Evolution of the Velocity Structure in the Diurnal Warm Layer

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ABSTRACT

The daily formation of near-surface ocean stratification caused by penetrating solar radiation modifies heat fluxes through the air–sea interface, turbulence dissipation in the mixed layer, and the vertical profile of lateral transport. The transport is altered because momentum from wind is trapped in a thin near-surface layer, the diurnal warm layer. We investigate the dynamics of this layer, with particular attention to the vertical shear of horizontal velocity. We first develop a quantitative link between the near-surface shear components that relates the crosswind component to the inertial turning of the along-wind component. Three days of high-resolution velocity observations confirm this relation. Clear colocation of shear and stratification with Richardson numbers near 0.25 indicate marginal instability. Idealized numerical modeling is then invoked to extrapolate below the observed wind speeds. This modeling, together with a simple energetic scaling analysis, provides a rule of thumb that the diurnal shear evolves differently above and below a 2 m s$^{-1}$ wind speed, with limited sensitivity of this threshold to latitude and mean net surface heat flux. Only above this wind speed is the energy input sufficient to overcome the stabilizing buoyancy flux and thereby induce marginal instability. The differing shear regimes explain differences in the timing and magnitude of diurnal sea surface temperature anomalies.

1. Introduction

On days with weak winds and a significant input of solar radiation into the upper ocean, a shallow diurnal warm layer (DWL) forms within the mixed layer (Fig. 1). This layer exhibits a near-surface temperature gradient due to the vertical divergence of penetrating radiation. The consequent stable stratification decouples the mixed layer and the DWL, with the latter capturing the momentum input from the wind. Whereas mixed layers typically have velocity differences of 0.05 m s$^{-1}$ between the surface and the bulk of the mixed layer, the presence of a DWL can induce velocity anomalies 5 times larger (Kudryavtsev and Soloviev 1990).

Termed the diurnal jet, the near-surface velocity anomaly has been observed to be 0.1–0.3 m s$^{-1}$ (Price et al. 1986; Kraus 1987; Sutherland et al. 2016; Shcherbina et al. 2019). By isolating the near surface, DWLs (and rain layers) make the surface slippery. For both types of layers, Shcherbina et al. (2019) showed that surface water was typically advected 3 km farther per day than water at 30 m. The shear that occurs between the diurnal jet and the mixed layer (0.03 s$^{-1}$; Sutherland et al. 2016; Bogdanoff 2017) is comparable to that found in estuarine flows (0.05 s$^{-1}$; Stacey and Pond 1997), at the base of internal solitary waves (0.05 s$^{-1}$; Moum et al. 2003), and in the sheared layer above the equatorial undercurrent (0.02 s$^{-1}$; Smyth et al. 2013).

Under weak forcing (wind < 2 m s$^{-1}$), clear sky, and high sun angle, the DWL is confined to the top 1–2 m (e.g., Soloviev and Lukas 1997; Ward 2006). By mid-afternoon, sea surface temperature (SST) can increase by order 1°C, with variations primarily caused by water clarity and insolation. The increased surface temperature increases heat fluxes through the air–sea interface (Matthews et al. 2014), whereas the increased stratification reduces turbulent mixing in the remnant mixed layer below the DWL. Turbulence dissipation rates in this remnant layer, for example, can drop by two orders of magnitude from their nighttime values (Brainerd and Gregg 1993; Moulin et al. 2018).

Under stronger winds, DWLs are less surface intensified and their associated SST anomalies are smaller because shear-induced mixing enhances heat transfer away from the surface. Indeed, heat flux and the turbulence induced by shear are closely related, so parameterizing heat transfer will require knowledge of...
DWL-induced shear. For example, Sutherland et al. (2016) show that the standard scaling of turbulence dissipation ($\varepsilon \propto u^3_*$, where $u_*$ is the friction velocity) better matches observations if it is adapted to include a linear proportionality to the observed shear ($\varepsilon \propto u^2_* S$).

DWLs also modify momentum transfer. Downward transfer of the momentum originating from wind is inhibited (Kondo et al. 1979), thereby altering the classical, unstratified Ekman transport relation and introducing external scaling parameters such as the insolation and the heating period as a fraction of the Coriolis period. A DWL is an inherently time-dependent process with dynamics that are not easily amenable to a physically meaningful, analytical solution. For example, to link time-averaged upper ocean currents to an effective, time-independent viscosity, Price and Sundermeyer (1999) had to invoke a complex-valued viscosity. As they note, the profile of a complex viscosity is as complicated as the current profile itself. Similarly, Wenegrat and McPhaden (2016) derived a mean viscosity that was significantly different from the mean of the time-varying values.

Horizontal differences in forcing, wind in particular, will lead to horizontal gradients in DWL properties and consequently horizontal gradients in upper ocean current structure. DWLs can therefore provide a mechanism for submesoscale and mesoscale horizontal mixing of surface waters by creating regions of convergence and divergence (see Bogdanoff 2017, chapter 4). Indeed, DWLs have a range of scales. They may vary over kilometers due to internal waves (Soloviev and Lukas 1997), or alternatively organize over more than 1000 km (Bellenger and Duvel 2009).

