Seasonal and Interannual Variations in the Daily Cycle of Winds over the Galápagos

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ABSTRACT

While the daily cycle of near-surface winds over the equatorial east Pacific has been studied in some detail, little is known about the daily cycle above the surface layer. Furthermore, the causes of the observed near-surface daily cycle are not well understood. A better understanding of the structure and forcing mechanisms at work on the lower-tropospheric winds over this region may increase our appreciation for the varying importance of local and remote atmospheric and oceanic processes. This study documents the daily cycle of lower-tropospheric winds over one of the Galápagos Islands during the late 1990s, as well as how it varied seasonally and interannually, using half-hourly profiler winds. The well-known zonal semidiurnal tide is evident in the data, as is a diurnal cycle that is predominantly meridional and may be driven by the results of convection over the Andes. In addition, the high vertical resolution of the wind profiles reveals a decoupling of the daily cycle of winds below ~500 m from those aloft during periods of cold (<23°C) SSTs. This decoupling, which is not evident in long-term mean profiles that effectively filter the daily cycle, may inhibit the vertical mixing of momentum or vertical propagation of tidal or wave signatures over the region.

1. Introduction

The mean daily cycle of near-surface winds over the equatorial Pacific has been well documented by Deser (1994) and Deser and Smith (1998, hereafter DS98). Deser (1994) documented the daily cycle, including diurnal and semidiurnal harmonics, using data collected by 14 Tropical Atmosphere–Ocean (TAO) buoys during the summer of 1992. She found that the meridional daily cycle was stronger than the zonal one and postulated a connection to flow associated with the ITCZ. The buoy winds also clearly showed the semidiurnal tide in the zonal wind component. DS98 extended the analysis of TAO winds in time and space. Their results, using 1993–96 data from 54 buoys, were qualitatively similar to those of Deser (1994). The semidiurnal tide dominated variations in $u$ except along the equator at 125°W, 110°W, and 95°W, while diurnal variations dominated variations in $v$ over the east Pacific “cold tongue.” DS98 were also able to identify the diurnal cycle of the divergence field over the equatorial Pacific and its seasonal variations in the east.

The mean daily cycle of winds above the surface of the equatorial Pacific is less well researched. A regional study of the daily cycle in lower-tropospheric winds was conducted by Gutzler and Hartten (1995) using data collected over the western Pacific. They, too, found evidence of the semidiurnal tide. They also found a daily signal that they attributed to the propagation of an outflow boundary from Indonesian convection. The extent to which those results hold for other seasons or years is not known. Similar work over the central or eastern Pacific is lacking.

We have used 4 years of profiler data from San Cristóbal in the Galápagos Islands to investigate the daily cycle of lower-tropospheric winds over the east Pacific Cold Tongue. In order to evaluate the stationarity of the results, we have computed a mean daily cycle over several time periods. Viewing the results as anomalies from a daily mean highlights the vertical structure of the flow and its variability, while spectral analysis enables us to offer hypotheses about the causes of some of the elements of the daily cycle.

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2. Data sources and analysis methods

a. Profiler winds

The primary data used in this study are half-hourly average wind profiles collected over one of the Galápagos Islands. Since 1994, the National Oceanic and Atmospheric Administration (NOAA)/Aeronomy Laboratory has operated a 915-MHz wind profiler and some surface instruments at the National Institute of Meteorology and Hydrology of Ecuador (INAMHI) observation station. This site is in Puerto Baquerizo Moreno, (0.90°S, 89.61°W), on the western tip of Isla San Cristóbal in the Galápagos. San Cristóbal, the archipelago’s easternmost island, lies in the heart of the east Pacific Cold Tongue (Fig. 1a); the SST front almost always lies north of the archipelago, and satellite imagery shows little SST gradient between 90° and 95°W. The island extends 43 km from southwest to northeast, with a long southern coast; its shape and topography are shown in Fig. 1b. The profiler is located at 8 m MSL (above mean sea level). To the east of Puerto Baquerizo Moreno, the land rises gently for 3.7 km to an altitude of about 150 m before rising more sharply over the next 7 km to a poorly mapped peak several hundred meters above sea level.

