Coupled Response of the Trade Wind, SST Gradient, and SST in the Caribbean Sea, and the Potential Impact on Loop Current’s Interannual Variability

Y.-L. Chang
National Taiwan Normal University, Taipei, Taiwan, and Princeton University, Princeton, New Jersey

L.-Y. Oey
Princeton University, Princeton, New Jersey, and National Central University, Jhongli, Taiwan

(Manuscript received 23 September 2012, in final form 7 February 2013)

ABSTRACT

Air–sea coupling in the IntraAmerican seas (IAS; Caribbean Sea and Gulf of Mexico) is studied through analyses of observational data from satellite, reanalysis products, and in situ measurements. A strong coupling is found between the easterly trade wind \(-U\) and meridional SST gradient \(\delta T/\delta y\) across a localized region of the southern-central Caribbean Sea from seasonal and interannual to decadal time scales. The \(\delta T/\delta y\) anomaly is caused by a variation in the strength of coastal upwelling off the Venezuelan coast by the wind, which in turn strengthens (weakens) for stronger (weaker) \(\delta T/\delta y\). Wind speeds and seasonal fluctuations in IAS have increased in the past two decades with a transition near 1994 coinciding approximately with when the Atlantic multidecadal oscillation (AMO) turned from cold to warm phases. In particular, the seasonal swing from summer’s strong to fall’s weak trade wind has become larger. The ocean’s upper-layer depth has also deepened, by as much as 50% on average in the eastern Gulf of Mexico. These conditions favor the shedding of eddies from the Loop Current, making it more likely to shed at a biannual frequency, as has been observed from altimetry data.

1. Introduction

In the Caribbean Sea (8°–20°N, 90°–60°W), the lower-tropospheric easterly trade wind \(-U\) reaches maxima twice per year, once in July and another time in January. Following the maxima, the easterly trade wind weakens to minima, once in October and another time around May. This biannual variation was found by Wang (2007) based on the monthly-mean National Centers for Environmental Prediction (NCEP) reanalysis data (2.5° × 2.5° resolution) at 925 hPa in the Caribbean Low-Level Jet region (CLLJ; defined as minus the zonal wind averaged over 12.5°–17.5°N, 80°–70°W; Amador 1998). The biannual variation was also found by Munoz et al. (2008) based on the 40-yr European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-40) dataset (1.125° × 1.125° resolution). The authors suggest that the variation is caused by changes in the northward pressure gradient across the Caribbean Sea because of the seasonal shift of the North Atlantic Ocean subtropical high (NASH), and that the pressure gradient is modulated, both in winter and summer, by the meridional thermal gradients across the southern Caribbean Sea (SCarS) because of land heating to the south. Cook and Vizy (2010) used a higher-resolution North American Regional Reanalysis dataset (approximately 32-km resolution) and found that heating over the land topography of northern South America is effective only during the South American monsoon season in boreal winter, but that in summer, the “westward expansion of the NASH tightens the meridional geopotential gradient over the Caribbean to produce the July wind speed maximum.” Wang (2007) found that the biannual variation of the CLLJ is asymmetric: the summer–fall weakening from 12 to 6 m s\(^{-1}\) is larger than the winter–spring weakening.
The biannual asymmetry was noted by Chang and Oey (2012) using a higher-resolution but shorter-period surface wind dataset (0.25° × 0.25° from 1988 to 2009) that incorporates also buoy and satellite measurements, and the phenomenon is prevalent throughout the Caribbean Sea. The meridional gradient $\partial T/\partial y$ of the sea surface temperature (SST) across the CLLJ region also varies biannually, with maximum gradients in August and January and minimum gradients in November and May, approximately in synchrony with the wind (Wang 2007). On the other hand, the SST itself varies annually with a single maximum (minimum) around October (February). Based on these observations and the suggestion by Inoue et al. (2002), Wang (2007) hypothesizes that the wind stress curl–induced Ekman pumping (suction) to the south (north) of the trade wind core would strengthen $\partial T/\partial y$, and the resulting sea level pressure gradient further strengthens the easterly wind (Lindzen and Nigam 1987). Wang (2007) suggested that the positive coupling may lead to the biannual variation of CLLJ. A lucid analysis of the atmospheric boundary layer process (e.g. Pedlosky 1987) associated with an SST gradient is given by Feliks et al. (2004), who show that an easterly (westerly) anomaly is produced for positive (negative) $\partial T/\partial y$ [see supplemental materials (SM)]. One expects that the positive $\partial T/\partial y$ would be predominantly contributed by the southern cooling from Ekman pumping because, in the absence of heat sources (and/or mean upwelling), Ekman suction alone in the northern half cannot increase SST. As a consequence, one also expects that the $\partial T/\partial y$ would be most sensitive to the southern cooling and warming anomalies because of stronger and weaker upwelling, respectively.

The biannual asymmetry in the easterly trade wind may in part be caused by the occurrence of the so-called mid-summer drought (MSD). As described in Magana et al. (1999), MSD results in a decrease in the amount of rain around July–August over southern Mexico and Central America and parts of the southwestern Caribbean. After the May and June rainy season caused by the northward shift of the ITCZ, the onset of MSD in July is associated with decreased convective activity (i.e., a cool anomaly source) over the eastern Pacific Ocean warm pool off the western coast of southern Mexico and Central America. According to the model of Gill (1980), a lower-tropospheric anticyclonic circulation anomaly develops centered to the northwest of the cool source over the western coast of Mexico (this location will be referred to as the MSD center), and it contributes to the strengthening of the easterly trade wind in the Caribbean Sea in July. With the anticyclonic anomaly over the warm pool, there are fewer clouds and more incoming solar radiation. In the following months, the warm pool SST rises above the threshold of around 28°C for increased low-level convergence and deep convection. This signals the termination of MSD near the end of August and early September and the reestablishment of the lower-tropospheric cyclonic circulation anomaly, which contributes to the weakening of the easterly trade wind in the Caribbean Sea in September and October. Both MSD (Magan et al. 1999) and air–sea coupling (Wang 2007) mechanisms may contribute to a strong and weak easterly trade wind in summer and early fall, respectively, making the max–min difference from summer to fall much more distinct than from winter to spring and explaining the asymmetry of the biannual wind variation.