Explicitly simulating DWLs in an ocean model requires high vertical resolution. Even 1-m spacing near the surface is inadequate for low wind scenarios. A climate model with a 10-m grid near the surface therefore needs to parameterize DWL physics. This is typically achieved, if attempted at all, by including a sublayer with idealized vertical profiles of temperature $T(z)$, and sometimes salinity and velocity $u(z)$. Several idealizations have been proposed for use in either climate models or operational procedures like SST corrections. These include linear $T$ and $u$ (Fairall et al. 1996), $T$ and $u \propto z^{-1/3}$ (Fine et al. 2015; Large and Caron 2015), and $T \propto e^{-z}$ with either a wind speed–dependent depth scale (Gentemann et al. 2009) or a depth-dependent phase lag (Matthews et al. 2014). These latter two studies illustrate the difficulty of idealizing even just the temperature profile: it is both depth and time dependent and its shape differs with wind speed.

Compared to the number of attempts to idealize $T(z, t)$, little attention has been given to $u(z, t)$ or, equivalently, vector shear $S(z, t)$. The lack of shear-focused studies reflects the challenge of making suitable near-surface measurements. Acoustic Doppler current profiler

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Fig. 1. Near-surface shear induced by a diurnal warm layer. An otherwise well mixed layer is forced by a weak wind of constant direction and magnitude. The bulk velocity oscillates at the Coriolis frequency (2-day period, 14°N). The addition of periodic penetrating solar radiation and heat loss induces near-surface stratification and, consequently, shear because the now-warmed layer traps the momentum input from wind. Contours and current arrows are derived from an idealized simulation (see section 3).
(ADCP) measurements in the near surface are challenging to obtain for a number of reasons including ship-wake contamination, the effects of surface gravity waves, the finite blanking distance of the ADCP, and bubbles associated with wave breaking (e.g., Marmorino et al. 1999; Gemmrich and Farmer 2004).

Overcoming the challenges of obtaining near-surface velocity measurements requires purposely designed platforms and, especially for the study of DWLs, a sufficiently high-vertical-resolution ADCP. Recent studies have attached ADCPs to either buoys or Lagrangian floats to obtain a continuous depiction throughout the day of shear across the DWL at 1-m vertical resolution (Prytherch et al. 2013; Wenegrat and McPhaden 2015; Sutherland et al. 2016; Shcherbina et al. 2019). A very high-resolution (0.03 m) depiction, albeit gappy in time, is also provided by Shcherbina et al. (2019). Richardson numbers (Ri) are calculated in some of these studies and are close to 0.25 in the sheared layer. Wenegrat and McPhaden (2015), however, do not capture the top 7 m, and Sutherland et al. (2016) call for additional, higher-resolution Ri estimates. Ultimately, we will build on these studies by providing corroborating evidence for a critical Richardson number, but we will also investigate lesser-studied aspects of DWL shear such as its inertial rotation and vertical distribution.

Ekman spirals are central to understanding upper ocean currents. Detecting them requires averaging over several inertial periods, yet interpreting this average during a fair weather day requires an understanding of how DWL shear and its inertial rotation evolve during the day (Price and Sundermeyer 1999). Accounting for inertial rotation may also improve corrections to subsurface-based estimates of SST. For a specific day, Prytherch et al. (2013) demonstrate a clear improvement of the temperature estimate by adding vector winds and Coriolis terms to the commonly used Fairall et al. (1996) correction. Otherwise, however, Prytherch et al. (2013) do not include any discussion of inertial turning from their dataset comprising 4700 days of measurements. Such rotation is also minimally considered by Shcherbina et al. (2019), who only note that the strong near-surface currents they observe at 10°N do not rotate significantly during their lifetime. Rotation is not applicable in Wenegrat and McPhaden’s (2015) study given their equatorial setting.

Arguably the most thorough, albeit partly qualitative, depiction of inertial turning in a DWL is provided by Sutherland et al. (2016) from observations in the subtropical North Atlantic (26°N). They demonstrate the turning by averaging the velocity over the top 5 m and calculating an anomaly relative to 25 m in order to show that the diurnal jet accelerates in the along-wind direction early in the morning and that a crosswind velocity component develops later with a lag consistent with the inertial period. They conjecture that the turning may explain, at least in part, the factor-of-3 discrepancy between their observed deepening rate of the DWL (1.8 m h⁻¹) compared to that in the equatorial Pacific (6 m h⁻¹; Smyth et al. 2013).

This paper focuses on the diurnal evolution of $S(z, t)$ in the DWL using observations from 13° to 18°N. Using velocity and temperature measurements from a new surface-following platform, we first demonstrate the quantitative link between the magnitudes of along- and crosswind components of shear. We then show that shear is concentrated at the base of the DWL, with a magnitude that is linked to the stratification through a critical Richardson number criterion. Idealized modeling, however, demonstrates that a critical Richardson number is not reached below wind speeds of 2 m s⁻¹, thus defining different regimes of diurnal SST anomalies at low and moderate wind speeds.

2. Observations from a surface-following platform

Data were obtained with SurfOtter (Fig. 2), a new surface-following platform designed for near-surface oceanographic measurements. Its body consists of a 3-m-long, 0.4-m-diameter aluminum tube attached to a 2-m-deep fin that is tethered to the ship with 200 m of cable. Near the body, the cable divides into a three-point bridle, much like a kite. This forces the body outboard so as to sample water unaffected by the ship.

