Diurnal Cycles of Near-Surface Currents Across the Tropical Pacific

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Abstract Solar radiative heat and wind-forced momentum can be trapped at the surface and transmitted into the ocean interior via a daily descending shear layer that mixes heat and momentum downwards. Here, we characterize the extent of this mechanism across the tropical Pacific via seven mooring deployments spanning the eastern and western tropical Pacific and the Inter-Tropical and South Pacific Convergence Zones. We find a wide range of diurnal cycles in temperature and velocity, from very weak cycles trapped in the upper 10 m to strong cycles that reach as deep as 60 m. These deeper diurnal cycles appear where the strong background shear above the equatorial undercurrent in the eastern equatorial Pacific helps destabilize the fluid, setting up a persistent state of marginal instability. At the sites located in these marginally unstable regions, we find a linear relationship between wind speed and the depth and strength of the diurnal cycle in velocity.

Plain Language Summary The daily cycle of daytime heating and nighttime cooling over the tropical ocean can inject heat and wind momentum from the atmosphere into the ocean. Daytime warm layers at the ocean surface can trap wind momentum, and the heat and momentum in these layers can be mixed downward each day. We explore this process at seven different locations across the tropical Pacific, finding that this near-surface daily heat and momentum mixes far deeper into the ocean near the equator than away from it. We find that the deeper mixing occurs only where strong near-surface currents make the fluid more “mixable,” so that the wind-driven jet can mix downwards more easily. We also find that stronger wind is associated with a stronger, deeper jet in these regions.

1. Introduction

The diurnal cycle of solar heating modulates the penetration of heat and momentum into the ocean interior. Daytime heating creates warm stratified layers in the upper few meters at the ocean surface, called the diurnal warm layer (DWL). Nighttime cooling destroys the DWL and transmits the heat in these warm layers downward via convective mixing. When wind blows over the warm afternoon ocean, wind momentum can be trapped and concentrated in this buoyantly isolated warm layer. Especially in the tropics, where Coriolis turning is weak, this layer can thus form a diurnal jet: a localized wind-driven current that forms at the ocean surface each afternoon.

The diurnal warm layer in the top few meters of the afternoon ocean can be several degrees warmer than the bulk SST (Ward, 2006); thus the DWL plays an important, widespread role in setting heat flux between the ocean and the atmosphere (Bellenger & Duvel, 2009; Fairall et al., 1996; Stuart-Menteth et al., 2003; Woolnough et al., 2007). The diurnal jet complicates the role of the DWL in modulating heat fluxes, since shear at the base of the jet can induce vertical mixing that tends to lower the temperature at the sea surface (Liu et al., 2016; Sutherland et al., 2016). Diurnal jets also complicate momentum fluxes between the atmosphere and the ocean, which depend upon the relative motion of the air to the ocean surface. Typically, near-surface ocean currents are measured at or below 10 m, which often lies below the DWL. Since the surface of the ocean responds directly to wind forcing while the deeper geostrophic currents are also influenced by remote wind forcing, a better understanding of the DWL influence on surface currents is needed for accurate calculation of air-sea momentum fluxes (Kudryavtsev & Soloviev, 1990; Wijesekera et al., 2020). Thus the dynamics of the diurnal warm layer can influence climate model solutions, prompting efforts to parameterize its effects on temperature and velocity profiles in the ocean surface and near-surface (e.g.,...
Most of these studies have explored the dynamics of diurnal jets in the extratropics, where they rarely reach deeper than 10–15 m, largely due to the turning with depth associated with Coriolis rotational effects (Hughes et al., 2020; Price et al., 1986). Near the equator, a combination of near-zero Coriolis and specific subsurface conditions enables the diurnal jet to reach far deeper via a descending shear layer. This deep diurnal jet has been well-observed at 0°, 140°W in the eastern equatorial Pacific, where very high background shears uniquely support the descending shear layer mechanism (Lien et al., 1995). The shallow transition between the westward wind-driven South Equatorial Current (SEC) at the surface and the eastward Equatorial Undercurrent (EUC) below sets up a flow regime of “marginal instability,” described by a Richardson number $R_i$ that lies near a critical value near 1/4 at which perturbations can grow through shear instability. The diurnal warm layer interacts uniquely with this type of near-surface flow regime: afternoon heating stabilizes the fluid enough to confine wind momentum in the surface layer, while the resulting shear destabilizes the fluid enough to induce mixing that extends the wind momentum deeper into the water column. This downward mixing of heat and momentum forms a descending shear layer that can extend the reach of the diurnal jet downwards, carry wind-driven momentum toward the ocean interior (e.g., Danabasoglu et al., 2006; Lien et al., 1995; Moulin et al., 2018; Moum et al., 1989; Pham et al., 2017; Smyth & Moum, 2013).

