1. Introduction

Recent satellite observations provide an unprecedented view of mesoscale oceanic eddies and their interactions with the atmosphere over the global ocean (Chelton et al. 2004; Xie 2004). One of the most notable forms of such mesoscale ocean–atmosphere interactions is tropical instability waves (TIWs), which are commonly observed phenomena over the equatorial cold tongue of the Pacific and Atlantic Oceans. TIWs are manifested as westward-propagating wavelike oscillations near the SST fronts in the equatorial cold tongues. Satellite observations reveal that SST perturbations are highly coherent in space with variations in surface wind (Hashizume et al. 2001; Jochum et al. 2007b). The focus of this analysis is to examine to what extent the recently completed high-resolution, and partially coupled, Climate Forecast System Reanalysis (CFSR) at the National Centers for Environmental Prediction (NCEP) (Saha et al. 2010) is capable of detecting ocean–atmosphere interactions associated with TIWs in the equatorial Pacific.

Numerous studies have been devoted to understanding TIWs and their influences on climate variability. TIWs have been recognized as a key contributor to upper-ocean momentum balance, heat budget, and nutrient distribution (Weisberg 1984; Chavez et al. 1999; Jochum and Murtugudde 2006; Moum et al. 2009). The
unique role of TIWs in air–sea interactions and climate variability is further highlighted by Xie (2004), who showed that small-scale SST variability associated with TIWs is positively correlated with the near-surface wind speed (WS), which differs from the out-of-phase relationship between SST and wind speed on basin scales. Furthermore, the TIW-induced atmospheric perturbations could potentially feedback onto the SST variability (Pezzi et al. 2004), with a net impact on the ocean mean state and the characteristics of El Niño–Southern Oscillation (ENSO) (Seo et al. 2007a; An 2008; Zhang and Busalacchi 2008; Imada and Kimoto 2012).

Given the significance of TIWs on different aspects of climate variability, it is important to understand details of the associated ocean–atmosphere interactions. Because of the paucity of in situ observations, the significance and complexity of TIWs was not recognized until near-all-weather-coverage microwave satellite measurements became available in the late 1990s. The atmospheric variations associated with TIWs involve processes throughout the depth of the planetary boundary layer (Hashizume et al. 2002; Xie 2004). However, satellite can only measure limited surface variables, making it difficult to understand detailed mechanisms of mesoscale air–sea couplings. Moreover, the short satellite data record has a limited utility in assessing the influence of TIWs on climate variability. Although numerical weather prediction (NWP) reanalysis products, which provide a three-dimensional rendition of various oceanic and atmospheric fields, could be particularly useful for understanding details in the atmospheric variations related to TIWs, their low horizontal resolution has been an impediment. For example, the widely used European Centre for Medium-Range Weather Forecasts (ECMWF) and NCEP reanalysis products were not able to resolve mesoscale atmospheric variability because of coarse model resolutions (Chelton 2005).

The recently completed CFSR has three novel features that could make it a valuable dataset for the analysis of TIWs: 1) a 31-yr period from 1979 to 2009 is available at much higher spatial and temporal resolution; 2) assimilation of a high-resolution daily SST analysis is at 0.25° grid resolution; and 3) the first-guess field for the analysis is generated from a 6-h coupled model integration, thus maintaining physical consistency among ocean–atmosphere interactions and their evolution. The purpose of this study is to assess the characteristics of the TIWs in the equatorial Pacific simulated by the CFSR and how well they agree with in situ and satellite observations, with the aim to inform the user community about the feasibility of CFSR as an analysis and monitoring tool for TIWs.

2. CFSR and validation

a. CFSR

The CFSR is the latest generation of NCEP reanalysis with the first guess from the NCEP Climate Forecast System (CFS), a fully coupled atmosphere–ocean–sea–ice–land model (Saha et al. 2010). The atmosphere component is the operational NCEP Global Forecast System (GFS), which is run at a spectral resolution of T382 (~38 km) with 64 vertical layers. Some of the parameterizations in the GFS include a simplified Arakawa–Schubert cumulus convection with cumulus momentum mixing (Pan and Wu 1995; Hong and Pan 1996), orographic gravity wave drag (Kim and Arakawa 1995), and a cloud microphysics scheme to determine cloud condensate prognostically. The CFSR uses the gridpoint statistical interpolation (GSI) data assimilation system for the atmosphere. All available conventional and satellite observations were assimilated in the CFSR. Here we only list observation sources relevant to this study: conventional observational inputs include data over land surface, ship, and buoy observations; radiosondes; pilot balloons (PIBALS); wind profiler; and aircraft. Satellite inputs include radiance assimilation, upper-air winds derived from geostationary satellite, surface winds from the Special Sensor Microwave Imager (SSM/I), European Remote Sensing Satellites (ERS-I and ERS-2), Quick Scatterometer (QuikSCAT) SeaWinds, and Naval Research Laboratory (NRL) WindSat scatterometers.