The data used in this study were post-processed as described in Riddle et al. (1996). This study uses October 1994 through December 1999 winds from ten altitudes between ~300 and ~2600 m and also from NOAA’s surface instruments. The profiler was not operational between early October 1995 and early August 1996, and winds below 500 m were not available from November 1998 through June 1999. Other gaps in the data are sporadic in time and height. Previous work by Hartten and Gage (2000) has shown that seasonal-mean wind profiles from this instrument are consistent with other observations in the area.

b. Tropical indices

Two different tropical indices were used to organize the data during this study. The Southern Oscillation index (SOI) is based on the normalized difference between surface pressures at Tahiti and Darwin; it gives a basinwide view of the state of the tropical Pacific, with negative (positive) values indicating warm (cold) ENSO conditions, respectively. The Niño-1+2 index is based on SSTs averaged over the region extending from 0° to 10°S and from 90° to 80°W. It gives a local view of the state of the east Pacific. Time series of monthly SOI and Niño-1+2 SSTs, and their anomalies calculated relative to the 1961–90 base period, were obtained from the National Centers for Environmental Prediction (NCEP); a 5-month running mean was then applied to filter out short-term fluctuations.

c. Near-surface atmosphere and ocean measurements

This study also made use of near-surface data collected from October 1994 through December 1999 by the TAO buoys moored in the east Pacific (McPhaden 1995). Daily air temperature, $T_{\text{air}}$, and SST from the buoys at 0°, 95°W and 2°S, 95°W were obtained from the TAO Project Office at NOAA/Pacific Marine Environmental Laboratory (PMEL). Air temperature was measured at 3 m MSL, while SST was measured 1 m below the surface. There were several data gaps exceeding 10 consecutive days at each site, as shown in Fig. 2, as well as some smaller gaps.
3. Results

3.1 Vector wind anomalies

Figure 3 shows the mean daily cycle for the Full9499 case plotted as vector wind anomalies from the daily average, together with the subtracted mean profile plotted to the right in vector form. Below 500 m, anomalies are northerly at night and southerly during the day, in agreement with the results of DS98. Above that, the anomalies look quite different; they are predominantly easterly from about 0700 LT to shortly after noon and northwesterly from about 1900 LT until early in the morning. Transitions in both speed and direction are fairly smooth between 500 and 2600 m, with the largest anomalies between about 1.0 and 1.5 km. Thus, the flow near the surface appears to be decoupled from that aloft. This decoupling occurs well below the shift from strong southeasterlies to weaker northeasterlies that marks the top of the trade inversion near 1750 m. If the daily cycle was frequently driven by a land/sea breeze circulation, we would expect to see daytime flow in the lowest 200 to 1000 m of the atmosphere directed toward the heated region for at least a few hours (Wexler 1946; Leopold 1949). The daily cycle in Fig. 3 seems inconsistent with a land/sea breeze, since the midday flow is directed neither toward the center of Isla San Cristóbal, nor the mountain along its south shore, nor the “center of mass” of the entire Galápagos archipelago.

Conditions during the SOLNiño and SOLNiña cases are shown in Figs. 4 and 5. The results from the SST_warm and SST_cold cases are similar. The SOLNiña results at 325 m are not plotted because fewer than 40% of the total possible observations were present at some times. During El Niño conditions, the daily cycle involves predominantly meridional anomalies at all heights, southerly (veering with height to southeasterly) during the middle of the day and north-northwesterly (veering to northerly) during the middle of the night. The anomalies below 500 m are very similar to those in Fig. 3. In contrast, during La Niña conditions the near-surface vector

| Table 1. Case study names, definitions, and dates. SST$^\ddagger$ refers to monthly anomalies of the Niño-1+2 index (Reynolds 1988). A 5-month running mean was applied to both the SST$^\ddagger$ and the SOI time series before they were evaluated according to the listed criteria. When available, profiler and TAO data were used from the first day of the initial month through the last day of the final month in each case. |
|---|---|---|
| Case | Description | Start date(s) | End date(s) |
| Full9499 | Monthly SOI \(-1\) | Oct 1994 | Dec 1999 |
| SOLNiño | Monthly SOI \(-1\) | Oct 1994 | Nov 1994 |
| SST_warm | Monthly SST\(+1\,^\circ C\) | May 1997 | Apr 1998 |
| SOLNiña | Monthly SST\(+1\,^\circ C\) | Aug 1998 | Feb 1999 |
| SST_cold | Monthly SST\(-1\,^\circ C\) | Jun 1996 | Oct 1996 |
| 95warm | 1995 warm season | Feb 1995 | May 1995 |
| 97warm | 1997 warm season | Feb 1997 | May 1997 |
| 99warm | 1999 warm season | Feb 1999 | May 1999 |
| 99cold | 1999 cold season | Jul 1999 | Oct 1999 |

* No profiler data were available for Jun and Jul 1996.
anomalies show a dramatic decoupling from the flow aloft, with directions similar to the Full9499 case but with magnitudes twice as large. From 500 to 2000 m the daily cycle is dominated by zonal anomalies, easterly from about 0700 to 1400 LT and westerly from about 1800 until 0300 LT. The largest amplitudes in this layer occur from about 750 to 1500 m.