This is the first study to present data obtained with SurfOtter, and we will therefore introduce all of its instrumentation. The focus of this paper, however, will be of a subset based on the vertical velocity shear and temperature-induced stratification.

Along the fin of SurfOtter (Fig. 2c), we placed five temperature loggers, four CTDs, four fast thermistors, and two GusTs (pitot tube, fast thermistor, pressure sensor, and accelerometer). Two temperature loggers were also placed near the nose where an ADCP is situated. In other deployments not discussed in this paper, a 6-m cable with additional instrumentation was attached to the bottom of the fin.

Turbulence inferred from the fast thermistors will be the focus of complementary studies. Unfortunately, we cannot easily combine the shear and turbulence analyses because the shallow velocity data are contaminated by acoustic reflections from the fin and the cable if present. In this paper, data within 3 m of the surface are contaminated. The ADCP will be pitched forward in future deployments to avoid the contamination.
The ADCP is a high-frequency (1000 kHz) Sentinel V, which sampled at 4 Hz and 0.5-m vertical resolution and had a maximum depth range of 15–20 m. Shear within the DWL is well portrayed with 15-min temporal averaging and 2-m depth averaging. We focus on shear as opposed to velocity because shear has the benefit of filtering out the bulk motion of the mixed layer, which is inconsequential for our purposes. In the schematic (Fig. 1), for example, the bulk motion is an inertial oscillation. This filtering could also be achieved by calculating a velocity anomaly relative to a depth below the warm layer. We will use shear, however, to avoid needing to define a reference velocity, which would be a significant complication given the limited ADCP range. Cursory velocity anomaly calculations indicate that they are 0.1 m s$^{-1}$, near the lower bound of values from past studies. For brevity, we also describe shear vectors as the quantities undergoing inertial rotation rather than velocity vectors.

Data were recorded on a slow transit northward (13°–18°N along 134.7°E at 1.5 m s$^{-1}$) during a segment of a 2-month-long field campaign in 2018 aboard the R/V Thomas G. Thompson in the Western Pacific. The project focuses on the Propagation of IntraSeasonal Tropical Oscillations (PISTON), which are 20–60-day time scale atmospheric oscillations in outgoing long-wave radiation, SST, and precipitation that are influenced by near-surface ocean properties.

Here we focus on four consecutive days (4–8 October 2018) with wind speeds of 1–5 m s$^{-1}$ and strong insolation with daily peaks of 1000 W m$^{-2}$. These conditions are conducive to establishing a DWL which, during these four days, formed above a 40–70-m-deep mixed layer. These four days represent a period of appreciable near-surface warming and during which we removed the cable from beneath the fin to enable velocity measurements starting at 3 m. The first day, with lower wind speed than the following three days, is excluded from some figures as the shear is presumed to occur in the top few meters where we were unable to observe it.

Temperature dominated the stratification for the specified period. It was measured directly in the top 2 m by the sensors on the fin. Below 10 m, temperature was recorded by the microstructure profiler Chameleon (Moum et al. 1995), which profiled every 6–12 min. In the top 10 m, Chameleon can measure the general temperature profile, but it underestimates temperature gradients because it measures water mixed and distorted by the ship. To more accurately estimate the temperature gradient over the top 10 m, we interpolate between SurfOtter and Chameleon values using either exponential or hyperbolic tangent profiles following Moulin et al. (2018).

An important factor in permitting a good measurement of DWL structure is the degree to which SurfOtter follows the free surface. To test this, we compare (i) the vertical displacement as determined by twice integrating an accelerometer oriented vertically to (ii) a separate pressure record that is converted to an equivalent depth. Before comparison, the nonhydrostatic component of pressure is removed from the pressure record with the method described in appendix A. An example (Fig. 3) time series dominated by 1-m amplitude, 11-s period
swell shows a root-mean-square displacement relative to the free surface of 0.05 m.

3. Idealized diurnal warm layer modeling

To supplement our observations, we created idealized simulations of DWLs. All simulations are one-dimensional (vertical) with solar radiation that penetrates the surface for 12 h centered about noon with a sinusoidal time dependence with a peak of 900 W m\(^{-2}\). Absorption of solar radiation is treated with a two-band profile (Jerlov type IA; Paulson and Simpson 1977). A constant surface heat loss that yields a net-zero daily heat input through the surface is chosen for convenience. The only other forcing is a constant wind stress. The influence of this wind is examined by running multiple simulations that differ only by the magnitude of the stress.

We use the MITgcm check point 67g (Marshall et al. 1997; Adcroft et al. 2004) at high resolution. Time steps are 10 s and grid cells are 0.1 m at the surface, increasing by 2% with each grid cell to a depth of 100 m. Overall, 150 depth cells are used, but much of this grid is surplus to requirements as the mixed layer depth is set to 40 m, comparable to the observed value.

Simulations are run for four days starting at midnight. Velocity, initialized as zero, displays an impulse response in the mixed layer, oscillating with the 2-day Coriolis period (a latitude of 14.4°N). A spin up period is unnecessary, however, because shear is independent of this oscillation and each day is identical for the intents and purposes of this paper.