The ocean component of the coupled model is the Geophysical Fluid Dynamics Laboratory (GFDL) Modular Ocean Model version MOM4p0d (Griffies et al. 2004), which is a finite-difference model with a zonal resolution of 0.5°, a meridional resolution that varies from 0.25° between 10°S and 10°N to 0.5° poleward of 30°S and 30°N, and 40 layers in the vertical with 27 layers in the upper 400 m. The ocean data assimilation for the CFSR uses a 3D variational (3DVAR) analysis scheme that incorporates a temporal and spatial varying background error covariance (Behringer et al. 1998) and is a modified version developed by Derber and Rosati (1989). Temperature and salinity profiles are assimilated at 6-h intervals using all of the observation within a previous 10-day window. The more distant an observation is in time from the target analysis, the less weight it receives in the assimilation. For the top level of the model (5 m), the temperature analysis is strongly nudged to the daily optimum interpolation (OI) SST product (Reynolds et al. 2007), and the salinity to the mean climatology based on the World Ocean Database 1998 (Conkright et al. 1999). Temperature observations used in the assimilation include profiles from expendable bathythermographs (XBTs), fixed mooring arrays—the Tropical Atmosphere

In this study, the set of CFSR variables analyzed includes SST, surface heat flux, surface wind stress, surface pressure, cloud cover, air temperature, wind velocity, specific humidity, geopotential height, and ocean current velocity. It is noteworthy that the 10-m wind for the CFSR is obtained based on extrapolating the wind at the lowest model level (~25 m over the ocean) using the stability-dependent surface friction velocity and assuming a logarithmic profile that is typical of wind profile in boundary layer under neutral conditions. The CFSR data are available at hourly resolution and 0.5° grid. Daily means derived from hourly output are used in this study. Readers are referred to Saha et al. (2010) for more detail about the CFSR.

b. Validation data

1) Satellite Data

SST is from the Tropical Rainfall Measuring Mission (TRMM) Microwave Imager (TMI). TMI data are not affected by clouds (except for regions of heavy rain), and hence provides essentially an uninterrupted record of the SST signatures of TIWs. Surface wind velocity is from the SeaWinds scatterometer on the QuikSCAT satellite. It is noteworthy that scatterometers measure small-scale ocean surface roughness, which is dependent on the surface stress. However, the surface stress is not just a function of wind speed at some height above the surface but also depends on the static stability near the surface. The QuikSCAT wind retrievals are considered as the equivalent neutral-stability wind at a reference height of 10 m above the sea surface (Chelton and Freilich 2005). We use the daily TMI SST (version 4, available since December 1997) and QuikSCAT wind products (available from July 1999 to November 2009) at 0.25° grid from Remote Sensing Systems (http://www.ssmi.com). These products were interpolated on the same 0.5° grid as for the CFSR.

2) In Situ Data

Daily ocean temperature at 1-m depth, referred to as TAO SST, wind velocity at 4 m, and current velocity measurements from the TAO mooring array (McPhaden 1995) are used as references to validate the CFSR-analyzed fields and satellite observations.

3) SST Analysis

Version 2 daily SST analyses on a 0.25° grid from the National Climatic Data Center (NCDC) (Reynolds et al. 2007) is also used in this study. This analysis blends ship, buoy, and satellite measurements of SST using an OI procedure (OISST). These data are available since November 1981 and were assimilated in the CFSR. The OISST was interpolated onto the same 0.5° CFSR grid.

c. Validation Results

TIWs typically have a period of about 30 days (Qiao and Weisberg 1995; Lyman et al. 2007; Shinoda et al. 2009). To describe variations associated with TIWs, a Butterworth filter with 20–40-day bandpass is applied to the time series of the data to emphasize variability centered at 30 days. Hereafter, all datasets discussed refer to filtered data.

Time series of the filtered CFSR SST and TMI SST were validated against the TAO SST at three mooring sites along 2°N at 140°, 125°, and 110°W. We note that the TAO SST is not an independent validation dataset, as it was assimilated in the CFSR. Figure 1a shows a segment (2006–09) of filtered SST at 2°N, 110°W. It shows that the CFSR SST (red line) agrees very well with the TAO SST (black line). Similar results were obtained at two other sites mentioned above (results not shown). Table 1 provides the comparison statistics at the three TAO mooring sites during 2001–08. The anomaly correlations (ACs) between the CFSR and TAO SST are all greater than 0.9, and the amplitude for the CFSR is roughly about 74% of the TAO according to regression coefficients (RCs). As expected, TMI SST is highly consistent with TAO SST (AC is 0.92–0.97, RC is about 1).

Monthly variance of the filtered SST at 2°N, 110°W during 1979–2009 is shown in Fig. 1b. The monthly variance is defined as the monthly mean of the squared daily filtered SST. The OISST (green line) underestimates the TIW variance before 2002 relative to TAO but agrees better afterward. Reynolds et al. (2007) documented that there is an increase in SST variance when the near-all-weather-coverage Advanced Microwave Scanning Radiometer (AMSR) was used from June 2002 onward compared to the period when only the Advanced Very High Resolution Radiometer (AVHRR) infrared satellite SST data were used before June 2002. Our result is consistent with their finding. We also note that the TIW variance in the CFSR SST (red line) is very weak before 1982 because only monthly mean SST was assimilated. The results indicate that discrepancy between CFSR and TAO during 1982–2002 can be partially attributed to weaker variance in the OISST, since the CFSR SST is strongly nudged toward the OISST.
WS perturbations from the CFSR and the QuikSCAT are also validated against the TAO measurements. The CFSR winds are highly correlated ($AC \geq 0.86$) with the TAO winds, and their amplitudes are about 80%–85% of the TAO (Table 1). It is noteworthy that the CFSR has higher correlation with TAO than the QuikSCAT at all three sites. This may be because the QuikSCAT measures winds relative to the moving sea surface, while the TAO provides wind estimates relative to a fixed location. Indeed, it has been demonstrated that departures of QuikSCAT from the TAO can be related to ocean surface currents (Kelly et al. 2001; Polito et al. 2001).