Observing the diurnal jet presents a challenge, since interference from waves, bubbles, fish, ships, buoy structures, and from reflections from the moving ocean surface itself can all impede the collection and interpretation of velocity data in the upper ocean. Some studies have employed ADCPs mounted on autonomous instruments (Hughes et al., 2020; Shcherbina et al., 2019; Sutherland et al., 2016), as well as higher-resolution microstructure profiler instruments (Lien et al., 1995; Sutherland et al., 2016; Wenegrat & McPhaden, 2015). A handful of studies have employed near-surface deployments of ADCPs or multiple acoustic current meters on moorings (Cronin & Kessler, 2009; Prytherch et al., 2013; Smyth et al., 2013; Wenegrat et al., 2014). These mooring studies have the advantage of longer observational periods, but many have suffered from noisier velocity observations, prompting experiments combining moored velocity measurements with collocated shorter-term microstructure observations.

These relatively long mooring records can nonetheless help to identify and explore lower-frequency modulations of the diurnal jet, and to expand the range of atmospheric and dynamical regimes seen during a specific experiment. Smyth and Moum (2013) used 5 years of observations to show that downward mixing on the equator can be suppressed in boreal spring, a finding confirmed via modeling by Pham et al. (2017). Wenegrat and McPhaden (2015) used an 8-month record to identify a similar seasonality in the Atlantic, finding that steady trade winds and a destabilized near-surface flow made for a stronger, deeper equatorial diurnal jet in the boreal summer and fall. Finally, a year-long record of velocity observations at 20°S, 85°W was used by Prytherch et al. (2013) to test the one-dimensional model of the diurnal jet given by Price et al. (1986) and refined by Fairall et al. (1996). They found reasonable agreement with this model in the upper 5–15 m of the ocean, with a 0.77 correlation between the observations and the model prediction of shear.

The present study deployed ADCPs on surface moorings across the tropical Pacific as part of a Tropical Pacific Observing System (TPOS) pilot project meant to explore the possibility of long-term monitoring in the ocean near-surface boundary layer. The pilot stations were operational Tropical Atmosphere Ocean (TAO) surface moorings that spanned the eastern and western equatorial Pacific and the Inter-Tropical and South Pacific Convergence Zones. Combining these shallow (7–65-m depth) velocity profiles with concurrent wind and radiation observations, we characterize the sub-7 m differences among diurnal jets across these various regimes, and identify a linear relationship between wind speed and the strength and depth of the eastern equatorial jet.

2. Data and Methods

2.1. Data Processing

The near-surface observations were sampled via instrument enhancements to standard operational National Data Buoy Center (NDBC) TAO moorings at 8°N, 165°E; 2°S, 165°E; 8°S, 180°; 0°, 180°; 0°, 155°W; 2°S,
The success of the upward-looking ADCP instruments was somewhat surprising given concerns about fish contamination of the acoustic signal in the near-surface tropical Pacific region (Plimpton et al., 1997). At 2°S, 165°E, 0°, 155°W and 0°, 180°, noise in the upper 40 m in each record might have indicated fish contamination, except that the expected corresponding increases in the ADCP beam amplitude were not observed. Two criteria were used to flag and reject between 12% and one-third of velocity data at each site: when the instruments tilted more than 20°, in accordance with the manufacturer’s specifications; and when the hourly difference between the ADCP and PCM were observed. A denser deployment of temperature sensors (circles) includes conductivity observations at instruments between the surface and 20 m. A point current meter (ovals) mounted between 12- and 24-m depth at each site is used to validate the ADCP observations. At 8°S, 180°; 8°N, 165°E; 0°, 180°; and 2°N, 125°W point current meters were deployed at 7-m depth as well. Meteorological enhancements (triangle) include short wave and long wave radiation and rain observations. Diamonds show standard TAO temperature sensors. (b) Map of mooring enhancements. (c) Duration of mooring enhancements. Gray lines show the duration of the deployment, colors show the extent of the near-surface velocity data that overlaps with wind, temperature, and salinity observations and was usable for this analysis.
power. Instrument noise, interference from surface waves, poorly sampled internal waves, or episodic mixing events likely all contribute to a relatively noisy signal from this sampling scheme. This noise complicated the identification of a diurnal signal in the hourly velocity record (Figures 2d and 2e), requiring the use of the composites and averaging in the analyses described below. The shallow limit of the upward-looking ADCP profile is set by sidelobe interference from the surface (Appell et al., 1991), and thus depends on how shallow the ADCP instrument was mounted; for these sites, the shallowest usable data occurred around 11-m depth. Velocity data were leveled using pressure measured at the ADCP instrument and interpolated to a fixed 2-m vertical grid.