Variability of the easterly trade wind is also associated with the tropospheric vertical wind shear $U_z$, which is one of several factors that influences tropical cyclogenesis. Inoue et al. (2002) find that in the Caribbean Basin, the number of tropical storms formed reaches maxima in June and October, separated by a relative minimum in July, which the authors suggest is associated with the onset of MSD when the easterly trade wind is strong. The strong trade wind is accompanied by higher $U_z$ as well as increased moisture flux divergence, preventing the organization of deep convection as well as suppressing it, thus also suppressing tropical cyclogenesis. The authors suggest that during the MSD, to the south of the axis of strong easterly wind, Ekman pumping cools SSTs in the southwestern Caribbean Sea (10°–15°N, 80°–75°W; this will be referred to as the Columbia Basin). In time, the cool SST results in decreased atmospheric convection, leading to weakened wind speeds (Wallace et al. 1989), decreased upwelling, and signaling a reversal of the environmental conditions that again favor cyclogenesis in September and October.

The coupling between $\partial T/\partial y$ and the easterly trade wind, and in turn the trade wind’s connection with vertical wind shear, operate also at interannual and longer time scales. Figure 1 shows these relations using the NCEP reanalysis data, from 1948 to 2010, and various correlations (Corr) with their 95% significances are printed (see also Table 1).\(^1\) Averaged September–October values over the CLLJ region are used, because in this region and during these months the weakening of the easterly trade wind encapsulates many of the coupled processes described above. At long time scales, the variations

\(^1\) Corr($A, B, lags$) is the maximum lagged correlation coefficient with lags in months and is positive if $A$ leads $B$. All quoted correlations are above the 95% significance level, and lags are omitted for zero-lagged correlation.
are related to the Atlantic Multidecadal Oscillation (AMO) (Schlesinger and Ramankutty 1994), and the AMO is also shown. Negative AMO indicates cooler SST in the northern Atlantic Ocean, approximately from 1900 to 1930 (not shown), and from 1963 to 1994, while positive phase indicates warmer SST, approximately from 1930 to 1963, and from 1994 to present. The trade wind (anomaly, same below) is seen to be correlated with $\partial T / \partial y$, $\partial P / \partial y$, and $\mathbf{U}$ (anomalies). Periods of stronger trade wind and SST gradients [hence also stronger atmospheric boundary-layer sea level pressure (SLP = $P$) gradients] approximately coincide, so does the stronger $\mathbf{U}$; Corr($-U, \partial T / \partial y$), Corr($-U, \partial P / \partial y$), and Corr($-U, \mathbf{U}$) are all positive. The trade wind (as well as the other three variables) are anticorrelated with AMO, so that they are generally weaker after 1994 when AMO > 0, and stronger from 1963 to 1994 when AMO < 0. These general trends are consistent with the findings of Wang (2007, Fig. 9b) that the trade wind is anticorrelated with the Caribbean Sea SSTs. There are notable exceptions, however, which may be related to ENSO events such as the negative (positive) trade wind anomaly from 1988 to 1990 (from 2001 to 2002) during La Niña (El Niño) (see Fig. 1a). Wang (2007) suggests that in summer, ENSO’s teleconnection to the Caribbean Sea SLP is such that the trade wind weakens (strengthens) during the period of cooler (warmer) tropical Pacific SST anomalies. Figure 1 also indicates that Corr($\partial T / \partial y$, AMO) is insignificant (or significant but weak according to the unsmoothed series; Table 1), which may reflect the fact that the $\partial T / \partial y$ cannot be adequately resolved with the NCEP 2.5° × 2.5° grid.

The above studies indicate therefore that the easterly trade wind in the Caribbean Sea affects and is affected by processes not only in the atmosphere and at the air–sea interface, but also in the ocean’s subsurface through Ekman pumping. The wind also seems to hold the key to our understanding of mesoscale eddy dynamics in the IntraAmerican seas (IAS; Caribbean Sea and Gulf of Mexico). Oey et al. (2003b) show in model calculations that the periods of Loop Current eddy shedding depend on Yucatan transport fluctuations driven by the Caribbean wind stress and wind stress curl and suggest that localized wind stress curl can spin up warm core rings in the northern Caribbean Sea. Chang and Oey (2012) propose that the biannual transport variation in the Yucatan Channel [observed by Rousset and Beal (2010)] is forced by the biannual easterly wind in the Caribbean Sea, and show that this contributes [the other contributor is the closely related wind variation in the Gulf of Mexico, as explained also in Chang and Oey (2012)] to biannual preferences in the months when eddies are shed (late summer and winter). Jouanno et al. (2012) suggest that the biannual easterly trade wind also forces the biannually varying southern Caribbean Current, which in turn becomes baroclinically and barotropically unstable, producing Caribbean eddies with a biannual preference (August–October and February–March). Chang and Oey (2013) found that the Loop Current’s growth

![Figure 1](image-url)
cycle is intimately related to the vorticity and transport fluctuations in the Yucatan Channel because of the piling up and retreat of warm water in the northwestern Caribbean Sea by the biannually varying trade wind.

In this work, we use observational data to further explore trade wind variability and to infer its connection with ocean processes at time scales from seasonal to interannual, emphasizing more on the latter, longer periods. Expanding the ideas of previous studies, we will demonstrate that trade wind in the IAS is intimately coupled to the SST gradient and SST over the SCarS. This regional air–sea coupling is particularly strong from summer to fall, is related to the MSD process of the eastern-equatorial North Pacific, and impacts the western portion of NASH. We will show that with the warming of the North Atlantic Ocean in the recent decades, mesoscale variability in the Loop Current appears to have undergone a basic characteristic change that is driven by the air–sea coupling. Section 2 describes the data used in this study. Section 3 uses time series and singular value decomposition (SVD) (Bretherton et al. 1992) to analyze the air–sea coupled processes. Section 4 presents observational evidences that the Loop Current’s interannual variability is related to the coupled wind, SST, and SST gradient process in the Caribbean Sea. Section 5 concludes the paper.