Four distinct mixing schemes are tested, encompassing a range of approaches to simulating upper ocean mixing: (i) a bulk model modified to include shear-induced mixing (PWP; Price et al. 1986), (ii) a scheme with diffusivity and viscosity that depend on Richardson number (PP; Pacanowski and Philander 1981), (iii) a second-moment closure scheme with two diagnostic turbulent length scales (GGL; Gaspar et al. 1990), and (iv) a semiempirical scheme with features such as non-local fluxes and a smooth transition in diffusivity between the mixed layer and thermocline (KPP; Large et al. 1994). All mixing schemes use the default parameters specified in the MITgcm. The PWP model is not available as an option and is instead run as a stand-alone simulation with fixed grid spacing of 0.1 m. Vertical grid resolution for the KPP simulations is set to 0.5 m because a 0.1-m grid induces unphysical artifacts associated with non-local fluxes. This is consistent with Van Roekel et al. (2018) who find that diurnally forced simulations with a 1-m grid are more reliable than those with a 0.1-m grid.

We primarily use the Gaspar et al. (1990) mixing scheme based on (i) its favorable prediction of shear compared with observations, and (ii) its sophisticated approach to quantifying turbulence. Appendix B details our evaluation of the mixing schemes.

4. Inertial rotation of the sheared layer

In the Northern Hemisphere, we expect a rightward (clockwise) turning of DWL shear throughout the day. This is well demonstrated on 7 October (Fig. 4).
The along-wind shear (denoted $\partial u/\partial z$) is stronger earlier in the day and the crosswind component $\partial v/\partial z$ increases later. There is also a clear correlation between shear and stratification, which we will return to in section 5.

Hereafter, we develop this qualitative illustration of inertial turning into a quantitative demonstration of the link between the magnitudes of the two shear components.

Consider a DWL of variable depth $h(t)$ embedded within a much deeper, otherwise unstratified mixed layer. Let $t = 0$ when the near surface first becomes stratified [approximately 0800 local time (LT); Fig. 4c]. An exact definition of the depth $h$ is not needed, but we treat it as the depth of fluid accelerated by the wind. With a constant wind stress $\tau$ in the $x$ direction, the bulk momentum equations for the layer of depth $h$ are

\[
\frac{\partial (u h)}{\partial t} = \frac{\tau}{\rho} + f(u h), \tag{1}
\]

\[
\frac{\partial (u h)}{\partial t} = -f(u h), \tag{2}
\]

which, for an initially motionless layer, have the solution

\[
u h = \frac{\tau}{f \rho} \sin(\omega t), \tag{3}
\]

\[
v h = \frac{\tau}{f \rho} \left[ \cos(\omega t) - 1 \right]. \tag{4}
\]

This is a standard problem that usually requires an additional expression or differential equation for $h(t)$ to solve the full system of equations (e.g., Pollard et al. 1972; Ushijima and Yoshikawa 2019). Here we seek only the link between the velocity components assuming constant $\tau$:

\[
u = u \frac{\cos(\omega t) - 1}{\sin(\omega t)}. \tag{5}
\]

Using 12 h as an estimate of the duration of a DWL before it is mixed away, Fig. 5 shows that the along-wind component is always larger for latitudes below 15°. By comparison, for latitudes of 30° and 60°, the crosswind component exceeds the along-wind component after 6 and 3.5 h, respectively.

To express Eq. (5) in terms of shear, we assume there is a finite-thickness sheared layer over which the bulk surface layer's velocity decreases. Given the thickness will be the same for both components, shear components are related in the same way as the velocity components:

\[
\frac{\partial v}{\partial z} = \frac{\partial u}{\partial z} \frac{\cos(\omega t) - 1}{\sin(\omega t)}. \tag{6}
\]

We choose by eye the layer over which we vertically average the shear, a subjective but necessary approach.
The shear is noisy, which precludes using a simple threshold criterion. Similarly, we do not use a temperature gradient threshold because it would be sensitive in another way. As noted in section 2, temperature between 2 and 10 m is interpolated rather than directly measured. Even if it were possible to define the sheared layer in terms of a threshold, the definition would still be subjective in our choice of threshold. An example of our choice of the boundaries of the layer is shown in Fig. 4. Where appropriate in Fig. 6, we low-pass filter the shear in time with a 2-h window (denoted $h/C_1$). It is possible to test how well the observed shear components agree with the relation of Eq. (6) for three of the four days. The other day has negligible shear below 3 m as will be described in section 5. On 7 October, the crosswind shear agrees well with the prediction (Fig. 6). This day, which was the example used in Fig. 4, is the best test of our simple prediction because it is the day with the most consistent wind forcing, in both direction and magnitude, as required by Eq. (6).

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Consistent winds are also present on 5 October and again the observed magnitude of the crosswind shear agrees with that predicted. The noticeable discrepancy starting at 2100 LT occurs because we transit through the edge of an eddy, during which an ascending band of shear and stratification is evident both in the Chameleon record and the shear record from a lower-frequency current profiler (not shown).

On 6 October, the change in direction of the wind throughout the day is appreciable. The speed in the nominal crosswind direction is persistently and appreciably negative between 0800 and 1200 LT. Conversely, after 1600 LT, it is positive and nearly the same magnitude as the speed in nominal along-wind direction. Consequently, the magnitude of the predicted shear in the crosswind direction is underestimated prior to approximately 1600 LT and overestimated thereafter.