Finally, substantial dropouts in wind observations occurred at 0°, 180° and 0°, 155°W. In these cases, we patched in daily reanalysis data from the CCMP version 2 wind vector product (Wentz et al., 2015) totaling 312 days out of 361 days at 0°, 180° and 257 days out of 342 days at 0°, 155°W. For the analyses below, we employ daily averages of the 10-min wind observations at each site, so the substitution does not introduce a sampling bias, especially since none of the stations exhibit a significant diurnal cycle of wind. The scatterometer’s regional average likely shows less temporal detail than the mooring observations, and may introduce any bias associated with the reanalysis product.

2.2. Estimating $R_i$

To evaluate each site’s sensitivity to vertical mixing, we compare buoyancy and shear squared by estimating the Richardson number

$$R_i = \frac{N^2}{\left|\frac{\partial u}{\partial z}\right|^2},$$

(1)

where $N^2 = \frac{\partial b}{\partial z}$ is the buoyancy frequency, $b = -g(\rho - \rho_0)/\rho_0$ is buoyancy, $g$ is gravity, $\rho$ is seawater density, and $\rho_0$ is the background density. $|u_z|^2 = (u_x)^2 + (v_y)^2$ gives the squared vertical shear, where $u = (u_x, v_y)$ is horizontal velocity.

This formulation of $R_i$ can be sensitive to values of shear squared near zero, so we follow Wenegrat and McPhaden (2015) and others by employing a reduced shear squared $M = |u_z|^2 - 4N^2$, which is equal to zero.
at the critical Richardson number $Ricr = 1/4$, indicating when the fluid is unstable to vertical shear. Noisy estimates of shear squared have led us to modify this reduced formulation further, with a looser criterion of

$$M = |u_z|^2 - N^2,$$

which is equal to zero at a critical Richardson number of $Ricr = 1$, at which the fluid can be susceptible to shear instability if perturbed. A $Ricr$ greater than 0.25 is not without precedent, as some studies (Fairall et al., 1996; Price et al., 1986; Prytherch et al., 2013) have empirically identified a critical bulk Richardson number of 0.65 for the descent of the shear layer.

2.3. Estimating $N^2$

Estimating the buoyancy gradient $N^2$ requires observations of both temperature and salinity in the near-surface layer. At each site, salinity observations extended from the surface at 1–20 m, with 5-m resolution. Below 20 m where salinity data are not available, we assume a constant salinity equal to that of the 20-m estimate. This method loses the nuance in diurnal stratification below 20 m that may be due to rainfall and other salinity events, but still illuminates the temperature-driven differences in buoyancy among the sites, as well as the dominant diurnal pattern in stratification at each site.

2.4. Estimating Shear Squared

Taking the vertical gradient and squaring an already noisy velocity signal presented a challenge for calculation of the shear squared necessary to estimate the Richardson number. We found that the velocity signal from the ADCP is too noisy for direct calculation of a diurnal composite of shear squared, especially in the upper 40 m at 2°S, 165°E, 0°, 155°W and 0°, 180°. To estimate the shear squared associated with the diurnal signal, we thus take a different approach. We identify the component of the velocity data associated with the diurnal signal by first calculating the diurnal composite of the velocity observations, and then estimating the squared shear from this diurnal composite. This may result in a depiction of diurnal shear squared that is lower in amplitude than the true signal, but still allows us to identify the pattern of the signal over the diurnal time scale.

3. Results

3.1. Diurnal Cycles in Temperature and Velocity

Daily diurnal cycles can be identified clearly in the observations of shortwave radiation (Figure 2b) and ocean temperature (Figure 2c). The diurnal cycle in wind at each site is weak, at least an order of magnitude smaller than the overall wind signal magnitude (Figure 2a). Relatively noisy ADCP data and an observational window that starts below 10-m depth means that diurnal composites are necessary to identify the surface-forced diurnal cycle in velocity. A record-mean diurnal composite loses the detail of the temporal variation at each station, but nonetheless allows us to make some overall comparisons among the sites.