Most of the results from individual warm and cold seasons are similar to those shown in Figs. 4 and 5, respectively. The exception is the 97cold case (Fig. 6), which had deep meridional flow instead of the decoupled flow seen in other cold periods. We attribute this to the strong sea surface warming accompanying the 1997 El Niño episode. It is not clear why the daily cycle over San Cristóbal was disrupted during July through October 1997, behaving very much like that during a typical warm season, while the annual cycle was not disrupted until later in the year, during the December–January–February (DJF) season (Hartten and Gage 2000).

Previous studies have shown that SSTs in the east Pacific cold tongue affect the overlying surface winds...
Fig. 5. The daily cycle of lower-tropospheric winds over San Cristóbal during months with SOI $\geq 1$ (SOLNiña case).

Fig. 6. The daily cycle of lower-tropospheric winds over San Cristóbal during the 97cold case, Jul through Oct 1997.

(Hayes et al. 1989; Wallace et al. 1989; Chelton et al. 2001). Wallace et al. (1989) hypothesized that the SSTs modify the stability of the atmospheric boundary layer and thereby affect the surface wind field. The fact that the vertical decoupling of the daily cycle of winds appears to occur during cold periods but not warm ones suggests that the effects of the near-surface stability extend well above the surface. To test that idea, we computed the daily air–sea temperature difference, $T_{\text{air}} - \text{SST}$, at both the $0^\circ$, $95^\circ$W and $2^\circ$S, $95^\circ$W buoys, then averaged daily $T_{\text{air}}$, SST, and $T_{\text{air}} - \text{SST}$ over the time periods shown in Table 1. (Data were not available from $2^\circ$S, $95^\circ$W during the 99cold case.) The results are plotted in Fig. 7. Cases were classified as “decoupled” or “coupled” based on visual inspection of Figs. 3–6 and analogous plots for the remaining cases. Surprisingly, the association between decoupling and the air–sea temperature difference is weak (Fig. 7b), although the only time the difference was nonnegative was at $2^\circ$S, $95^\circ$W during the decoupled cases. A much clearer separation between decoupled and coupled cases is seen in the SST data, with SSTs $< 23^\circ$C at both buoys during decoupled cases and $>24^\circ$C during coupled ones. (While the division between decoupled and coupled flow generally occurs at SST = $23^\circ$C, SST was slightly higher than that when averaged over the full time period studied;
yet Fig. 3 shows a slight decoupling of the flow. This could be an artifact of averaging flow fields that are distributed in a bimodal fashion, or it could reflect different data gaps in the wind and TAO observations.)

b. Diurnal and semidiurnal cycles

Spectral analysis of the mean daily cycle was used to construct Fig. 8, which shows profiles of the amplitude and phase (local time of maximum amplitude) for both the diurnal and semidiurnal harmonics of $u$ and $v$ during warm and cold season cases. The two harmonics are most easily studied by focusing on these cases. Figure 8 shows that the diurnal harmonic usually accounts for most of the total variance in both $u$ and $v$. However, the semidiurnal harmonic exhibits coherence in time and space that makes its study interesting as well. The decoupling of the low-level flow from that aloft during cold conditions is not obvious when the data are viewed solely in terms of the first two daily harmonics although, as discussed below, some of the decoupling effects can still be seen.

Considering first the semidiurnal harmonic, in the zonal wind it is fairly consistent in amplitude and phase; the mean amplitude of the data plotted in Fig. 8 (all heights and cases) is 0.18 m s$^{-1}$, and the mean phase is 1510 LT. The atmospheric semidiurnal tide phase is invariant with height, and maximum westerly anomalies are expected at 0344 and 1544 LT (Whiteman and Bian 1996). Figure 8 shows that semidiurnal $u$ anomalies with amplitude 0.15 to 0.22 m s$^{-1}$ and phase 1514 to 1614 LT were common above 700 m during all the averaging periods. Below 700 m, phases tended to be earlier than expected by tidal theory, as was the case with the TAO data studied by DS98. In addition, low-level amplitudes tended to be higher than average during some seasons (95warm and the decoupled flow cases 98cold and 99cold) and lower during the others. This is in contrast to the results of DS98, who found semidiurnal fluctuations $\leq$0.10 m s$^{-1}$ at the cold tongue buoys along 140°, 125°, and 110°W and semidiurnal fluctuations of 0.15 and 0.13 m s$^{-1}$ at the 0°, 95°W and 2°S, 95°W buoys. They suggested that a very stable boundary layer over the cold tongue might be keeping the full magnitude of the semidiurnal tide from propagating to the surface. However, our results seem to indicate the opposite at San Cristóbal, namely that decreased vertical mixing during periods of increased lower boundary layer stability permits a higher-amplitude semidiurnal tide to exist near the surface. (Fig. 7 shows that 95warm was the coupled season with the smallest mean air–sea temperature difference.) Such a scenario would require the presence of enough water vapor in the lowest 500 m of the atmosphere to allow solar heating to drive tidal motions there. Further exploration of this contradictory evidence is left for future research.