One may argue that good agreement between the CFSR and TAO winds at the TAO mooring sites is largely because the TAO winds were assimilated in the model. It gives rise to the question of whether variables in the CFSR above the surface (i.e., in the atmosphere) or below the surface (i.e., in the ocean) where observational data are sparse also contain signatures related to the TIWs. Thus, we next examine eddy kinetic energy (EKE) in the CFSR, a quantity that is generally used to represent TIW activity in the ocean (e.g., Weisberg and Colin 1986; Jochum et al. 2004). Details of atmospheric response to TIW-induced SST are discussed in section 3.

Figure 2 shows the time evolution of EKE derived from the CFSR and TAO acoustic Doppler current profiler (ADCP) measurements on $0^\circ$N, $110^\circ$W. Similar to previous studies (Jochum et al. 2004; Seo et al. 2007a), the EKE is defined as $EKE = (u'^2 + v'^2)/2$, where $u'$ and $v'$ are daily 20–40-day bandpass-filtered zonal and meridional current velocities, respectively. The CFSR shows clear TIW signals in the top 55 m that are largely in phase

| Table 1. Comparison statistics of CFSR/TMI with TAO SST, CFSR/QuikSCAT 10-m wind speed (WS10) with TAO wind at three mooring sites along $2^\circ$N at $140^\circ$W, $125^\circ$W, and $110^\circ$W during 2001–08. Shown are root-mean-square difference (RMS) and standard deviation (std dev) of TAO data. Note all data are daily bandpass filtered. |
|-----------------|-----------------|-----------------|-----------------|-----------------|-----------------|-----------------|
|                 | SST\(^\circ\)C |                 |                 |                 |                 |                 |
|                 | $2^\circ$N, 140\(^\circ\)W | $2^\circ$N, 125\(^\circ\)W | $2^\circ$N, 110\(^\circ\)W |                 |                 |                 |
|                 | CFSR | TMI | CFSR | TMI | CFSR | TMI | CFSR | QSCAT | CFSR | QSCAT | CFSR | QSCAT |
| AC              | 0.93 | 0.97 | 0.9 | 0.92 | 0.91 | 0.97 | 0.91 | 0.89 | 0.9 | 0.87 | 0.86 | 0.82 |
| RMS (std dev)   | 0.12 | 0.08 | 0.18 | 0.21 | 0.21 | 0.13 | 0.20 | 0.23 | 0.2 | 0.23 | 0.23 | 0.26 |
|                 | (0.3) | (0.3) | (0.39) | (0.39) | (0.5) | (0.5) | (0.48) | (0.48) | (0.46) | (0.46) | (0.44) | (0.44) |
| RC              | 0.74 | 1.04 | 0.73 | 1 | 0.76 | 1.02 | 0.85 | 0.94 | 0.80 | 0.86 | 0.85 | 0.86 |

FIG. 1. Time series of (a) daily bandpass-filtered SST ($^\circ$C) and (b) monthly variance of filtered SST ($^\circ$C) for CFSR SST (red), TMI (blue), OISST (green), and TAO SST (black) at $2^\circ$N, $110^\circ$W. Monthly variance is defined as monthly averaged daily variance of each time series.
with those in the TAO ADCP measurements. However, the strength of TIW signals is underestimated, especially in the subsurface. The full TAO time series shows the EKE reached a maximum of 600 cm$^2$ s$^{-2}$ at 55-m depth, about 1.5 times the amplitude estimated by the CFSR (400 cm$^2$ s$^{-2}$). The EKE for the CFSR at 85-m depth has a peak amplitude around 100 cm$^2$ s$^{-2}$ and is about 25% of the TAO estimate.

Horizontal and vertical distribution of EKE reflects the regions where TIWs are prominent. Figure 3 shows the annual mean of the horizontal and vertical distribution of EKE. The horizontal distribution of EKE (Fig. 3a) coincides with that of the observed TIW-associated SST variance (Fig. 4b), with the maximum center located north of the equator. In the vertical, the EKE is largely confined to the upper 100 m, particularly in the mixed layer with the maximum EKE at the surface. The vertical section across 110°W (Fig. 3b) shows the eddy energy extends deeper north of the equator in the region of the South Equatorial Current. Along the equator the eddy energy is largely confined within 130°–110°W in the top 50 m (Fig. 3c). These spatial structures are consistent with previous observations and model simulations (e.g., Weisberg and Colin 1986; Qiao and Weisberg 1998; Masina et al. 1999). The amplitude of the eddy energy, however, is weaker than reported in some previous studies. For example, Baturin and Niiler (1997) computed velocity component covariances from more than 1900 Lagrangian drifters in the tropical Pacific. They found that

the meridional eddy velocity covariance at 15 m has a maximum of over 1250 cm$^2$ s$^{-2}$ at around 2°N, 120°W, giving an estimate for the maximum EKE that would be more than 625 cm$^2$ s$^{-2}$, a much higher value than the CFSR estimate (~220 cm$^2$ s$^{-2}$). Masina et al. (1999) showed that the EKE along the equator at 70-m depth peaks around 210 cm$^2$ s$^{-1}$, which is also higher than the CFSR estimate.