The diurnal cycle in near-surface temperature is fundamentally different between the western Pacific sites (color contours, left column, Figure 4) and the eastern equatorial sites (color contours, right column, Figure 4). At all seven sites, we see surface warming that starts in the early morning and continues until midafternoon. Nighttime cooling corresponds with lower SST and convective overturning within the surface ocean mixed layer, until daytime surface heating begins anew. This surface heating and cooling clearly extends downwards over the course of the day at all five sites, but reaches more deeply at the eastern equatorial sites, with a +0.05 °C temperature anomaly reaching 10–15-m depth at the western and off-equatorial sites and 20–30-m depth at the eastern equatorial sites (Figure 4).

A diurnal cycle as large as 10 cm/s can be seen in the velocity at most of the stations as well (Figure 4, line contours). For simplicity, zonal and meridional velocities are translated into along-wind velocities with respect to the record-mean direction of the winds (arrows, Figure 3d) which are largely toward the west and exhibit very little diurnal cycle at each site. To demonstrate the full 24-h structure of the diurnal jet, we show
the velocity composites with respect to the record-mean velocity profile, rather than assuming a “blank slate” profile in the early morning. Velocity anomalies are calculated with respect to the bottom-most observational bin to remove the influence of the barotropic tide. Thus the along-wind jet appears as a positive anomaly (purple contours, Figure 4), and the absence of the jet appears as a negative anomaly (orange contours, Figure 4). During daytime heating and stratification (filled red contours, Figure 4), velocities in the uppermost layers at some of the equatorial sites tend to strengthen and to turn toward the direction of the wind (purple contours, Figure 4).

The difference between the western and eastern equatorial sites occurs in the diurnal cycles of velocity as well. At the western sites, the diurnal change in velocity is shallow, perhaps not even entering the observational window at 8°S, 180°. At the eastern equatorial sites, the diurnal change in velocity is strong at 10-m depth at all four sites, on the order of a 5–10 cm/s night to day difference. The depth penetration of this diurnal jet differs among the equatorial sites and is deepest in the east, from about 30-m depth at 0°, 180° to 50–60-m depth at 2°S, 140°W and 2°N, 125°W.

At the shallowest observed depths, the daily maximum in velocity tends to occur slightly later—between 2 and 6 hours, depending on the site—than the daily maximum in near-surface temperature. We interpret this lag to indicate the time necessary for surface heating and wind stress momentum to reach the observational window set by the ADCP. At the top of the observation window, the windward jet reaches its maximum in the late evening, propagating downwards, and weakening through the night until it is mixed away. At the eastern equatorial sites, the jet reaches as deep as 60 m and lasts into the early morning, when nighttime convective mixing has cooled the water that lies above. Note that since we calculate the jet with respect to the bottom-most observed velocity, its values reduce to zero at the base of the ADCP sampling, but certainly could extend deeper into the water column than shown here.

3.2. Marginal Instability at the Eastern Equatorial Sites

Several factors influence the depth and speed of the diurnal jet, including latitude, wind strength, insolation, and the “mixability” of the near-surface ocean. This last factor, set by the balance between stratification and background shear that is expressed by $R_i$, likely plays the largest role in setting the overall difference between the western and eastern equatorial sites. Sun et al. (1998), Moum et al. (2009), Inoue et al. (2012), Smyth et al. (2013), Smyth and Moum (2013), and others have shown that the shallow
SEC-to-EUC transition at 0°, 140°W represents a specific regime of marginally unstable flow, where the background Richardson number varies about a critical value of 1/4. In this marginally unstable state, small perturbations to the flow can instigate mixing, making the eastern equatorial region particularly susceptible to surface-forced shear instability compared to the rest of the near-surface ocean. Cronin and Kessler (2009) showed that the mean state of the near-surface flow varies near $R_i = 1/4$ at 2°N, 140°W as well. The depth of the record-mean diurnal jet at the eastern equatorial sites suggests a similarly marginally unstable flow regime at all four eastern equatorial sites, as far west as 0°, 180° and as far east as 2°N, 125°W. 

