (Shah and Bou-Zeid 2014), a consequent reduction of the surface wind speed, and thus smaller \( u_0 \). In addition, downward transport of heat could maintain warmer air temperatures near the surface and positive \( \delta \theta_v \) because the SST does not respond to the surface heat flux on short time scales. This downward transport might contribute to larger \( \delta \theta_v \) with stronger turbulence and reduce the presumed decrease of turbulence with increasing \( \delta \theta_v \).

We now examine the bivariate variation of \( u_0 \) (section 3) in \( \delta \theta_v-V \) space (Fig. 7) after excluding short-fetch wind directions between 330° and 360°. Values of \( u_0 \) were jointly binned into boxes defined by intervals of \( V \) and \( \delta \theta_v \), and then averaged. In Fig. 7, \( u_0 \) does not depend systematically on \( \delta \theta_v \). The failure of \( u_0 \) to generally decrease with increasing \( \delta \theta_v \) for weaker winds contrasts with existing surface-layer similarity theory.

The weak relationship between \( u_0 \) and \( \delta \theta_v \) for weaker winds and the general absence of very small values of \( u_0 \) imply that frictional decoupling (see the introduction) is rare for this dataset. The \( u_0 \) for the smallest interval of \( V \) still averages to values greater than 0.05 m s\(^{-1}\). The frictional decoupling may be absent because the warm-air advection is weaker compared to that for the decoupling cases in the literature because the land–sea temperature contrasts were less at the ASIT site than in the some of the other studies. Also, flow against the swell, non-stationary submeso motions, and possible advection of turbulence from land for the shortest fetches in the present dataset may all contribute to the turbulence energy, as explored below.

7. Wind direction and wave state

The larger \( u_0(V) \) for northerly flow compared to southerly flow in this study (Fig. 6) is consistent with the impact of the swell that normally propagates from the south with a phase speed of about 6–8 m s\(^{-1}\) at this site. The observed difference of averaged \( u_0 \) between the northerly and southerly flow decreases with increasing \( V \) (Fig. 6), which might be due to the expected decreased impact of swell for stronger winds and the expected increased contribution of the wind-driven waves to \( u_0 \). The increase of \( u_0 \) for the interval of weakest winds with a northerly component (Fig. 6) is consistent with the influence of the wave state because the effect of swell on \( u_0 \) is expected to be greatest for the weakest winds. Advection of turbulence from land could not be numerically evaluated but cannot be ruled out as a significant influence on \( u_0(V) \) for northerly flow.

The averaged value of the north–south momentum flux \( \overline{w} \overline{v} \) (section 3) for the weakest southerly flow, which follows the faster northward-moving swell, vanishes for small but finite southerly flow \( v \) (Fig. 8). This behavior is consistent with the impact of swell for the weakest wind cases where the wind-driven downward

![Fig. 7](image1.png)

![Fig. 8](image2.png)
flux of $v$ is partially cancelled by upward flux of $v$ generated by the swell.

The stress direction is roughly aligned with the wind direction for most of the measurements (Fig. 9), although the scatter is significant partly because of the short 1-min averaging. For the northwesterly flow, the stress direction tends to be directed more from the north than the wind direction. For southwesterly flow, the stress direction tends to be directed more from the south than the wind direction. These tendencies are consistent with the east–west orientation of the swell crests that is expected to enhance the stress component in the direction perpendicular to the crests.

These deviations of the stress direction from the wind direction do not depend significantly on the wind speed (cf. maroon dashed line with green dashed line in Fig. 9). It is not known if the swell for large $V$ can modify the stress direction more than it can modify the stress magnitude. Deviations of the stress direction from the wind direction could also result from modification of the shear direction by baroclinicity (Geernaert et al. 1993), Ekman effects, and height-dependent advection. The data do not allow us to evaluate these processes. However, the analysis in this section generally supports the expected influence of swell on the stress for weak winds.