3. TIW characteristics

In this section, the 2001–08 daily data are used to assess spatial structures of ocean–atmosphere variability associated with TIWs in the CFSR. Results are then compared with satellite observations and previous numerical modeling studies. Following Hashizume et al. (2001), all variables were bandpass filtered for a period of 20–40 days and a zonal wavelength of 890–2200 km to highlight TIW-related variability.

a. Horizontal structure

In this analysis temporal variance of filtered SST is used to represent TIW activities. As shown in Figs. 4a,b, the spatial distribution of variance simulated by the CFSR compares well with the TMI estimate, but the amplitude of variance is underestimated. The weaker variance in the CFSR SST is partially attributed to the weaker amplitude of the OI SST, which was assimilated into the CFSR. Both the CFSR and the TMI SST exhibit two
distinct branches of strong TIW activities. The northern branch centered at 2°N is much stronger and broader than the southern one centered at 2°S.

Consistent with previous studies, TIW-related SST variations exhibit pronounced seasonal and interannual variability. The seasonality in the TIW activity can be seen clearly in the SST variance averaged over 0°–4°N (Figs. 4c,d). The maximum variation center moves westward from 90°W in May to 140°W in December, resembling the seasonal evolution of the equatorial cold tongue (Horel 1982). Except for weaker amplitude, similar seasonality is also found in the southern branch (not shown). On interannual time scales, TIW activity varies in concert with the phase of ENSO. As shown in Fig. 5, TIW signals are much stronger and extend farther west during the 2007 La Niña conditions than during the 2002 El Niño conditions. Meanwhile, the CFSR winds are about 70% of the observed. Therefore, we inferred that the underestimation of CFSR SST perturbations largely accounts for the underestimation of wind perturbations in the CFSR.

To study covariability with TIW-induced SST, bandpass-filtered fields are regressed on bandpass-filtered SST at 2°N, 125°W. Figure 6 shows the regression patterns of SST and surface wind from the CFSR, TMI, and QuikSCAT during active TIW seasons (June–January) in 2001–08. The covariability between SST and surface wind in the CFSR is remarkably similar in spatial structures and in magnitude to that from the TMI SST/QuikSCAT wind. Both the CFSR and observations reveal that the mean southeasterly trade winds are enhanced over warm SST and reduced over cold SST, leading to a nearly in-phase relationship between SST and wind speed. The magnitude of the response in wind speed varies from 0.2 to 0.6 m s⁻¹ °C⁻¹.

TIW-induced 10-m wind perturbations give rise to changes in surface wind stresses. The regression pattern of surface wind stress is shown in Fig. 7a. As expected, the pattern of wind stress response is almost identical to that of surface wind (Fig. 6a). The wind stress perturbations are in phase with SST, with stronger (weaker) wind stresses over warm (cold) SST anomalies. We note that the impact of ocean currents on the surface wind stress is absent in the CFSR because the model estimates the winds relative to the fixed ocean surface. This is different from the QuikSCAT wind stresses, which are...
determined by relative motions between the surface winds and the ocean currents (Kelly et al. 2001; Chelton et al. 2004).

The effect of SST on the derivatives of wind stress depends on the alignment of the direction of wind stress and SST gradient (Chelton et al. 2001). As shown in Fig. 7a, pronounced wind stress curl responses take place near the equator, where the winds blow parallel to isotherms and on the eastern flank of northward-pointing cusps (near 5°N). The regression pattern is consistent with previous satellite-based observational studies (Chelton et al. 2001, 2004). In addition, Fig. 7b shows that the centers of upwelling (downwelling) perturbations (vertical velocity at 55-m depth) are nearly collocated with surface current divergence (convergence) at 15-m depth, indicating a physical consistency in the ocean circulation fields is maintained by the CFSR.

The significant responses of wind stress curl suggest that the atmosphere could feed back onto TIW through Ekman dynamics (Chelton et al. 2001). For example, Spall (2007) suggested that the SST-induced wind stress curl perturbations might impact the baroclinic instability of the ocean through Ekman pumping. Using a high-resolution regional coupled model, Seo et al. (2007a) showed that the perturbation of Ekman pumping due to TIW-induced wind stress curl is small compared to the perturbation that resulted from baroclinic instability in the ocean. Similar to Seo et al. (2007a), the magnitude of Ekman pumping vertical velocity from the CFSR (∼3 × 10⁻⁶ m s⁻¹ °C⁻¹) is only one-fifth of the vertical velocity at 55-m depth (∼1.5 × 10⁻⁵ m s⁻¹ °C⁻¹) along 2°N, where upwelling (downwelling) perturbations are strongest (Fig. 7b). The difference in the amplitude suggests that the feedback effect of TIW-induced Ekman pumping variability is a minor contribution to TIW growth compared to the baroclinic energy source intrinsic to the ocean. The results are consistent with Seo et al.’s (2007a) study.

Ocean eddies modify surface wind stresses via two mechanisms: 1) near-surface wind perturbations induced by TIW-associated SST anomalies and 2) ocean currents altering the relative motion between the atmosphere and the ocean (e.g., Chelton et al. 2004). Changes in the surface wind stress, in turn, can feed back on the evolution of ocean eddies. Using a numerical model and satellite observations, Small et al. (2009) investigated the impact of wind stress
feedback on the TIWs. They found that the Ekman pumping due to the surface current effect alone leads to significant damping of the TIWs. Since the impact of surface current on the surface wind stress was not included in the CFSR, it may indicate that the impact of TIW-induced Ekman pumping was underestimated in the CFSR.