2. Datasets and methods

Daily SST data at 1/8° resolution from 1982 to 2009 are from the Group for High Resolution Sea Surface Temperature (GHRSSST; https://www.ghrsst.org/data/). Ocean surface winds from 1988 to 2009 are from Cross-Calibrated Multi-Platform (CCMP) Ocean Surface Wind at 6-hourly and 1/8° spatial resolution. NCEP reanalysis data at a coarser resolution (2.5°) but longer period covering from 1948 to 2010 are also used. Whenever possible, we will use the higher-resolution GHRSSST and CCMP datasets to analyze the air–sea coupling process. The buoy SST and wind speed from the National Data Buoy Center (NDBC) (http://www.ndbc.noaa.gov/) are also used in some analyses. The Loop Current data from 1974 to 1992 were obtained from the literature; they were compiled by previous investigators using a combination of satellite–SST images as well as in situ and ship measurements to identify eddy separations and eddy-shedding periods and to infer Loop Current variability (Sturges et al. 1994; Vukovich 1998; Sturges and Leben 2000; Leben 2005). Although these data were already used and interpreted in various ways in these previous works, the earliest data from the 1970s and early 1980s may be more prone to inaccuracies. Data in the late 1980s are believed to be more reliable as Advanced Very High Resolution Radiometer (AVHRR) and other satellite products became more widely available. After 1992, gridded satellite altimetry data is available from Archiving, Validation, and Interpretation of Satellite Oceanographic data (AVISO; http://www.aviso.oceanobs.com/). More accurate mean sea surface height (e.g., Rio et al. 2009) as well as model and observation reanalysis data specifically designed to study the Loop Current (Oey et al. 2003a; Oey and Wang 2009) are also available. Weekly gridded AVISO products as well as an extension of the analyses of Oey et al. (2003a) and Oey and Wang (2009) through 2010 are used in this study to infer the dates when eddies were shed. Finally, to diagnose long-period changes in the ocean’s upper-layer depths, monthly ECMWF ocean analysis data (Balmaseda et al. 2008; http://apdrc.soest.hawaii.edu/datadoc/ecmwf_oras3.php) at 1° × 1° from 1959 to 2009 are used.

All data are organized into a monthly format (unless they already come in as monthly data). Monthly climatologies are calculated based on the full record period. To study interannual variations, anomalies are then computed by subtracting the monthly climatologies from the corresponding dataset. The 95% significances given for various correlations are computed as 1 − (1 − 0.95)2(F − 1), where F is the degree of freedom calculated as N/τ_N, N is length of time series and τ_N is the dot product of the autocovariances of the two time series.

| −U in CLLJ | 0.36 (−0.11) | 0.48 (0.1) | 0.70 (0.1) | −0.36 (−0.11) |
| T | 0.48 (0.1) | 1 | 0.52 (0.11) | 0.52 (0.11) |
| P | 0.92 (0.13) | 0.52 (0.11) | 1 | 0.29 (0.1) |
| U | 0.70 (0.12) | 0.29 (0.1) | 0.78 (0.14) | 0.82 (0.14) |
| AMO | −0.36 (−0.11) | −0.25 (−0.14) | −0.44 (−0.19) | −0.5 (−0.2) | 1 |
3. Shift in characteristics of IAS winds with warming of Atlantic SST

Figure 1 suggests a shift in the trade wind and other related characteristics as AMO transitions from negative to positive around 1994. To study the differences before and after the transition, we pick two 5-yr periods: Pr1 from 1989 to 1993 and Pr2 from 2004 to 2008, when AMO is in the cold and warm phases respectively, and when also we have contemporaneous high-resolution wind (CCMP: 1988–2009) and SST (GHRSSST: 1982–2009) datasets. We caution, however, that, because AMO has a typical period of about 50–70 years and the cold-to-warm phase transition around 1994 is gradual, the two chosen periods may not, therefore, be representative of the air–sea state during cold and warm phases, respectively. Nonetheless, Pr1 and Pr2 are characterized respectively by anomalously stronger and weaker trade wind, SLP gradients, and $U_w$, which are apparently related to AMO (Fig. 1), and they will also be seen to represent two contrasting periods of Loop Current eddy-shedding characteristics. Other ancillary evidence will be provided below to indicate that the two chosen periods display characteristics that are consistent with cold or warm sea surface states.