Inertial turning of the sheared layer is particularly clear in our idealized model (Fig. 7). That is, Eq. (6) works especially well at predicting the crosswind shear (Fig. 7h) because the wind stress in the crosswind direction is always zero. The data are also not noisy and we can objectively define an appropriate layer to average the shear over (dashed lines in Figs. 7e,f): for a given time, the dashed lines enclose the vertical region in which the temperature gradient is at least half as large as the maximum temperature gradient for that time.

5. Differing shear regimes at low and moderate wind speed

The colocation of shear and stratification in the 7 October example described earlier (Fig. 4) is a hint that the DWL may be in a state of marginal instability (Thorpe and Liu 2009). This can be confirmed by demonstrating that the Richardson number ($R_i = N^2/S^2$, where $S$ is shear and $N$ is buoyancy frequency) fluctuates about a critical value of 0.25. In this marginal state, there is a balance in which wind forcing works to increase shear and decrease $R_i$, but shear-induced mixing works to increase $R_i$. Here we present results in terms of the reduced squared shear ($S^2 - 4N^2$), a quantity more...
robust to noise than Ri. When $S^2 - 4N^2 > 0$, flow is unstable to shear instability. Ultimately, we will demonstrate that this occurs provided wind speeds are sufficiently large ($\approx 2\, \text{m s}^{-1}$).

For the three days when wind speeds are at least $3\, \text{m s}^{-1}$, the strongest shear is largely confined within regions in which the temperature gradient exceeds 0.04 K m$^{-1}$ (Figs. 8e,h,k). The correlation is most obvious on 5 October followed by 7 October, and last 6 October. Periods of weaker correlation arise when our fitted temperature profiles above 10 m (either exponential or hyperbolic tangent as noted in section 2) are less accurate. For example, at 1530 LT 6 October (Fig. 8h), Chameleon profiles show a local temperature peak at 3–5 m deep that cannot be captured by our fitting procedure.

The $S^2 - 4N^2$ values are close to zero within the layers of strong stratification (Figs. 8f,i,l). In these three panels, we might have expected large swaths of white indicating $S^2 - 4N^2 = 0$. Instead, however, there are both positive and negative values. There are two possible explanations for this. As described by Smyth et al. (2019), marginal instability does not lead to a steady Ri, but rather a value that fluctuates about critical as turbulence grows and decays. More likely, however, the fluctuations result from the limitations of estimating the exact temperature profile in the top 10 m and hence the exact depth and magnitude of $N^2$ peaks.

As a statistical distribution, we expect our estimates of $S^2 - 4N^2$ values within the regions of strong temperature gradient to be centered near but below zero. Although the values should fluctuate about zero as noted above, the distribution may also incorporate values where or when marginal instability is not active. Figure 9 shows histograms of $S^2 - 4N^2$ for all points at which $\partial T/\partial z$ is at least 0.04 K m$^{-1}$ (the inner contours shown in Fig. 8). The distributions for each of the three days appear broadly as expected.

Inferring marginal instability in a DWL is not a new finding. Kudryavtsev and Soloviev (1990), for example, estimate a bulk Ri across the diurnal thermocline of 0.3 ± 0.1. Further, it is an expected result given that
there exist numerous oceanic and atmospheric processes that straddle a 1:1 line in $S^2$, $4N^2$ parameter space [see list on page 3 of Smyth et al. (2019)]. DWLs are merely a process near the higher end of this line. This instability at the base of the DWL means that turbulence-enhanced mixing is critical to DWL evolution, and that numerical representation requires a parameterization of this mixing.

At low wind speeds, however, the DWL may not be in a state of marginal instability. In these cases, stratification dominates ($S^2 - 4N^2 < 0$) in the top few meters. Unfortunately, many studies including this one have trouble capturing shear in the top few meters. In particular, we did not measure near-surface shear on 4 October when winds were only 1–3 m s$^{-1}$, and when the mixing is insufficient to appreciably deepen the DWL until convective cooling begins around 1700 LT (Fig. 8a).

If there is a wind speed below which shear-induced mixing is negligible, this implies that diurnal SST anomalies should be approximately independent of wind speed in this low-wind speed regime. At zero wind speed, there is no shear. Conversely, at moderate wind speed, shear decreases with wind speed because the velocity anomaly of the DWL remains approximately constant, but the thickness of the sheared and stratified layer increases with wind speed due to increased mixing (Price et al. 1986). Using our idealized model, we look for the wind speed at which shear production tends to overcome buoyancy input in a bulk sense. To do this, we run 11 separate simulations with idealized forcing that

![Fig. 8. Relative magnitude of shear and stratification on four consecutive DWLs. (left) Stratification determined from an interpolation (section 2). Two contours ($\partial^2T/\partial z = 0.02$ and $0.04$ K m$^{-1}$) are repeated to aid comparison. Wind speeds noted are the 25th–75th percentile of the values over the 12 h shown. (center) Squared shear. (right) Reduced squared shear, which is positive when the Richardson number is less than 0.25.](image-url)
differ only by the wind speed and repeat this for each of the four mixing schemes we are testing.

At a wind speed of 2 m s\(^{-1}\) shear begins to overcome stratification. In the low-wind regime (<2 m s\(^{-1}\), Figs. 10a,b), the maximum shear occurs at 2 m, but \(S^2 - 4N^2\) is negative throughout the majority of the DWL. (Fig. 10 uses the default GGL mixing scheme; see appendix B for equivalent figures with alternate mixing schemes.) Maximum shear occurs at 2 m s\(^{-1}\) (Fig. 10c).