The availability of the CFSR data allows inspection of variables that are not directly observed. As the largest

FIG. 5. Longitude–time diagram of filtered SST (shaded, °C) and 10-m WS (contour, m s⁻¹) along 2°N for (a) 2002 and (c) 2007 cold seasons for the CFSR. (b), (d) As in (a) and (c), respectively, but for TMI SST and QuikSCAT wind. The WS interval is 0.3 m s⁻¹, and the zero contours have been omitted for clarity.
Regression values shown in Fig. 6 are confined in the region north of the equator (0°–6°N), this area will be used to discuss TIW-related variability for some other variables in the CFSR. Figure 8 shows the regression coefficients of air temperature at 2 m, specific humidity at 2 m, surface pressure, and the planetary boundary layer (PBL) height.

For 2-m air temperature, the amplitude of TIW-related anomalies is about 0.5 K (Fig. 8a). The largest temperature perturbations are located slightly downstream (i.e., northwest) of the SST extremes due to advection by the mean southeasterly wind flow (Small et al. 2003). The specific humidity maximums (minimums) are closely collocated with the convergence (divergence) of surface winds (Fig. 8b). The PBL height response is nearly in phase with SST, with a deepened (shallower) PBL over warm (cold) SST anomalies (Fig. 8d). The magnitude of extremes reaches 80 m °C⁻¹. It implies that the TIW-associated SST anomalies affect the vertical stratification of the whole PBL. The relationship between the PBL adjustment and SST is consistent with ship-based measurements by Hashizume et al. (2002). Their study demonstrated that the PBL height perturbation could be as much as 500 m in response to a 2°C SST anomaly. It is roughly 3 times the CFSR (160 m).

Sea surface pressure gradients can be decomposed into two terms: the thermal pressure gradient and the interfacial pressure gradient (Hashizume et al. 2002; Cronin et al. 2003). The former is inversely related to near-surface virtual temperature and the latter is associated with PBL height adjustment. Note that virtual temperature is proportional to air temperature and specific humidity. Figure 8c shows that the surface pressure maxima are located between the air temperature and specific humidity minimums, while they are displaced east of the maximum of PBL height. It implies that surface pressure perturbations are mainly influenced by changes in the hydrostatic component. In the CFSR, the magnitude of pressure perturbations reaches ~10 and +8 Pa °C⁻¹.

The amplitude of surface pressure perturbation and its phase relationship with SST perturbation are similar to the analysis based on TAO measurements (Cronin et al. 2003). However, our results disagree with radiosonde observations (Hashizume et al. 2002), which showed little TIW signals in surface pressure. Hashizume et al. (2002) suggested that the interfacial pressure gradient tended to cancel the thermal pressure gradient, leading to the absence of surface pressure response. It is shown in Fig. 8d that the magnitude of TIW-induced PBL height perturbations is about one-third of Hashizume et al.’s (2002) result. It indicates that the interfacial pressure gradient effect might be underestimated in the CFSR, giving rise to overestimation of the thermal pressure gradient effect. In contrast, the discrepancy between the CFSR and the radiosonde observation could be due to different time periods used in Hashizume et al. (2002) and the current study. The former focused on vertical soundings during 21–28 September 1999, while the latter is based on the 2001–08 period. More measurements and modeling studies are needed to understand the differences.

Figure 8c shows the wind perturbations are similar to the ones expected from pressure-driven flow. Using a high-resolution regional atmospheric model, Small et al.

### Table 2. Slopes of linear least squares fit lines of 10-m wind speed (WS) and SST obtained from CFSR, QuikSCAT wind speed, and TMI SST. Values were computed over the regions 150°–110°W along 2°N during active TIW seasons (June–January) from 2001 to 2008.

<table>
<thead>
<tr>
<th>WS (CFSR)/SST (CFSR)</th>
<th>WS (QSCAT)/SST (TMI)</th>
<th>WS (CFSR)/WS (CFSR)/SST (TMI)</th>
<th>SST (CFSR)/SST (TMI)</th>
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<td>0.60</td>
<td>0.66</td>
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**Fig. 6.** Regression coefficients of filtered 10-m WS (contour, m s⁻¹ °C⁻¹), wind velocity (vectors, m s⁻¹ °C⁻¹), and SST (shading, °C) onto filtered SST at 2°N, 125°W during active TIW seasons (June–January) in 2001–08 from (a) CFSR winds and SST, and (b) QuikSCAT winds and TMI SST.
(2003) showed that a westward shift of thermal pressure relative to SST anomalies induces an anomalous wind pattern that is roughly in phase with SST, consistent with the pressure-driven mechanism (Lindzen and Nigam 1987). In contrast, the in-phase relationship between SST and wind speed is also consistent with the vertical mixing hypothesis, which suggests SST perturbation changes the surface wind speed by modifying the stability of the atmospheric boundary and vertical momentum mixing (Hayes et al. 1989). Further studies are needed to understand which mechanism is responsible for SST-induced changes in the low-level wind in the CFSR.

b. Vertical structure

Figure 9 shows the vertical structure of TIW regression coefficients in a longitude–height cross section at 2°N. The climatological-mean PBL height inferred from the CFSR is superimposed on each panel (red line). Air temperature perturbations (Fig. 9a) are mainly confined in the PBL, with a weak reversal in sign above 900 hPa. The pattern is consistent with the analysis by Small et al. (2003). However, Hashizume et al. (2002) reported that large cold temperature anomalies ride directly above the warming effect of SST anomalies based on radiosonde observations. This strong inversion feature is not observed in the CFSR. Geopotential height extremes (Fig. 9b) are out of phase with air temperature extremes and are consistent with the perturbation in surface pressure (Fig. 8c).