Figure 2 shows wind anomaly composites superimposed on SST anomalies averaged for the months of September–October for the respective periods. During Pr2 (Pr1), anomalously warm (cold) SST is seen in the SCarS off the Central and South American coast. The wind anomaly is westerly (easterly), thus weakening (strengthening) the trade wind. The westerly (easterly) wind anomaly extends into the eastern-tropical Pacific Ocean where the SST is generally warmer (cooler) during Pr2 (Pr1), and where also a cyclonic (anticyclonic) wind anomaly is centered near the MSD center over the western coast of Mexico ($22^\circ$N, $105^\circ$W). While the easterly anomalies strengthen the trade wind in the Atlantic and Caribbean, they weaken the generally southwesterly (convergent) winds off the coast of Panama and Mexico in the ITCZ of the eastern Pacific (ca. $5^\circ$–$15^\circ$N and $90^\circ$–$110^\circ$W) (Hastenrath 2002). Incorporating the ideas presented by Magana et al. (1999), we deduce from Fig. 2b (Fig. 2a) that, during Pr2 (Pr1), the late summer rainy season in September–October after MSD would produce anomalously more (less) precipitation over the southern part of Mexico and Central America, which is in fact observed (see SM). The confinement of warming (cooling) along the SCarS coast (Fig. 2) indicates weakened (strengthened) coastal upwelling and means that $\partial T/\partial y$ is decreased (increased) during Pr2 (Pr1). Both seasonal and interannual surface heat fluxes from NCEP reanalysis show greater losses during Pr2 (than Pr1; Fig. 1e) despite the weakened wind, and therefore cannot contribute to the southern Caribbean warming (cooling) observed in Fig. 2 during Pr2 (Pr1). Comparing with AMO (Fig. 1e, thick line), years of cold (warm) AMO generally coincide with, but slightly lead by 2–3 years, periods of negative (positive) anomalies of latent heat flux indicating decreased (increased) heat losses from the ocean. The coastal warming also indicates that, as mentioned previously, the SST gradients are predominantly contributed by the SST changes over the southern portion of the Caribbean Sea. However, the SST cooling and warming during Pr1 and Pr2 are asymmetric to the west and east of approximately $70^\circ$W longitude: the cooling during Pr1 is strongest in the Columbian Basin in southwestern Caribbean Sea while the warming during Pr2 is largest in the eastern end of the Caribbean Sea. The asymmetry suggests a departure of simple upwelling and downwelling along an idealized straight coast. The west-side cooling asymmetry may be caused by the stronger upwelling (in addition to coastal upwelling) in the Columbian Basin because of Ekman pumping by the positive wind stress curl (Gordon 1967) when the easterly trade wind is stronger in September–October during Pr1. On the other hand, as the trade wind weakens during Pr2, a setup of isotherms by cyclonically propagating Kelvin and other coastally trapped waves would result in the anomalous convergence of warmer water near the Lesser Antilles island chain near $60^\circ$W, producing the east-side warming. These SST asymmetries result also in asymmetric wind anomaly patterns seen in Fig. 2. According to Gill (1980), the cyclonic (anticyclonic) wind anomaly is located to the northwest of the warm (cold) source. The anticyclone’s center (Fig. 2a) is over the Campeche Bay in the southwestern Gulf of Mexico ($21^\circ$N, $95^\circ$W) close to the MSD center. Since it is of opposite sign than the cyclonic trough that characterizes the second rainy season in September–October after the MSD (Magana et al. 1999), it acts to weaken the trough, and hence also the precipitation as mentioned above. During Pr2, the cyclone’s center (Fig. 2b) is to the east and northeast of

---

2 Pr1 may include influences from the strong 1988/89 La Niña. To test this, we repeated the composite and other related analyses taking 1990–94 as the cold-phase period. The results are very similar (details in SM).
Cuba at 21°N, 73°W, separate from the MSD center, and the wind anomaly patterns therefore consist of two cyclones (Fig. 2b).

To assess the sensitivity of the composites (Fig. 2) to longer periods of cold and warm phases of AMO, we average for 13 years from the beginning year of GHRSST, 1982, through 1994 to represent the cold phase, and another 13 years from 1996 through 2008 to represent the warm phase. For the wind, we use the NCEP reanalysis data. The results indicate similar features as in Fig. 2: warm (cool) SST and westerly or weakening (easterly or strengthening) trade wind anomalies in the Caribbean during the warm (cold) phase of AMO from 1996 to 2008 (from 1982 to 1994); despite the coarser resolution of the NCEP wind, cyclonic (anticyclonic) flow centered to the north of Cuba (over the southwestern Gulf of Mexico) still exists during the warm (cold) phase.

The wind and SST anomalies in the SCarS are also linked to wind and SST responses in other parts of the North Pacific and Atlantic Oceans (Fig. 2c). The Pr2
(Pr1) is generally associated with cooler (warmer) eastern-central tropical–subtropical Pacific SSTs, cooler (warmer) Pacific decadal oscillation (PDO) in the extratropics of the Pacific Ocean, and weaker (stronger) NASH in the extratropics of the North Atlantic, consistent with features detailed in Wang (2007) (see SM).

Monthly SST and $\partial T/\partial y$ from GHRSST and easterly trade wind from CCMP averaged in the CLLJ region are shown in Figs. 3a,b for Pr1 and Pr2. The SST variation is annual with a minimum from February through March and a maximum in October, although there is a small (statistically insignificant) peak in June during Pr2. The SST during Pr2 is from 0.4° (January) to 0.8°C (October) warmer than Pr1. In contrast, the $\partial T/\partial y$ is biannual with maxima in both January–February and July–August and minima in May and around October–November (Fig. 3b). The SST gradients are generally weaker during Pr2 especially from summer to fall. The easterly trade wind is also biannual (Fig. 3a), with maxima from June (Pr2) through July (Pr1) and in January, and minima in May and October. There is a larger-amplitude seasonal fluctuation during Pr2 than Pr1, especially from the summer maximum to fall minimum when it is 50% larger (than from winter to spring), similar to the corresponding variation in $\partial T/\partial y$. Wind fluctuations in IAS have indeed become stronger in the past decades. Figure 3c plots the trade wind (i.e., $-U$) averaged in CLLJ. There were six events in the past 21 years in which the CLLJ wind speeds were weaker than the mean minus two standard deviations (i.e., $-U < 4.6 \text{ m s}^{-1}$: 1990, 1999, 2004, 2005, 2007, and 2008), all of them were around September–October, and four out of six occurred during Pr2. The standard deviation of the wind stress also shows an increasing trend throughout the 1988–2009 period (Fig. 3c, dash curve). The increase after 1994 coincides with the gradual change in AMO from cool to warm phases (see Fig. 1). Figures 3d,e compare Pr1 and Pr2 wind stresses and the corresponding standard deviations. While the Caribbean Sea trade winds are weaker in September–October during Pr2 (Fig. 2), the 5-yr wind means and standard deviations (Figs. 3d,e) are stronger, in particular because of stronger wind in summer and winter (Fig. 3a). The increases are large in the SCarS, and there are smaller increases over the eastern Gulf of Mexico. Figures 3f,g plot wind speed and standard deviation anomalies recorded at NDBC station 42003 (26.04°N, 85.61°W; the longest available wind measurements over the ocean in IAS) in the eastern Gulf of Mexico. The trends in the past ca. 20–30 years (Figs. 3c,f,g) are for both increasing wind speeds and stronger fluctuations with a transition from generally negative to positive anomalies near 1994.