Across all mixing schemes, shear increases with wind speed for the 0.5, 1.0, and 1.5 m s\(^{-1}\) wind speed simulations. Maximum shear then tapers off to a peak at 1.7–2.0 m s\(^{-1}\) (Fig. 11a). The peak for the PP scheme, which is difficult to discern on the scale of Fig. 11a, occurs at 2.0 m s\(^{-1}\).

Over the 0–2 m s\(^{-1}\) range, the time at which maximum shear occurs decreases from 1800 to 1600 LT for the GGL and PP schemes (Fig. 11c), with 1800 LT being the onset of convective cooling and hence the time at which the shear starts being mixed away. For these two schemes, the change in maximum SST is negligible in the 0–2 m s\(^{-1}\) range (Fig. 11b).

A 2 m s\(^{-1}\) wind speed arises in other studies focused on diurnal SST anomalies. Webster et al. (1996) parameterize anomalies differently above and below 2 m s\(^{-1}\). Similarly, Lukas (1991) parameterizes diurnal SST anomalies with an inverse dependence on the squared wind speed, but only for wind speeds above 2.7 m s\(^{-1}\). Gentemann et al. (2009) demonstrate that temperature profiles lose their exponential decay with depth at wind speeds above 2 m s\(^{-1}\). Finally, Scanlon et al. (2013), to best tune their model to match observed SST, argue that 2 m s\(^{-1}\) separates regimes in which the effects of wave breaking need to be included or not in the near-surface turbulence coefficients. In contrast to Scanlon et al. (2013), we argue that the separation is predominantly explained by the changes in mixing induced by DWL shear, with wave breaking (or lack thereof) having a smaller influence. Webster et al. (1996) and Gentemann et al. (2009) also both mentioned the importance of mixing or a lack thereof.

A rough scaling helps suggest why 2 m s\(^{-1}\) arises as the wind speed corresponding to maximum shear. This will occur when the kinetic energy input into the DWL by wind first exceeds the potential energy input by solar radiation (see also Pollard et al. 1972). The kinetic energy in the DWL (in m\(^2\) s\(^{-2}\)) follows from Eqs. (3) and (4):

\[
\frac{u^2 + v^2}{2} = \frac{\tau}{h_0 \rho_a} [2 - 2 \cos(ft)].
\]

The wind stress is related to the wind speed \(U\) by

\[
\tau = C_d \rho_a U^2,
\]

where \(C_d\) is a drag coefficient (\(1.25 \times 10^{-3}\)) and \(\rho_a\) is the air density (1.2 kg m\(^{-3}\)).

The buoyancy input in the same units as Eq. (7) is

\[
J_b = \frac{g \alpha}{\rho_a c_p} J_q f,
\]

where \(J_b\) is a buoyancy flux (m\(^2\) s\(^{-3}\)), \(g\) is gravitational acceleration, \(\alpha\) is the thermal expansion coefficient of seawater (\(3 \times 10^{-4}\) K\(^{-1}\)), \(\rho_v\) and \(c_p\) are the density and specific heat capacity of seawater, and \(J_q\) is the heat flux through the air–sea interface minus the radiative flux that penetrates deeper than \(h\). Although \(J_q\) is sinusoidal in time, for simplicity we will treat it as a constant by using an appropriate time average described later.

Equating the right-hand sides of Eqs. (7) and (9) and solving for the wind speed gives \(U_{cr}\), the critical wind speed at which the kinetic energy from wind equals the potential energy input by surface heating:

\[
U_{cr} = \left[\frac{2g \alpha f^2}{\rho_a c_p^2} \frac{h^2 J_q f}{2 - 2 \cos(ft)}\right]^{1/4}.
\]

At \(U < U_{cr}\), the kinetic energy input to the DWL by the wind is smaller than the potential energy input by daytime heating. Consequently, mixing should be minimal. For \(U > U_{cr}\), the potential energy is insufficient to stabilize the DWL against the increased kinetic energy and shear-induced mixing ensues.

The second fraction in Eq. (10) includes all components for which an estimate is required. We set \(t = 4.75\) hours, half of the 9.5-h period during which the net heat flux is into the ocean [equivalent to \(P_Q\) defined by...
Price et al. (1986); $h = 2\text{ m}$, the typical depth of the maximum temperature and velocity gradients at low wind speeds (see, e.g., Fig. 10); and $J_q = 185\text{ W m}^{-2}$, the time-averaged net heat flux convergence (absorbed radiation − surface cooling) in the top 2 m during the heating period. The heat flux $J_q$ is lower than may be inferred from Fig. 7g because $J_q$ excludes radiation that

FIG. 10. Shear exceeds stratification above a wind speed of $2\text{ m s}^{-1}$. Except for the varying wind speed, the five simulations shown are identical: 14.4°N latitude; diurnally varying, penetrating solar radiation that peaks at 900 W m$^{-2}$; and the Gaspar et al. (1990) mixing scheme is used.