This broad zonal extent makes sense given the unique characteristics of the central and eastern equatorial region: a state of marginal instability can be considered an example of a “self organized criticality” that can occur in a forced, stratified shear flow (Smyth et al., 2019). In the ocean, situations that exhibit this unique balance of background shear and stratification include some estuarine flows, certain types of river plumes, and the shallow reversing current structure of the eastern equatorial Pacific and Atlantic (see review in Figure 4. Diurnal composites of along-wind velocity anomaly relative to the bottom-most observational bin (contours; purple-orange colorbar) and temperature anomaly (color shading; red-blue colorbar). Western sites are shown in the left-hand column, central, and eastern equatorial sites in the right-hand column. Anomalies are taken with respect to the time-mean along-wind velocity and the time-mean temperature profiles. Along-wind velocity is calculated with respect to the direction of the record-mean wind, shown in Figure 1. Diurnal composites are smoothed via a 3-hour triangle filter and repeated for clarity.
Smyth et al. (2019). We might expect a state of marginal instability wherever this type of flow structure dominates in the Pacific. At all four eastern equatorial sites, the transition from the surface westward-flowing SEC to the shallow underlying eastward EUC creates a strong background near-surface shear that is largely not present at the western or off-equatorial sites (Figure 3b). For the eastern equatorial sites, the zero-crossing of the zonal velocity profiles shows the mean depth of transition from westward SEC to eastward EUC. The mean transition between the SEC and the EUC appears within our observational window at about 40-m depth at 0°, 155°W and 2°S, 140°W (Figure 3b), but strong background shear also dominates the flow at 0°, 180° and 2°N, 125°W, where a deeper EUC lies below the sampling of the ADCPs deployed here (Johnson et al., 2001).

The 2°S, 165°E site provides an interesting contrast. A strong vertical shear exists in the mean zonal current due to a weak eastward current overlying a deeper westward current. Trade wind forcing here acts to reduce the vertical shear (Figure 3b), making the water column less susceptible to mixing, perhaps explaining why a mean deep penetration of the diurnal cycle is not observed at this site. Further work is needed to explore how the diurnal cycle might vary during transient westerly wind events acting on these types of reversing current structures (Cronin et al., 2000).

3.3. Diurnal Evolution of \( \text{Ri} \)

The background state of marginal instability helps to explain why the diurnal layer extends more deeply at the eastern equatorial sites. A comparison of the diurnal evolution of \( \text{Ri} \) at each of these sites partly illuminates the process by which these deep shear layers descend. At all seven sites, the temperature-driven diurnal cycle in \( N^2 \) is clear, with an increase in stratification beginning around 6 a.m. at the surface and peaking in the midafternoon. For the shear squared component of the \( \text{Ri} \) calculation, we estimate the diurnal velocity composite as described in Section 2.4, noting that this operation may result in a lower amplitude for the diurnal shear squared than might have resulted from a direct diurnal composite of the hourly shear squared signal.

To evaluate the balance between \( N^2 \) and shear squared, we calculate a modified version of reduced shear squared that reaches zero when \( \text{Ri} = 1 \) (Equation 3). This diagnostic \( M \) helps to indicate when buoyancy or shear dominates in the fluid. At several of the sites, we can see that shear appears to dominate in the upper 15 m or so of the ocean (red, Figure 5), despite strong afternoon stratification (Figure 4). At the eastern equatorial sites at 0°, 155°W, 2°S, 140°W, and 2°N, 125°W, we observe a pattern of descending shear dominance, where shear appears to dominate near at the top of the near-surface observational window starting around noon and descends over time. This signature of the descending shear layer closely aligns with the baroclinic acceleration of the diurnal jet (Figure 4), suggesting as in other studies that shear instabilities provide the mechanism for transferring the diurnal jet downwards.

At 0°, 180°, a similar pattern may be present—a slightly lighter blue can be observed beneath the shallow daytime red layer in Figure 5d—but has proven more difficult to identify from the data. At the off-equatorial sites at 8°S and 8°N, weak stratification and very weak shear squared combine such that the diurnal warm layer does not appear to penetrate into the observational window in the record-mean analysis. At 2°S, 165°E, very low stratification means that shear dominates everywhere, but no coherent descending diurnal pattern can be observed.