The larger standard deviation after 1994 has a strong seasonal component contributed in particular by the anomalous weakening of the easterly trade wind from summer to fall (Fig. 3a), which uniquely distinguishes Pr2 from Pr1. To examine the spatial structure of the anomaly, Fig. 4a compares between Pr1 and Pr2 the difference between fall and summer zonal wind stresses, $\delta \tau_{z}$ (where $\tau$ is kinematic zonal wind stress), which is always positive in the Caribbean Sea and is larger in years of weaker fall and/or stronger summer trade winds. In the Gulf of Mexico, the seasonal zonal wind is 180° out of phase with the Caribbean wind (Chang and Oey 2012), and the $\delta \tau_{z}$ is negative and stronger (i.e., more negative) in years of stronger fall and/or weaker summer easterly winds (Fig. 4a). It is clear that $|\delta \tau_{z}|$ is larger during Pr2 than Pr1 over the Caribbean Sea and in the eastern Gulf of Mexico. Figure 4b plots the corresponding changes between fall and summer in the wind stress curl $\delta (V \times \tau_{z})$ during Pr1 and Pr2. In the northern (southern) Caribbean Sea the changes are positive (negative) because the wind stress curls in the respective regions weaken from summer to fall. The change is again larger in Pr2 than Pr1. From the south to the north across the Caribbean Sea, the normal condition is for a negative $\delta (V \times \tau_{z})$ in September or October, as well as stronger values of wind stress curl in June or July, and therefore a large negative (positive) $\delta (V \times \tau_{z})$, resulting in weakened Ekman-induced upwelling (downwelling) in the southern (northern) Caribbean Sea during fall. The situation is reversed during Pr1. The westerly wind anomaly over the Caribbean Sea during Pr2 weakens the trade wind, and the resulting weakening of coastal upwelling (downwelling) along the southern (northern) coast contributes to weaker $\partial T/\partial y$. In Fig. 2b, the westerly wind anomaly in the Caribbean Sea is convergent over the warm SST anomalies (SSTA; $\Delta T > 0.5^\circ$C), which favors upward convection and the cyclonic anomalous flow centered to the east and northeast of Cuba, as mentioned previously.

In the Gulf of Mexico, $\delta (V \times \tau_{z})$ is generally positive indicating the normal weakening of the Gulf-wide anticyclonic wind stress curl from summer to fall (Fig. 4b) (Gutierrez de Velasco and Winant 1996). Complex anomaly patterns are seen during Pr2, with positive and negative values over the region that extends northwestward from the Yucatan Channel to northerncentral Gulf of Mexico. These extended patterns appear to reflect stronger fluctuating winds over a warmer eastern Gulf (Fig. 2b; Figs. 3c,f,g) during Pr2 when, as will be
FIG. 3. (a) CLLJ-averaged composite Pr1 (black) and Pr2 (red) monthly easterly wind (CCMP; dashed line; mean = 7.4 m s$^{-1}$) and SST anomalies (GHRSST; solid line; mean = 27.8°C). (b) $\partial T/\partial y$ [°C (250 km)$^{-1}$] for Pr1 (black) and Pr2 (red). (c) CLLJ-averaged zonal wind (solid line; lhs y scale; m s$^{-1}$), thin horizontal line has mean of 7.4 m s$^{-1}$, dotted horizontal lines are std dev of 1.4 m s$^{-1}$, and thick dashed is the 360-day low-pass wind stress std dev (rhs y axis scale; m$^2$ s$^{-2}$). (d)-(e) Pr1 and Pr2 kinematic wind stress (vectors; reference vector provided in m$^2$ s$^{-2}$) and std dev (color shading; 10$^{-4}$ m$^2$ s$^{-2}$) based on the full 5-yr data. (f)-(g) NDBC station 42003 (26.04°N, 86.61°W) wind speed and std dev anomalies (m s$^{-1}$).
shown in section 4, the Loop Current has become deeper and eddies are more abundant.

The anomalous, summer-to-fall weakening of the trade winds during Pr2 is closely related to the ocean. Figure 4c shows that the SST anomaly (SSTA; color shading) averaged over the SCarS (10°–17°N, 85°–60°W) is significantly correlated with the $\delta T_{s}^{*}$ anomaly (open bars). The year when SCarS warms tends to coincide with when the September or October wind of the same year weakens anomalously. The exceptions are 1990 ($\delta T_{s}^{*}$ anomaly > 0 but SSTA < 0) and 1995 and 1998 ($\delta T_{s}^{*}$ anomaly < 0 but SSTA > 0), which reflect the influence of ENSO and the Atlantic warm pool (AWP) on the atmospheric conditions over the Caribbean Sea (Gray 1984; Wang et al. 2006). The extra-deep cumulus convection in the eastern Pacific during an El Niño is associated with anomalous westerly in the upper-tropospheric wind (200 hPa; Gray 1984). The associated Hadley cell produces anomalous easterly wind near the surface over the Caribbean Sea. The surface trade winds therefore tend to be stronger around September–October of 1995 and 1998 following the December 1994 and 1997 El Niño [consistent also with the findings of Wang (2007)], resulting in negative $\delta T_{s}^{*}$ anomaly despite the positive AWP in both years. The situation is reversed after La Niña. Although 1989 is near the end of the 1988/89 La Niña, its effect as well as the influence of the positive AWP in 1990 both contribute to the positive $\delta T_{s}^{*}$ anomaly (Fig. 4c).