The vertical structure of zonal wind velocity $U$ is shown in Fig. 9c. Negative (positive) $U$ anomalies lie slightly to the east of warm (cold) SST. Above 940 hPa, anomalies change sign and reach extremes around the top of the PBL. A west–east dipole structure also exists in the meridional wind velocity regression map (Fig. 9d). The amplitude of SST-induced meridional wind velocity $V$ perturbations is weaker than that of the $U$ anomalies, yet the response can reach to 750 hPa. The vertical structures of the wind field response and their phases relative to SST for the CFSR are generally consistent with previous numerical model studies (Xie et al. 1998; Small et al. 2003).

As shown in Figs. 8 and 9, wind convergence and divergence, along with changes in static stability, set up a thermally direct circulation cell (Fig. 9e, vectors). The perturbations in the vertical velocity (Fig. 10f) are consistent with this paradigm. The resultant vertical circulation also leads to vertical transport in surface moisture and vertical variations in water vapor. The largest specific humidity anomalies are observed around 925 hPa, close to the mean PBL height (Fig. 9e).

Previous studies have suggested that there are strong correlations between TIW-induced SST and low-level clouds. Deser et al. (1993) first noticed enhanced (reduced) cloud anomalies over warm troughs (cold crests) of SST waves. A regression analysis of TMI data showed significant cloud liquid water variations are associated with TIW-induced SST perturbations, indicating the influence of SST wave on cloud (Hashizume et al. 2001).

Following these studies we next examine whether CFSR is capable of capturing the covariance between low-level cloud and SSTs. Figure 10a shows the regression coefficient pattern of the low-level cloud cover (LLCC) in the CFSR. The magnitude of LLCC perturbations reaches 5% °C$^{-1}$, and the phase is consistent with the perturbation of low-level wind convergence. The amplitude is close to that from the satellite observation (4.2% °C$^{-1}$) by Deser et al. (1993). The LLCC perturbations also show a propagating feature that is consistent with the SST anomalies and reinforces the confidence in the statistical significance of low-level cloud response in the CFSR (Fig. 10b). It is noted that the
maximums of the model LLCC perturbations are roughly in quadrature to Deser’s results. Two mechanisms have been proposed to explain how low clouds respond to SST anomalies. Deser et al. (1993) suggested that formation of low cloud can be attributed to thermal convection associated with cold-air advection from the equator to the warm sector. Xie et al. (1998) proposed that perturbation of wind convergence favors cloud formation by increasing humidity near the PBL top. Our analysis shows the extremes of wind convergence (divergence) perturbations coincide with maximums (minimums) of cloud anomalies as well as water vapor anomalies (Figs. 10a and 8b), lending support to the convergence mechanism.

c. Impacts of TIW-induced atmospheric responses on the surface heat budget

In the previous sections, we documented that TIW-induced SST perturbations give rise to significant response in atmospheric wind, air temperature, water vapor, and cloud fields—factors that play important roles in determining surface heat fluxes. In this subsection, we will quantify how these variables impact surface flux variability and compare our estimates with previous observation and model studies.

The responses of ocean surface heat flux quantities are shown in Figs. 11 and 12. Positive coefficients mean the ocean gains heat. The perturbations of net heat flux (Fig. 11a), which are defined as the sum of shortwave radiation (SW), longwave radiation (LW), latent heat flux, and sensible heat flux, are almost out of phase with SST perturbations, suggesting a negative feedback. The peak net heat flux responses to SST perturbations are about +40 and −40 W m⁻² °C⁻¹.

The perturbations of latent heat flux (Fig. 11b) and sensible heat flux (Fig. 11e) are also out of phase with SST perturbations. The magnitude of latent heat flux responses is about 30 W m⁻² °C⁻¹. In contrast, the sensible heat flux is 4 W m⁻² °C⁻¹, SW is 8 W m⁻² °C⁻¹, and LW is 4 W m⁻² °C⁻¹. It suggests that changes in latent heat flux are the dominant factor accounting for the net heat flux perturbations. The magnitude of the latent heat perturbations in the CFSR is somewhat lower than previous estimates of 50 W m⁻² °C⁻¹ by Zhang and McPhaden (1995) from in situ data and 40 W m⁻² °C⁻¹ by Thum et al. (2002), estimated based on a combination of satellite and in situ data.