3.4. Eastern Equatorial Diurnal Cycle and Wind Strength

Equatorial observations in the Atlantic (Wenegrat & McPhaden, 2015) and the Pacific (Smyth & Moum, 2013) suggest that a seasonal combination of factors, including the strength and steadiness of the trade winds, can cause a weaker, shallower jet in boreal spring and a stronger, deeper jet in boreal summer, winter, and fall. In the Pacific, Smyth and Moum (2013) and Pham et al. (2017) connect this effect to the seasonality of the deep-cycle turbulence that can be triggered by the diurnal jet, finding that shoaling of both the EUC and the thermocline can suppress vertical mixing in boreal spring, though a springtime slackening of the trade winds may play a role as well.
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Here, we focus on the wind piece of the puzzle by comparing the diurnal jet across a range of wind conditions at all seven stations. At some of the sites, sorting the velocity record by daily mean wind speed can result in diurnal composites of velocity that significantly deepen and strengthen with stronger winds (Figure 6). To compare wind speed to jet speed, we calculate diurnal composites of near-surface velocity across a range of wind speeds at each separate mooring station. To construct each composite, we first sort the station's velocity record by the daily mean wind speed, from the day with the lowest wind speed to the day with the highest wind speed. We then take a diurnal composite of velocity for 120-day bins, where the days are sorted from weakest to strongest daily mean wind speed. To produce a range of composites, we employ a running 120-day bin, such that the first composite comprises the 120 weakest-wind days, the second composite comprises the second through 121st weakest-wind days, and so forth. Borrowing Schudlich and Price (1992)'s terminology, we calculate $\Delta v$, the diurnal change in along-wind velocity, as a measure of the strength of the jet for each of these composites. As in Figure 4, this $\Delta v$ is calculated with respect to the bottom-most observational bin, to remove the influence of the barotropic tide.

Figures 7a and 7b show the distribution of these wind-binned diurnal composites for each station, with the distribution of 30 composites represented by each box. As above, we see a pronounced difference between

Figure 5. Diurnal composites of reduced shear squared $M = \sum u^2 + \sum v^2 \cdot \Lambda^2$. 

Figures 7a and 7b show the distribution of these wind-binned diurnal composites for each station, with the distribution of 30 composites represented by each box. As above, we see a pronounced difference between
If the seasonal correspondence between wind speed and near-surface marginal instability gave the full explanation for the observed relationship between wind speed and jet speed at the eastern equatorial sites, we would expect to see a clear seasonality in the jets themselves. We were not able to find an obvious difference in the strength and depth of the jets at these sites in boreal spring vs. the other seasons, and indeed did not find an obvious seasonality even in the wind records, except slightly weaker winds in March/April/May at 2°S, 140°W. Thus, further work is necessary to unpack the observed linear relationship between wind speed and jet speed, perhaps with a longer observational record and/or a more targeted ADCP sampling scheme such that the daily diurnal jet could be resolved and compared to the daily local wind forcing.

Fitting the eastern equatorial estimates (including the sites at 0°, 180°, 0°, 155°W, 2°S, 140°W, and 2°N, 125°W) yields a clear relationship between wind speed and diurnal jet speed: for every increase of 1 m/s in wind speed, the jet appears to increase by approximately 1.7 cm/s at 17-m depth (Figure 7a), and by 1.1 cm/s at 31-m depth (Figure 7b), a relationship that steadily decreases with depth (Figure 7c). We find this empirical relationship between wind speed and diurnal jet amplitude only at sites in the central and eastern equatorial band. A similar linear relationship is not observed at the western and off-equatorial sites, although this wind-dependency may simply occur shallower than our observational window.

The fundamentals of the local physics governing the diurnal jet mechanism that are described in Price et al. (1986) and Fairall et al. (1996) do not predict the observed linear dependency upon wind speed. In these one-dimensional models, it is the thickness of the jet—rather than the jet velocity—that scales with wind stress. Thus, we expect stronger wind forcing not to drive a faster jet, but rather to drive a deeper jet over which the increased wind stress momentum is distributed ("mainly" because we are ignoring the secondary effects of time-dependent forcing and entrainment of momentum from below). The near-linear relationship between wind speed and diurnal velocity anomaly $\Delta v$ observed in this experiment must then be explained by other factors.

We might explain some of the observed wind dependence via larger-scale, lower-frequency controls on the state of marginal instability, rather than on the local dynamics of the jet itself. Bjerknes feedbacks and basin-scale Rossby and Kelvin wave responses link seasonal wind variation to the rise and fall of both the EUC and the thermocline in the eastern equatorial Pacific (Yu & McPhaden, 1999). In boreal spring (March/April/May), slackening trade winds coincide with shoaling of the EUC and the thermocline. At 0°, 140°W, Pham et al. (2017) demonstrated that this shoaling increases $\text{RI}$ in the ocean near-surface, such that a state of marginal instability is no longer present, and the diurnal jet fails to penetrate below the DWL in the spring. A predominance of these shallow springtime jets could yield the low composite estimate of $\Delta v$ at the observed depths. In the other seasons, a predominance of deep jets could yield the higher composite estimate of $\Delta v$. Trade winds in the region follow a similar seasonal pattern, slackening in the boreal spring and picking up in the other seasons, so that a low springtime $\Delta v$ might correspond to a large-scale slackening of the winds, and a higher $\Delta v$ might correspond to stronger winds in the other seasons.
4. Summary and Discussion