a. Air–sea coupling

The $T_{s}/\delta y$ supports the SLP gradient and correlates with the trade wind. A weakened trade wind, which may be associated with the termination of MSD, reduces the supply of cooler water to the surface along the SCarS coast and reduces the SST gradients, which further weaken the wind, hence the SST gradients. While not necessary in the argument, as isotherms flatten, stratification increases and the mixed layer thins, which traps heat more uniformly near the surface and also reduces $T_{s}/\delta y$. The upshot is a positive coupled cycle in which trade wind weakens, coastal upwelling is reduced, sea surface warms, and the SST gradient decreases, which leads to further wind weakening, SST warming, etc. The coupling is more effective during Pr2 in the warm phase of AMO because of the corresponding decreases in vertical wind shear (Fig. 1d; Lee et al. 2011) and static stability of the atmosphere. Decreased stability is evidenced in satellite and NCEP data, which indicate that during Pr2, precipitation and latent heat flux loss (Fig. 1e) increase, while both shortwave and longwave radiation are less (see also SM). Latent heat flux dominates the surface heat flux, consistent with previous studies in the tropical North Atlantic Ocean (e.g., Foltz and McPhaden 2006). The associated increased mixing and thicker atmospheric boundary layer (ABL) can be shown to lead to stronger SST gradient–induced wind responses (appendix A).

Air–sea coupling is indicated by the near symmetry of the lagged-correlation coefficients of the same sign for plus or minus lags between $-U$ and $\delta T_{s}/\delta y$ (Fig. 5), indicating reinforcement of the atmosphere by ocean and vice versa (Chang et al. 1997). The SST gradients and $-U$ are significantly correlated, with a maximum of 0.49 at 95% significance level when the wind leads by 1 month. As discussed previously in Fig. 1 and Table 1, they are also significantly correlated with $\delta T_{s}/\delta y$ and $U$.

Instead of NCEP reanalysis, high-resolution (1/4° × 1/4°) CCMP and GHRSST datasets are used to conduct coupling analysis using SVD (Bretherton et al. 1992; Wang et al. 2003). Although the period is shorter (1988–2009), it includes the recent AMO transition from cold to warm phases near 1994. The SVD applied to two fields seeks to identify pairs of coupled spatial patterns (modes 1, 2, etc.), with each pair explaining a fraction of the covariance between the two fields. Figure 6 shows the SVD coupled spatial patterns (SPs) and expansion coefficients (ECs) between $\delta T_{s}/\delta y$ and zonal wind $U$ in the Caribbean Sea (8°–17°N, 55°–90°W). The SPs are plotted as homogeneous correlation maps (correlations of EC’s with data fields $\delta T_{s}/\delta y$ and $U$, respectively) whose squares ($\times 100$) give the percentage of variance explained. Analyses based on both the monthly time series (Figs. 6a–d) and time series with seasonal cycle removed (Figs. 6e–h) are shown; the similarity between their SPs indicates similar coupling processes. The total series (Fig. 6c) displays strong seasonal fluctuations but with substantial interannual variations the fraction of which [computed by projecting the interannual EC1 (Fig. 6g) onto the total series] is about 50%. The first mode is dominant with a square covariance fraction SCF1 of 58% (48% for interannual); SCF2 and higher are each <10% and are not shown. The $SP1(\delta T_{s}/\delta y)$ is concentrated in the upwelling region...
along the northern Venezuelan coast (ca. 75°–62°W), and has two local maxima to the west and east of about 70°W (Figs. 6a,e), reflecting the anomalous upwelling and downwelling asymmetries noted previously in Fig. 2, and a local minimum in the Columbian Basin. The SP1(∂T/∂y) couples with an SP1(U)-field that is negative and that has a maximum magnitude slightly to the west (80°–72°W; Figs. 6b,f). The EC1(U) and EC1(∂T/∂y) are significantly correlated (at the 95% significance level; Figs. 6d,h) and are nearly symmetrical about the zero lag where the correlation coefficient is 0.78 (0.68 for interannual), up to ±1-month lags. The negative lag indicates ∂T/∂y leading, and shows feedback from the ocean to the atmospheric boundary layer pressure.

Fig. 4. (a) Zonal wind stress \( \tau_{z} \) \( \delta \tau_{z} \) = \( S_{t} (\text{Max}_{t} - \text{Min}_{t}) \), where \( \text{Max}_{t} \) and \( \text{Min}_{t} \) are from June to October, and \( S_{t} \) is the sign of fall (September or October) minus summer (June, July, or August) \( \tau_{z} \). The first white contour \( |\delta \tau_{z}| \) is 8 \( \times 10^{-2} \) m² s⁻² and the contour interval (CI) is 2 \( \times 10^{-2} \) m² s⁻². (b) Wind stress curl \( \delta(\nabla \times \tau_{z}) \) (m s⁻²) during (left) Pr1 and (right) Pr2. (c) Anomalies of \( \delta \tau_{z} \) (m s⁻¹) in CLLJ region and SST (with seasonal cycle removed; °C) in the southern Caribbean Sea (color shading; ca. 10°–17°N, 85°–60°W). Correlation of \( \delta \tau_{z} \) with SST (0.54) and 95% significance (0.37) are shown.
gradient in the manner discussed previously. The positive lag (wind leads) indicates a wind-driven ocean response (i.e., upwelling and/or downwelling because of both the presence of the coast and contribution from wind stress curl) that drives the $\partial T/\partial y$ and the significant lag extends further to longer than +1 month, suggesting ocean responses at longer time scales. The positive correlation means that along the northern Venezuelan coast, since the $\text{SP1}(\partial T/\partial y)$ and $\text{SP1}(U)$ there are generally of opposite signs (Figs. 6a,b; Figs. 6c,f), periods of strongest trade winds ($-U$) and upwelling coincide. The notable exception is in the Columbian Basin where both $\text{SP1}(\partial T/\partial y)$ and $\text{SP1}(U)$ are of the same sign, so that periods of strong trade winds and weak $\partial T/\partial y$ in the basin coincide. These results are inconsistent with the air–sea coupling idea (e.g., Inoue et al. 2002; Wang 2007) caused by SST anomalies in the Columbian Basin that are driven by wind stress curl–induced Ekman pumping (suction), which presumably would couple positive $\partial T/\partial y$ with the easterly trade wind. Instead, the SVD results suggest the following physical interpretations. The strong trade wind forces a coastal upwelling jet (along the northern Venezuelan coast), which advects cool water into the northern region of the Columbian Basin, thus negating any (positive) $\partial T/\partial y$ that might

Fig. 5. Lag-correlation coefficients between 90-days low-passed $-U$ in CLLJ region and SST gradient (12.5°–15°N, 80°–70°W) using the NCEP reanalysis data (1948–2010) showing near symmetry about the max $\approx 0.5$ (when $-U$ leads by 1 month); dotted line indicates values below the 95% significance.