Following the bulk parameterization approach (Fairall et al. 1996), latent heat flux can be written as $\text{LH} = \rho C_e L (\Delta q)$ and sensible heat flux as $\text{SH} = \rho C_p C_h U (\Delta T)$, where $\rho$, $U$, $C_p$, and $L$ are air density, 10-m wind, specific heat of air, and the latent heat of vaporization, respectively; $C_e$ and $C_h$ are the exchange coefficients for moisture and heat, respectively; $\Delta q = q_s - q_a$, where $q_s$ and $q_a$ are the saturation specific humidity at the SST and the near-surface air specific humidity, respectively; and $\Delta T$ is the difference between SST and near-surface air temperature. To the first order, the perturbations of latent heat flux can be written as

$$\text{LH}' = \rho C_e L (\langle U \rangle) (\Delta q') + \rho C_e L (\langle \Delta q \rangle) U'$$

where primes denote bandpass-filtered fields and angle brackets denote mean values. Thus, both changes in
humidity difference and wind speed can contribute to latent heat flux anomalies. As shown in Fig. 11c, the peak of SST-induced water vapor perturbations generates about a 25 W m$^{-2}$ °C$^{-1}$ latent heat flux response, which is about 2.5 times the contribution due to wind speed perturbation (~10 W m$^{-2}$ °C$^{-1}$) (Fig. 11d). This suggests that the TIW-associated latent heat flux perturbations are mainly attributed to TIW-associated changes in water vapor. Similar analysis suggests that sensible heat fluxes associated with TIW are mainly affected by SST perturbations directly (Fig. 11f).

TIW-associated variability of low-level cloud cover also has an important impact on the surface heat budget by affecting the radiation flux. Figure 12 shows the regression coefficient maps of each component of the radiation heat flux at the ocean surface. The extremes of SW flux are located east of SST centers by 4° and follow a similar pattern to that of the low-level cloud cover regression map (Fig. 10a). The magnitude of the shortwave flux is about 9 W m$^{-2}$ °C$^{-1}$. Our analysis suggests that warm SST perturbations induce increased low-level cloud cover, hence reducing the amount of solar radiation reaching the sea surface (downward SW, Fig. 12c), which in turn tends to cool the warm SST anomaly (negative feedback). Assuming a typical SST wave amplitude of 2.5°C (Deser et al. 1993), the change in shortwave radiation flux will be 22.5 W m$^{-2}$, similar to the estimate (25 W m$^{-2}$) by Deser et al. (1993). The regression coefficients of longwave radiation fluxes and their phases relative to SST and cloud are shown in Figs. 12b,d,f, respectively. It is inferred that the upward and downward longwave flux perturbations are associated with SST and cloud cover perturbations, respectively. The magnitude of total longwave radiation fluxes at the surface is about 4 W m$^{-2}$ °C$^{-1}$.

4. Summary and discussion

The NCEP CFSR represents a new reanalysis effort with the first guess from a high-resolution coupled system, thus offering prospects for improved simulation of TIWs. The purpose of this analysis is to investigate and validate the ocean–atmosphere covariability arising in the presence of the tropical Pacific TIWs in the CFSR. We used multiyear daily high-resolution CFSR data to describe the temporal and spatial features of ocean and
atmosphere variability associated with TIWs. Consistent with previous studies, TIW-induced SST variations exhibit pronounced seasonal and interannual variability that are tightly connected with cold tongue variations.

Similar to the TMI SST/QuikSCAT winds, an in-phase relationship between SST and surface wind speed was well represented in the CFSR. Both datasets reveal that surface wind response is on the order of 0.6 m s$^{-1}$ °C$^{-1}$. Comparisons with the TAO observations reveal that the temporal variations of SST and wind anomalies associated with TIWs are accurately replicated in the CFSR, but the magnitudes are underestimated by about 25% and 20%, respectively. The weaker SST perturbations might be partially attributed to the weaker variance in the daily OISST that is assimilated in the CFSR. The coupling strength in the CFSR, measured by the wind response per unit SST anomaly, is about 10% weaker in the CFSR than its satellite counterpart.

The TIW-associated SST perturbations also affect air temperature and water vapor variability, which appear to give rise to surface pressure perturbations. The extremes of surface pressure are located downstream of SST extremes. The magnitude of pressure perturbations reaches $-10$ and $+8$ Pa °C$^{-1}$. The amplitude and phase of sea surface pressure to SST are similar to the analysis from TAO measurements (Cronin et al. 2003). The expected pressure-driven flow was in phase with SST, similar to the regression pattern of surface wind in the CFSR. It suggests that the pressure-driven mechanism (Lindzen and Nigam 1987) plays an important role in leading the surface wind response to SST, consistent with findings by Small et al. (2003) from a high-resolution atmospheric regional model, although the vertical mixing mechanism (Hayes et al. 1989) may also have a contributing factor. Detailed examinations of physical processes are beyond the scope of this study, but the availability of the CFSR data will help clarify the mechanisms by which the wind responds to SST in future studies.

The analysis also indicates that surface wind convergence and pressure fields, together with changes in static stability set up a thermally direct circulation cell. The resultant vertical motion causes large variations in water vapor. Clear TIW-associated signals are also found in low-level cloud cover. The cloud cover response is about 5% °C$^{-1}$. The amplitude is close to that from the satellite observation (4.2% °C$^{-1}$) by Deser et al. (1993). Our analysis suggests that TIW-associated perturbations of wind convergence favor cloud formation by increasing moisture near the PBL top, which is in line with the convergence mechanism proposed by Xie et al. (1998).

CFSR allows for a detailed investigation of thermodynamic feedback of surface heat fluxes associated with TIWs. Both latent and sensible heat fluxes are approximately out of phase with SST perturbations, and the responses are about 30 and 4 W m$^{-2}$ °C$^{-1}$, respectively. Further analysis shows that TIW-induced water vapor perturbation is the primary factor contributing to changes in latent heat flux, while SST-induced wind perturbation plays a secondary role.