We have shown that the near-surface diurnal cycle in temperature and velocity is diverse across the tropical Pacific. Surface-forced heat and momentum penetrates much deeper at the central and eastern equatorial sites (at 0°, 180°, 0°, 155°W, 2°S, 140°W, and 2°N, 125°W) than at the three western stations. We hypothesize that the difference between these sets of stations is due to the marginally unstable background flow in the central and eastern equatorial Pacific, where a shallow SEC-to-EUC transition results in high background shear in the stratified near-surface ocean. At these locations, the diurnal jet carries wind-forced momentum as deep as 60 m, connecting wind forcing to the EUC on a diurnal time scale.

The diurnal evolution of shear and stratification at each site reinforces this view, showing a daytime shear layer that descends over the course of the day at the eastern equatorial sites (red, Figures 5e–5g and perhaps Figure 5d). At the western sites, no obvious descending diurnal pattern emerges in the reduced shear squared $M$ that compares the relative dominance of shear and stratification. Consequently, the diurnal jet is likely trapped above the observational window at the western sites, though the record-length diurnal composites shown here may exclude episodic deep diurnal jets that could occur at any of the sites.

At the eastern equatorial stations, we see a robust relationship between wind strength and jet strength. As far west as the 0°, 180° station and as far off the equator as 2°N and 2°S, a clear dependence between the magnitude of the diurnal cycle of velocity and daily mean wind speed emerges (Figure 7). A local explanation for this relationship is not obvious. The fundamentals of the one-dimensional physics described by Price et al. (1986) and Fairall et al. (1996) imply that the speed of the diurnal jet should be approximately independent of wind forcing: strong winds may produce a stronger wind-momentum flux, but this momentum is distributed over a deeper diurnal layer such that the modeled jet speed at any particular depth remains largely unaltered.

In the near-surface eastern equatorial region, marginal instability can be controlled by basin-scale physics that are not captured by the local one-dimensional model. Nonlocal seasonal controls on both marginal instability and wind speed could yield a correlation between wind speed and jet speed that results in the observed dependence between the two. Boreal springtime slackening of the trade winds in the region can correspond with the springtime disappearance of the marginally unstable state necessary for a deep diurnal...
jet; low winds correspond with a stable near-surface, and thus a shallow diurnal jet, while higher winds in the other seasons correspond with a marginally unstable near-surface and thus a deeper diurnal jet. We note that this seasonal correspondence does not specifically predict the observed linear relationship between wind speed and jet speed, though. Instead, the relationship could be higher order; if we predict that the wind momentum trapping depth does not depend on wind strength, for example, then we might expect the jet speed $\Delta v$ to be proportional to wind stress, and thus to wind speed squared.

Ultimately the data was not adequate to detect an obvious seasonality in the diurnal jet at these eastern equatorial sites, so further work is necessary to explain the observed dependence between wind speed and jet speed. A point-by-point comparison of wind speed to Richardson number would help to illuminate the observed dependence, but the temporal resolution of our velocity sampling scheme meant that an hourly estimate of both velocity and shear squared is dominated by noise. The ADCP instruments were set to observe for either 2 or 6 min per hour in order to conserve battery life; future deployments seeking to resolve the diurnal jet on a daily basis would have greater success with more frequent or longer sampling per hour.

Finally, though deep diurnal jets can certainly form in midlatitudes, the unique circumstances along the equator mean that systematic deep jets can occur throughout much of the year, enhancing the connection between the atmosphere and the thermocline on a diurnal time scale. Thus to further monitor and observe the role of the diurnal cycle in facilitating ocean-atmosphere communication, a focus on the central and eastern equatorial region would be most valuable.

Data Availability Statement

The current meter and ADCP data can be accessed via: https://www.pmel.noaa.gov/ocs/ndbc-tpos-moor-enhancement-pilot-project. Other moored data can be accessed via: https://www.pmel.noaa.gov/tao/drupal/disdel/. CCMP Version-2.0 vector wind analyses are produced by Remote Sensing Systems; the wind data are available at www.remss.com.