Fig. 6. SVD spatial patterns $\text{SP1}$ (a,e) $\partial T/\partial y$ from GHRSSST with CI 0.2 and (b,f) zonal wind $U$ from CCMP with CI 0.05; (c,g) expansion coefficients EC1s; (d,h) Corr[EC1($\partial T/\partial y$),EC1($U$),lag], which is positive if $U$ leads, and solid if above 95% significance. The columns are based on (a–d) monthly time series and (e–h) on series with 22-yr (1988–2009) monthly composite removed.
have been otherwise produced by the positive wind stress curl in the basin. It is coastal upwelling, not the wind stress curl–induced Ekman pumping, that is coupled with the overlying wind via the offshore SST gradient ($\partial T/\partial y$) that it supports. Everything else (i.e., wind stress and Coriolis parameter) being equal, divergence and convergence caused by coastal constraint is much more effective than wind stress curl in forcing vertical movements of isotherms, by a factor of approximately $L_{CB}/R_c \approx 5–10$, where $L_{CB}$ is the width of the Columbian Basin (or scale of the wind curl) and $R_c$ is the baroclinic Rossby radius that defines the coastal upwelling zone (e.g., Allen 1980).

The wind anomaly pattern caused by the coupling is shown in Fig. 7, which plots the correlation between the September–October wind and $-EC1(\partial T/\partial y)$ from SVD. The 22-yr (1988–2009) monthly composite of both the wind and $\partial T/\partial y$ have been removed. Vectors are plotted only if the correlations are above the 95% significance. Shaded region is where the SVD is computed.

Figure 6c (or Fig. 6g) indicates that the EC1 amplitudes are larger (smaller) after (before) 1994. Figures 8a,b plot these EC1 amplitudes and the 22-yr composite monthly values. The monthly plot (Fig. 8b) shows a significant biannual cycle for both the wind and $\partial T/\partial y$, with amplitudes that are largest from summer (ca. June–July) to early fall (October) (cf. Fig. 3). The weakening of the wind and $\partial T/\partial y$ are centered on August, which seems to be consistent with the notion above that coupling during wind weakening may be triggered by the termination of MSD. By similar reasoning, the early onset of MSD may initiate the coupling process when the trade wind and $\partial T/\partial y$ are strengthening in June. Figure 8a shows that the EC1 amplitudes have increased for the past two decades. The transition from small to large amplitudes appears to be near 1994 especially for $\partial T/\partial y$, but a robust increasing trend for both is found after ca. 2000–01.

Finally, the SVD analysis using wind and SST (not shown) (instead of $\partial T/\partial y$) indicates that warmer SST covaries with weakened trade wind, but the patterns tend to be broad over the entire Caribbean Sea, and the correlation is weak (a maximum of 0.38). Taken together with the more localized region of stronger correlation for the trade wind and $\partial T/\partial y$ off the Venezuelan coast, these results suggest a general weakening of the trade wind in the warm phase, but the coastal upwelling process is crucial in explaining the physical coupling between the trade wind and ocean. The wind–SST analysis also does not show that the Columbian Basin’s cooling covaries with weakened trade wind as conjectured by Inoue et al. (2002), suggesting the dominance of the SST gradient–induced wind response (Lindzen and Nigam 1987; Feliks et al. 2004) over the vertical shear adjustment mechanism (Wallace et al. 1989).

b. Effects of NASH and South American continental land heating

Previous works (Wang 2007; Munoz et al. 2008; Cook and Vizy 2010) have suggested that the CLLJ variability may be part of the gyre-scale response forced by the shifting and varying intensity of the NASH. Also, in summer and fall, the strong coupled response discussed above may be influenced by thermal gradients caused by heating over the South American continent (Munoz et al. 2008). These issues are discussed in appendix B in which we show that influences from NASH and summer
land heating are weak. Our analysis tends to support the suggestion of Cook and Vizy (2010) that, in summer, continental land heating does not play a significant role.

4. Changes in Loop Current behavior in a warming climate

Do changes in the characteristics of the wind forcing as the northern Atlantic SST transitions from cold to warm phases, described above, alter also Loop Current variability in the Gulf of Mexico? Recent process-modeling studies demonstrate a strong dependence of Loop Current eddy shedding to wind forcing and suggest that such a possibility may indeed exist (Chang and Oey 2010, 2012). Before we can answer the question, some background information is necessary.

Chang and Oey (2012) found that biannual variation in the trade wind forces a corresponding biannual transport.
through the Yucatan Channel. The Loop Current has a tendency to shed eddies as the wind weakens from summer to fall, and also from winter to spring. The process depends also on the wind within the Gulf of Mexico (Chang and Oey 2010). Figure 9a shows the calendar months when eddy shedding is recorded, from 1974 to 2010. There are 24 eddy-shedding events in 1974–94 and 23 in 1995–2010, with periods (interval between 2 events) of ca. 4–19 months. The periods appear to be shorter after 1994, and the transition near 1995 coincides with the phase change of the AMO (Fig. 9b). However, this apparent connection of the shedding periods with AMO, while interesting, is tentative as the earliest data from the 1970s and 1980s may be prone to inaccuracies. Data in the late 1980s and thereafter are more robust, and these show the preferences for biannual shedding: in July–October (red dots) and January–March (blue dots). The biannual shedding characteristics are also seen in Fig. 9c in terms of the so called seasonal histogram, in which the number of eddies are plotted as a function of the calendar months (Chang and Oey 2012). The pre-1989 dataset (black bars) shows no biannual signal. On the other hand, the histogram based on the dataset after 1989 (gray bars and line connected by open circles) shows a clear biannual signal, with two local minima: November–December (no eddy) and May (1 eddy), and two maxima: September and March.