The net surface heat flux is further modified by the formation of low-level clouds over warm water. The low-level clouds reduces net shortwave radiation at the surface by about 9 W m$^{-2}$ °C$^{-1}$, slightly smaller than the estimate by Deser et al. (1993) (10 W m$^{-2}$ °C$^{-1}$). The total net surface heat flux response is about 40 W m$^{-2}$ °C$^{-1}$, and is dominated by the change in latent heat flux. Assuming that the 40 W m$^{-2}$ net heat flux is distributed over a mixed-layer depth of 40 m, the resulting SST tendency is 0.02 K day$^{-1}$. For a typical TIW periodicity of 30 days, the negative anomaly in net heat flux would produce 0.6°C cooling over warm waters. It suggests that TIW-associated heat flux perturbation yields a negative thermodynamic feedback to the TIW-associated SST by reducing (increasing) the net surface heat flux over warmer (colder) SST, consistent with previous studies.
We note that some of the comparison between the CFSR and the observational data may not be an independent validation. For example, the comparison of winds between the CFSR and TAO/QuikSCAT observations is not a rigorous way to validate CFSR surface winds, since these observations were assimilated in the CFSR. Instead, the comparisons should be interpreted as the degree to which the CFSR assimilation either maintains or corrupts the observational information. One way to assess the impact of data assimilation would be to conduct a similar analysis using a free run of the model or to conduct observing simulation experiments wherein specific observational platforms are not included. This, however, is beyond the scope of the current study.

Although the CFSR is able to provide a consistent representation of the ocean–atmosphere system—surface fluxes, low-level cloud cover, etc.—some deficiencies in the CFSR were also identified. For variables that are not constrained by the assimilation of observations, differences between the CFSR and observations can often be substantial. For example, the oceanic TIW EKE is underestimated substantially in the CFSR compared with the TAO ADCP and previous studies. The weak EKE could be related to the systematic bias in the mean state in the CFSR. The bias in the ocean temperature in the CFSR, relative to the TAO data, switched from a weak cold bias to a strong warm bias suddenly around the end of 1998, and has been diagnosed to be related to the assimilation of the Advanced Television and Infrared Observation Satellite Operational Vertical Sounder (ATOVS) data (Xue et al. 2011; Zhang et al. 2012). The onset of warm bias in the equatorial eastern Pacific weakens the SST front in the cold tongue, which probably explains the weakened TIW activities in the CFSR, consistent with the recent study by Wen (2009). Using a regional coupled model, Wen (2009) showed that when a large subsurface warming exists near the equatorial Atlantic Ocean, the activity of TIWs is substantially reduced, owing to the weakened SST front (Yu et al. 1995).

Large uncertainty is also found in the surface pressure signature associated with the TIW. The amplitude and the phase of sea surface pressure to TIW-associated SST perturbation are similar to the analysis from the TAO measurement (Cronin et al. 2003), while they disagree

![FIG. 11. Regression coefficients of (a) net surface heat flux (shaded, W m⁻² °C⁻¹), (b) latent heat flux (shaded, W m⁻² °C⁻¹), (c) latent heat flux response due to Δq' (shaded, W m⁻² °C⁻¹), and (d) latent heat flux response due to u' (shaded, W m⁻² °C⁻¹). Positive values mean the ocean gains heat from the atmosphere. Regression coefficients of SST (contours at +0.4°C and −0.4°C) are overlaid in each plot.](image-url)
with radiosonde observations (Hashizume et al. 2002). The discrepancy between the CFSR and the radiosonde could be attributed to the different time periods analyzed. The former is based on the 2001–08 CFSR record, while the latter focused on a short period possibly covering three TIW waves only. In contrast, the discrepancy might imply some vertical structures were not well represented in the CFSR. The radiosonde showed that the PBL is capped by a strong temperature inversion. However, this feature was not observed in the CFSR. It implies that the thermal pressure gradient effect may be overestimated in the CFSR and hence the surface pressure response.

Numerical modeling studies have demonstrated that TIWs play an important role in the equatorial heat budget by modulating horizontal advection and vertical mixing that involve air–sea heat and momentum flux exchanges. Recent studies have also suggested TIWs could contribute to the asymmetry of El Niño and La Niña (with larger amplitude for El Niños) and the irregularity of ENSO (An 2008; Zhang and Busalacchi 2008; Imada and Kimoto 2012). By examining TAO data and satellite observations, Jochum et al. (2007a) showed that zonal advection by TIWs is a significant contributor to the mixed-layer heat budget. However, a comprehensive understanding of the influence of TIW on climate variability is lacking, as some key variables are not available in observations. Our results demonstrate the capability of the CFSR in capturing mesoscale surface variations associated with TIWs, which are traditionally poorly represented in previous reanalysis. The high resolution and long record (1982–2009) of the CFSR, and its real-time extension, will provide an unprecedented opportunity to monitor and to characterize the space–time behavior of TIWs, and to help advance our understanding of the role of TIWs in the climate system. Finally, since small-scale ocean–atmospheric interactions were commonly observed near SST fronts throughout the global ocean (Chelton et al. 2004; Xie 2004), the 0.5° resolution of the CFSR might be sufficient to resolve oceanic mesoscale eddies with spatial scales greater than 100 km. Similar analyses of small-scale coupled variability outside the tropical Pacific basin may also be explored using the CFSR data.
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