FIG. 11. Shear within the DWL evolves differently in low- and moderate-wind speed regimes, with the separation at approximately $2\text{ m s}^{-1}$. The maximum shear differs between the four mixing schemes, but the wind speed at which this occurs is largely independent of mixing scheme. At wind speeds above $2\text{ m s}^{-1}$, the maximum SST during the afternoon is primarily a function of shear-induced mixing. Evolution of the shear throughout the day is shown for certain wind speeds in Figs. 10 and B2.
penetrates deeper than 2 m. With these choices, \( U_{cr} = 2.4 \text{ m s}^{-1} \), which is sufficiently close to 2.0 m s\(^{-1}\) given the simplicity of our scaling.

Large changes in our choices for \( J_q, t, f, \) or \( h \) still give \( U_{cr} \approx 2 \text{ m s}^{-1} \). The 1/4 exponent in Eq. (10) makes \( U_{cr} \) robust and arises because the buoyancy input into the layer of depth \( h \) increases linearly with \( J_q \) [Eq. (9)], whereas its kinetic energy increases as \( U^4 \) [Eqs. (7) and (8)]. For example, increasing \( J_q \) to 300 W m\(^{-2}\) or decreasing it to 100 W m\(^{-2}\) encompasses a large range of physically plausible values, yet gives \( U_{cr} = 2.6 \) and 2.0 m s\(^{-1}\), respectively. Similarly, when keeping other parameters constant, \( U_{cr} \) changes by only 0.3 m s\(^{-1}\) for latitudes between 1° and 60° and changes by only 0.5 m s\(^{-1}\) for choices of \( t \) between 3 and 7 h.

A detailed examination of the sensitivity of Eq. (10) in Fig. 12 shows that within a wide range of plausible input values, \( U_{cr} \) is within \( \pm 30\% \) of 2.0 m s\(^{-1}\). Given this limited sensitivity, while also recognizing that variability exists, we consider the 2 m s\(^{-1}\) threshold as a convenient rule of thumb, especially for the tropics, where day length and peak insolation vary minimally throughout the year.

With increasing wind speed beyond 2 m s\(^{-1}\), \( S^2 - 4N^2 \) is zero or slightly positive throughout much of the DWL (Figs. 10d,e). The maximum shear and maximum SST, and the associated times of day of these maxima, all generally decrease (Fig. 11), with the strongest dependence on wind speed being the 2–4 m s\(^{-1}\) range. The time of maximum SST for the PWP scheme varies widely because of abrupt changes in the depth of the mixed layer in the top few meters (see, e.g., Figs. B2d,e).

Maximum SSTs arising earlier in the day with increasing wind speeds have also been observed by Kondo et al. (1979). Similarly, Large and Caron (2015) predict that the timing of the peak afternoon SST depends on diurnal shear. We therefore conclude that the diurnal evolution of shear and its consequent mixing, or lack thereof, predominantly explains the different regimes of warming at low and moderate wind speeds. It also points to a need for shear measurements much closer to the surface to confirm the predicted change in behavior in low-wind conditions.

6. Conclusions

Sea surface temperature increases of only a few tenths of a degree are sufficient to appreciably alter lateral transport in the near-surface ocean. Horizontal gradients between horizontally adjacent diurnal warm layers of differing magnitude can play a dynamic role by introducing convergence and divergence in the surface ocean. This goes beyond their oft-cited thermodynamic role in heat fluxes through the air–sea interface.
Here we have observationally demonstrated key aspects of this modified lateral transport, expressed in terms of shear: (i) the link between the two components in a frame of reference aligned with the wind, and (ii) the existence of marginal instability, or lack thereof at low wind speed.

It is well understood that near-surface ocean properties depend on wind speed and insolation. DWL shear evolution is central to explaining these dependencies. Notably, it explains why several previous studies identify a wind speed at, or near, 2 m s\(^{-1}\) about which diurnal sea surface temperature anomalies behave differently with respect to wind speed. Although the magnitude of the anomaly is most important, its timing also appreciably differs with wind speed, with the peak varying by one hour depending on the extent to which shear-driven mixing causes the DWL to descend.

Improved understanding DWL shear will feed back into operational procedures like DWL corrections for subsurface temperature measurements and idealizations of DWL velocity profiles needed for climate models. This will be aided by additional near-surface shear and complementary turbulence observations. Nevertheless, our combined observational and simulated results presently indicate that at low wind speeds (<2 m s\(^{-1}\)) velocities are confined to the top few meters and temperature profiles will depend primarily on water clarity and insolation rather than wind. Above this wind speed, the presence of marginal instability implies that shear exhibits a profile equivalent to stratification, one that starts off surface-confined in the morning and progresses to one with a subsurface maximum later in the day.