Turning now to the diurnal harmonic, the amplitude of this wave in each component ranges from about 0.2 to over 1.0 m s$^{-1}$, with somewhat higher amplitudes and a smoother profile in $u$ than in $v$. In the zonal component, the time of maximum positive anomalies is usually 2200 to 0200 LT above 700 m, although there are some departures from this, especially above 2000 m in decoupled cases and during the 98warm case. The phase of the observed diurnal harmonic rarely agrees with the phase of the predominant zonal diurnal atmospheric tide (1, 1, 1), whose maximum westerly anomalies are expected at $\sim$1400 LT near the surface and at $\sim$1700 LT near 3 km (Williams and Avery 1996). [The (1, 1, 1) notation indicates that this tidal mode has a frequency of 1 oscillation per day, a zonal wavenumber of 1 and westward propagation, and a meridional (Hough function) index of 1.] In the meridional component, a midday phase is common, although an early morning maximum is found between 500 and 1000 m in the decoupled and 98warm cases.

Figure 8 shows that the zonal and meridional diurnal harmonics often appear to be about 180° out of phase;
Fig. 8. Profiles of the amplitude and phase of the (left) zonal and (right) meridional diurnal and semidiurnal waves for the warm and cold season cases. The phase is plotted as the local time of maximum positive values. The diurnal harmonic is represented by a solid diamond; the semidiurnal harmonic by an open square. In the zonal phase profiles, the hour centered on the theoretical phase of the zonal semidiurnal tide is shaded gray. In the meridional phase profiles, the hour centered on 1200 LT is shaded gray.
component led the diurnal tide. The next obvious cycle is the low-level one that matches the expected characteristics of the zonal semi-diurnal wave. The 10-14-h phase difference was 177°.

We conducted a further analysis by forcing the phase of \( u \) to lead the phase of \( v \) and focusing on the 10–14-h phase-difference window. Data from 98 Warm never fell into that window and, as noted earlier, the profile of the 98 Warm zonal diurnal cycle phase is quite different from those of the other cases, so we have excluded 98 Warm data from this analysis. The remaining cases leave us with 64 \((u, v)\) phase pairs at various heights. The phase difference between the diurnal harmonics fell within the 10–14-h phase-difference window during 33 out of these 64 instances, or more than half the time. Considering only those 33 heights and times during which the diurnal harmonic of \( u \) led that of \( v \) by 10–14 h, the mean phase difference was 177° (11.8 h). The mean amplitude and phase of the \( u \) diurnal harmonic of those 33 values was 0.37 m s\(^{-1}\) and 69° (0436 LT), while the mean amplitude and phase of the \( v \) diurnal harmonic was 0.44 m s\(^{-1}\) and 252° (1648 LT). We can thus define a typical lower-tropospheric condition during six of the studied seasons as a diurnal cycle with northwesterly anomalies in the middle of the night and southeasterly anomalies in the middle of the day (Fig. 9). This signal is fairly robust, although it is absent during the 98 Warm season and confined to the 1–2-km range during cold seasons with decoupled flow. It appears, therefore, to be a large-scale feature rather than a response to local conditions.

4. Discussion

What are the causes of daily cycles seen in the San Cristóbal profiler data? The easiest cycle to identify and explain is the semidiurnal (12 h) cycle in \( u \), which while small, is nearly ubiquitous (Fig. 8) and which closely matches the expected characteristics of the zonal semidiurnal tide. The next obvious cycle is the low-level one dominated by a roughly diurnal cycle in \( v \) (Figs. 3–6), which appears to be trapped at low levels during times when the mean SST is lower than 23°C and mixed upward during times when the SST is warmer than 24°C. Finally, there is the pure diurnal (24 h) cycle, which in its typical form is such that \( u \) leads \( v \) by about 12 h and at about 80% amplitude.