8. Nonstationary submeso variations

Nonstationary submeso motions can augment $u_*$ ($V$) because the smallest submeso motions may not be completely resolved by the 1-min-averaged wind components so that $u_*$ is larger for a given value of the resolved $V$. In addition, nonstationary winds can more effectively generate turbulence compared to stationary winds with the same averaged $V$ (Mahrt and Thomas 2015). This enhancement is evidently due to nonstationary distortion of the wind profile. Here, we quantify nonstationary submeso motions in terms of the absolute value of the wind speed change over 6-min periods (section 4). For $V < 4$ m s$^{-1}$, $u_*$ increases systematically with increasing $\delta_1 V$ (Fig. 10a), and the rate of increase increases with increasing $\delta_1 V$. Grachev et al. (2011) reported that the stress based on velocity fluctuations from 10-min averages was typically greater than the stress estimated from the inertial dissipation method for light winds and noted the potential contribution of low-frequency submeso/mesoscale motions to the stress computed from the velocity fluctuations.

The increase of $u_*$ with increasing $\delta_1 V$ at the ASIT tower is not indirectly due to a relationship between $\delta_1 V$ and $V$ because $\delta_1 V$ actually decreases slightly with increasing $V$ at this site. The submeso activity might be larger for shorter-fetch conditions because submeso motions generated over land can be advected out to sea. However, at this site, $\delta_1 V$ is relatively independent of wind direction (not shown) except for a modest increase with fetches less than 5 km. For fetches greater than roughly 5 km, $\delta_1 V$ does not decrease with further increase of fetch even up to fetches greater than a few hundred kilometers. The submeso motions could still be partly related to land because ducted gravity waves in the atmospheric boundary layer can propagate long distances. An example of particularly coherent wavelike phenomena in the stably stratified boundary layer near the ASIT tower can be found in Mahrt et al. (2014).
The value of $u_0$ is not systematically related to $\delta \theta_v$ (Fig. 10b), as also observed in section 6. The weak relationship between $u_0$ and $\delta \theta_v$ and the importance of nonstationary submeso motions ($\delta I_V$) suggest reconsidering similarity theory for weak winds.

9. Conclusions

For the coastal zone site of this study, $u_0(V)$ for weak winds is significantly larger for more active nonstationary submeso motions. However, $u_0(V)$ is not systematically related to the air–sea difference in virtual potential temperature $\delta \theta_v$, even for values of $\delta \theta_v$ greater than 3°C. For offshore flow against the swell, $u_0(V)$ is larger compared to that for onshore flow. However, advection of the turbulence from land could contribute to the larger $u_0$ at the tower for short-fetch offshore flow. Short-fetch cases include thin, stable boundary layers with depths of only a few tens of meters.

For weak winds, $u_0(V)$ increases with increasing submeso variability even when the submesos are resolved by the calculation of $V$. Evidently $u_0(V)$ is greater with nonequilibrium conditions due to profile distortion including near-surface wind maxima, inflection points, and wind directional shear, as found in previous studies over land (see the introduction).

The relationship of $u_0(V)$ with $\delta \theta_v$ is very weak except for short-fetch offshore flow with $V$ stronger than the transition value of about 4 m s$^{-1}$. Although the relationship between $u_0(V)$ and $\delta \theta_v$ is weak and undetectable for most cases, the variation of $z/L$ accounts for relatively large variation of $\phi_m(z/L)$ for this dataset (Edson et al. 2013). However, $z/L$ is an internal stability parameter that is strongly influenced by variation of $u_0$ and is not well correlated with the stratification. This realization may explain why $\delta \theta_v$ can be relatively unimportant yet the turbulence and nondimensional shear are sensitive to the value of $z/L$.

These results must be interpreted with caution because of measurement and analysis uncertainties (section 2). With shallow, offshore boundary layers, even the 6-m observational level may underestimate the surface flux.

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