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**APPENDIX A**

**Interpreting Pressure Records from SurfOtter**

To infer SurfOtter’s ability to follow the surface in section 2, we removed the nonhydrostatic component of the pressure record. Here we briefly show how this is achieved using linear wave theory and standard notation. For a monochromatic wave \(\eta = A \cos(kx - \omega t)\), the pressure at a fixed depth \(z\) (see, e.g., Chen 2005) is

\[
p(z, t) = -\rho g z + \rho g A e^{k z} \cos(kx - \omega t). \tag{A1}\]

For a spectrum of waves, the vertical decay of the waves as a function of wavenumber must be accounted for.
The conversion between wavenumber and frequency for deep water waves is

\[ k = \frac{4\pi^2f^2}{g}. \]  \hspace{1cm} (A2)

If \( \hat{\eta}(f) \) is the Fourier transform of \( \eta(t) \) then

\[ p(z, t) = -\rho g z + \rho g F^{-1} \left[ \hat{\eta} \exp \left( \frac{4\pi^2f^2}{g} z \right) \right], \]  \hspace{1cm} (A3)

where \( F^{-1} \) is the inverse Fourier transform. For interpreting pressure records from SurfOtter, we evaluate the record at \( z = \eta(t) - d \), where \( d \) is the nominal depth of the pressure sensor below the sea surface and \( \eta(t) \) is calculated by twice integrating a vertical accelerometer within the same instrument:

\[ p(z = \eta - d) = -\rho g (\eta - d) + \rho g F^{-1} \left[ \hat{\eta} \exp \left( -\frac{4\pi^2f^2}{g} d \right) \right]. \]  \hspace{1cm} (A4)

Although \( d \) and \( \eta \) have comparable magnitudes, we let \( \eta - d \approx -d \) in the exponential. This makes the inverse Fourier transform simple because it avoids mixing the time-dependent \( \eta(t) \) and frequency-dependent \( \hat{\eta}(f) \) terms within the argument of the transform. It also ensures the argument of the exponential is always negative, thereby avoiding numerical issues. Indeed, exponential decay is only an approximation to the true wave dynamics when \( z \approx 0 \), but a more accurate treatment (e.g., Smit et al. 2017) is beyond the scope of this study.

Our simple method is sufficient, at least near the bottom of the fin, given the clear agreement between the recorded pressure and that derived from Eq. (A4) (Fig. A1). The difference between the two quantities was defined as the effective displacement shown in Fig. 3.

**APPENDIX B**

**Comparison of Model Mixing Schemes**

The choice of mixing scheme used in our idealized model leads to markedly different predictions of the vertical profile of shear throughout the day (Fig. B1). Comparing these predictions against observed profiles guides our decision on which scheme or schemes are appropriate for simulating DWLs.

For all schemes, simulations are forced with 5 m s\(^{-1}\) winds and compared with vertical profiles of shear from four times of day from the average of three consecutive days when the wind consistently blew at 4–5 m s\(^{-1}\) and daily maximum sea surface temperature anomalies were relatively constant.

**Fig. B1.** Evaluation of predicted DWL shear using commonly used turbulence schemes. For each time of day shown, the observed curve is the average of all data from 2-h windows [e.g., 0900–1100 LT for (a)] from three consecutive days (5–7 Oct) when winds were consistently 4–5 m s\(^{-1}\). Profiles from simulations are single time slices.
were 0.4°C. For this wind speed, the maximum predicted shear for all four times is reasonably independent of the mixing scheme. This is not true at different wind speeds (Fig. 11a).

There is not one scheme that clearly correlates best with our limited observations. Nevertheless, for this study we consider the second-moment closure scheme (GGL) as the default scheme from the choices here.

Fig. B2. Reduced squared shear as predicted by three different mixing schemes: PWP, PP, and KPP. Each column is equivalent to Fig. 10 except for the mixing scheme.
Wade et al. (2011) also invoked this scheme when modeling DWLs in the equatorial Atlantic. Both this scheme and PP agree with the general shape of the observed profile at all times of day. (The double peaks in the observations at 1000 and 1600 LT should be treated with caution.) A key difference between these two schemes is their predictions of the depth of maximum shear. GGL better predicts this depth compared to observations at 1600 LT, whereas PP better predicts it at 1900 LT.

Given our observation that $S^2 - 4N^2 \approx 0$ (Fig. 8), which is equivalent to a critical Richardson number of 0.25, we initially expected the PP scheme to work well because of its simple Richardson number dependence. However, PP was developed back when vertical grid spacing for a three-dimensional numerical model was typically 10–100 m at best. It therefore needed to account for underresolved shear, but this makes it overly viscous for our purposes (note the long tails below the maximum shear). In this scheme, diffusivity and viscosity start increasing above a background value well before Ri reaches critical.

Because of this viscosity increase in the PP scheme, $S^2 - 4N^2$ is negative everywhere for wind speeds of 1–3 m s$^{-1}$ (Figs. B2f–j). The same occurs for the KPP scheme (Figs. B2k–j) because it includes a similar Ri dependence (see Fig. 3 of Large et al. 1994) for interior mixing. Although KPP includes a sophisticated boundary layer component, this is negligible during the day because the strongly stabilizing buoyancy flux leads to a boundary layer thickness of a fraction of a meter. Unlike PP and KPP, the PWP scheme predicts $S^2 - 4N^2$ fields (Figs. B2a–e) similar to that predicted by GGL, which were shown in Fig. 10. PWP explicitly includes a critical Ri = 0.25 constraint, so it is expected that $S^2 - 4N^2 = 0$ at sufficiently high wind speeds.

Shear observations at lower wind speed will be particularly helpful in differentiating the skill of different mixing schemes. Note, for example, the large differences between mixing schemes in the maximum shear predicted at 2 m s$^{-1}$ wind speed (Fig. 11a). Further observations may also enable tests against recent improvements to near-surface mixing parameterizations (e.g., Kantha and Clayson 2004; Esters et al. 2018; Van Rockel et al. 2018). We currently have too few degrees of freedom to undertake such tests.

REFERENCES