DS98 also found a strong meridional diurnal cycle in the TAO surface winds about 5° west of the Galápagos. They suggested that the diurnal cycle of near-surface wind divergence was linked to a deep meridional tropospheric circulation. Earlier, Gutzler and Hartten (1995) had hypothesized that the daily cycle of lower-tropospheric winds over part of the western equatorial Pacific could be caused by an inflow boundary associated with the diurnally varying deep convection over the island of New Guinea propagating outward at gravity wave speeds. In the east Pacific, an obvious source of deep convergence at a similar longitude to San Cristóbal would be convection in the ITCZ. Over the seven warm or cold seasons studied here the average latitude of the outgoing longwave radiation (OLR) minimum along 90°W (assumed to be collocated with the ITCZ) varied from 12.5°N (1999 cold season) to 0.2°S (1998 warm season). However, the phase of the 17-m meridional diurnal cycle varied little, from 1135 LT (1995 warm season) to 1347 LT (1999 cold season). The correlation between the 17-m phase and the distance between San Cristóbal and the ITCZ is only 0.21. Figure 8 shows that the phases of the meridional diurnal cycle in the free atmosphere (i.e., above about 2 km) were quite similar to those near the surface. Thus, a simple relationship between the meridional diurnal cycle at San Cristóbal and the ITCZ does not seem to exist.

Recently Garreaud and Muñoz (2004) have used a regional model to identify a daily cycle off the west coast of central South America that is driven by convection over the mountains. Their model extends northward only to (7.5°S, 87.0°W); however, northward extrapolation of the convectively induced upsurge wave at 800 hPa that propagates off the mountains during one month (November 2001) indicates the upsurge wave would likely reach the Galápagos near 0900 UTC (0300 LT; R. Garreaud, 2003, personal communication). If the upward motion in this wave were supported by convergence toward the wave axis, that is, northwesterlies ahead and southeasterlies behind, then this is a few hours sooner than might be expected based on the diurnal cycle shown in Fig. 9, which has a null value (corresponding to maximum vertical motion) at about 0700 LT. However, the possibility that the diurnal cycle over San Cristóbal is driven by continental convection is intriguing and deserves further study.

5. Conclusions

Long-term-mean daily cycles of lower-tropospheric winds over San Cristóbal in the Galápagos Islands have...
been computed for the entire 4-yr data record; for months during El Niño/warm events, for months during La Niña/cold events, and for individual warm and cold seasons. There is a strong daily cycle during all averaging periods. The profiler winds do not show any signs of a land/sea breeze circulation; this increases our confidence that they are representative of conditions over the eastern near-equatorial cold tongue. There is often evidence of a semi diurnal tidal signature in the zonal wind, especially at and above 800 m, but not of the (1, 1, 1) diurnal tide.

During El Niño/warm events, warm seasons, and the 1997 cold season the daily cycle projects predominantly onto meridional anomalies and is fairly consistent with the 1998 and 1999 cold seasons it projects onto zonal anomalies above 500 m and meridional anomalies below that. The decoupling of the low-level flow during these periods is presumably caused by the enhanced stability of the lowest levels of the atmosphere caused by the very cold underlying SSTs. During periods with coupled flow, SSTs recorded by the TAO buoys at 0°, 95°W and 2°S, 95°W were >24°C, while during periods with decoupled flow SSTs were <23°C. The 1997 cold season (Fig. 5) behaved differently than the other cold seasons, lacking the decoupling of the near-surface flow from that aloft and therefore having a deep meridional anomaly pattern. We attribute this to the strong sea surface warming accompanying the 1997 El Niño episode. The decoupling appears to lead to a higher-amplitude semi diurnal tide at low levels, although further work on the forcing of that tide in the lower troposphere is necessary to confirm that. The decoupling also seems to affect the transport of momentum between the lowest and middle levels of the troposphere.

The low-level daily cycle is dominated by the meridional diurnal (24 h) mode. During six of the seven seasons studied, more than half of the observed winds showed the zonal component of the diurnal cycle to lead the meridional component by 10 to 14 h. This results in a typical diurnal cycle of northwesterly anomalies in the middle of the night and southeasterly anomalies in the middle of the day. The diurnal cycle near the surface is inconsistent with simple theories relating it to convergence into convection in the ITCZ but may be consistent with convergence into an upshear wave propagating away from convection over the South American mountains.

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