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References

Appell, G. F., Bass, P. D., & Metcalf, M. A. (1991). Acoustic doppler current profiler performance in near surface and bottom boundaries. *IEEE Journal of Oceanic Engineering, 16*(4), 390–396. https://doi.org/10.1109/48.40903
Bellenger, H., & Duvel, J.-F. (2009). An analysis of tropical ocean diurnal warm layers. *Journal of Climate, 22*(13), 3629–3646. https://doi.org/10.1175/2008jcli2598.1
Cronin, M. F., & Kessler, W. S. (2009). Near-surface shear flow in the tropical Pacific cold tongue front. *Journal of Physical Oceanography, 39*(5), 1200–1215. https://doi.org/10.1175/2008jpo4064.1
Cronin, M. F., McPhaden, M. J., & Weisberg, R. H. (2000). Wind forced reversing jets in the western equatorial pacific. *Journal of Physical Oceanography, 30*(4), 655–676. https://doi.org/10.1175/1520-0485(2000)030<0655:wfrijj>2.0.co;2
Danabasoglu, G., Large, W. G., Tribbia, J. J., Gent, P. R., Briegleb, B. P., & McWilliams, J. C. (2006). Diurnal coupling in the tropical oceans of CCSM3. *Journal of Climate, 19*(11), 2347–2365. https://doi.org/10.1175/jcli3739.1
de Boyer Montégut, C., Madec, G., Fischer, A. S., Lazar, A., & Iudicone, D. (2004). Mixed layer depth over the global ocean: An examination of profile data and a profile-based climatology. *Journal of Geophysical Research: Oceans, 109*, C1. https://doi.org/10.1029/2004JC002378
Fairall, C. W., Bradley, E. F., Godfrey, J. S., Wick, G. A., Edson, J. B., & Young, G. S. (1996). Cool-skin and warm-layer effects on sea surface temperature. *Journal of Geophysical Research, 101*(C1), 1295–1308. https://doi.org/10.1029/95JC03190
Hughes, K. G., Moum, J. N., & Shroyer, E. L. (2020). Evolution of the velocity structure in the diurnal warm layer. *Journal of Physical Oceanography, 50*(3), 615–631. https://doi.org/10.1175/jpo-d-19-0207.1
Inoue, R., Lien, R.-C., & Moum, J. (2012). Modulation of equatorial turbulence by a tropical instability wave. *Journal of Geophysical Research, 117*, C11009. https://doi.org/10.1029/2011JC007767
Johnson, G. C., McPhaden, M. J., & Firing, E. (2001). Equatorial Pacific Ocean horizontal velocity, divergence, and upwelling. *Journal of Physical Oceanography, 31*(3), 839–849. https://doi.org/10.1175/1520-0485(2001)031<0839:epohvd>2.0.co;2
Kudryavtsev, V. N., & Soloviev, A. V. (1990). Slippery near-surface layer of the ocean arising due to daytime solar heating. *Journal of Physical Oceanography, 20*(5), 617–628. https://doi.org/10.1175/1520-0485(1990)020<0617:snslo>2.0.co;2
Lien, R.-C., Caldwell, D. R., Gregg, M. C., & Moum, J. N. (1995). Turbulence variability at the equator in the central Pacific at the beginning of the 1991–1993 El Niño. *Journal of Geophysical Research, 100*(C4), 6881–6898. https://doi.org/10.1029/94JC03312
Liu, C., Köhl, A., Liu, Z., Wang, F., & Stammer, D. (2016). Deep-reaching thermocline mixing in the equatorial Pacific cold tongue. *Nature Communications, 7*, 11576. https://doi.org/10.1038/ncomms11576
Moulin, A. J., Moum, J. N., & Shroyer, E. L. (2018). Evolution of turbulence in the diurnal warm layer. *Journal of Physical Oceanography, 48*(2), 383–396. https://doi.org/10.1175/jpo-d-17-0170.1
Moum, J. N., Caldwell, D. R., & Paulson, C. A. (1989). Mixing in the equatorial surface layer and thermoline. *Journal of Geophysical Research, 94*(C2), 2005–2022. https://doi.org/10.1029/98JC002005
Moum, J. N., Lien, R.-C., Perlin, A., Nash, J. D., Gregg, M. C., & Wiles, P. J. (2009). Sea surface cooling at the equator by subsurface mixing in tropical instability waves. *Nature Geoscience, 2*(1), 761–765. https://doi.org/10.1038/ngeo657