The shedding of Loop Current eddies can be interpreted as resulting from a balance between the volume influx \( Q \) through the Yucatan Channel, which grows the Loop, and the westward eddy's migration [velocity \( C_i = -2\beta R_o^2/3 \) for a lens eddy (Nof 1981), where \( R_o \) is the Rossby radius based on the matured eddy], which tends to peel the eddy from the Loop (Pichevin and Nof 1997; Nof 2005). The model results from Chang and Oey (2012) show that the Yucatan Channel transport is negatively correlated with the zonal wind stress and wind stress curl in the Caribbean Sea (see their Figs. 2d,e). From their 21-yr monthly climatology (see their Fig. 2a), we obtain the empirical estimate

\[
\frac{\partial Q}{\partial t_{sf}} \approx -1.2 \text{ Sv}/(10^{-5} \text{ m}^2 \text{ s}^{-2}),
\]

where \( sf \) is from summer to fall, that is, a decrease of 1.2 Sv (1 Sv = \( 10^6 \text{ m}^3 \text{ s}^{-1} \)) per 0.01 N m\(^{-2}\) decrease in the easterly trade wind stress (\( \delta \tau^e > 0 \)). From the composite of the Yucatan transport from Rousset and Beal (2010) based on 3.5-yr observations (which also show biannual variation), we obtain \( \frac{\partial Q}{\partial t_{sfObs}} \approx -3.5 \text{ Sv}/(2.5 \times 10^{-5} \text{ m}^2 \text{ s}^{-2}) \) (here the \( \delta \tau^e \) is from observed summer–fall CCMP wind stress composite; Chang and Oey 2012) \( \approx -1.4 \text{ Sv}/(10^{-5} \text{ m}^2 \text{ s}^{-2}) \), which agrees well with (1). As the wind weakens, the \( Q \) also decreases, which then results in the Pichevin and Nof imbalance that makes the Loop Current more susceptible to eddy shedding. In the interpretation of Chang and Oey (2012), the Loop Current behaves as a forced and weakly chaotic system. The time periods when eddies are shed (\( \pm 1–2 \) months) are mostly determined by the wind forcing, while the exact timing (\( \pm 1–2 \) weeks) is determined by other more rapidly evolving factors, most notably the baroclinic instability and upper–lower layer mass coupling (Xu et al. 2013).

We can now attempt to address the question posed at the beginning of this section. We focus on Pr1 and Pr2 when the eddy-shedding data is more reliable, and when as mentioned previously the high-resolution CCMP and GHRST are also available. The two periods cover two distinct phases (Fig. 9b): cold phase for Pr1 and warm phase for Pr2. Yet, the two periods have quite distinct eddy-shedding characteristics (Fig. 9a): 5 eddies during Pr1 and 9 eddies during Pr2, making them particularly interesting and significant to analyze.\(^6\) The summer-to-fall

\(^6\) From Fig. 9a, the Pr1 may also be chosen from 1990 to 1994. The results are very similar.
shedding occurs in both periods, while winter-to-spring shedding also occurred during Pr2. Our aim is to explain these differences based on previous modeling works (Chang and Oey 2010, 2012) and the unique differences in the warm and cold phases.

The decreased transport as a result of summer–fall weakening of the wind stress during Pr1 and Pr2 can be estimated from (1) and the observed wind [either Fig. 3a using the wind drag formula from Large and Pond (1981), or directly from Fig. 4a]:

\[
\Delta Q|_{Pr1ws} \approx -6 \text{ Sv} \quad \text{and} \quad \Delta Q|_{Pr2ws} \approx -10 \text{ Sv}.
\]

These estimates show that as the trade wind weakens from summer to fall, the supply of warm water into the Loop weakens by an amount that in both periods is a significant percentage of the total volume flux of approximately 12 Sv carried by a Loop Current eddy (Chang and Oey 2010). Process modeling (Chang and Oey (2012), see also their online auxiliary materials; also Chang and Oey (2013)) and realistic simulations (Xu et al. 2013) indicate that the value \( \Delta Q \approx -6 \text{ Sv} \) is a typically large value when the Loop Current can become more susceptible to shedding an eddy; this explains why summer-to-fall eddy shedding can occur during both Pr1 and Pr2.\(^7\)

A similar calculation for weakening of the wind stress from winter to spring yields the following:

\[
\frac{\partial Q}{\partial \tau^x}|_{ws} \approx -0.6 \text{ Sv}/(10^{-5} \text{ m}^2 \text{ s}^{-2}),
\]

where \( ws \) is from winter to spring, so that, since the weakening of the trade wind from winter to spring is also approximately half of the value from summer to fall, the corresponding decreased transports are

\[
\Delta Q|_{Pr1ws} \approx -1.5 \text{ Sv} \quad \text{and} \quad \Delta Q|_{Pr2ws} \approx -2.5 \text{ Sv}.
\]

These values are small (four times weaker than the values from summer to fall). The weakening of the easterly trade wind alone is therefore unlikely to be the only cause for the increased number of eddies shed from winter to spring during Pr2. Since the Pichevin and Nof imbalance depends on both the size of the matured Loop Current eddy that is about to shed (i.e., \( R_o \)) as well as on the inflow (i.e., \( Q \)), we conjecture that the upper layer (the Loop Current) during Pr2 is deeper in the eastern Gulf of Mexico. This would make \( C_i \approx -2\beta R_o^2/3 \) larger during Pr2, enabling the eddy to be more easily peeled off of the Loop Current.

There are at least two ways in which the upper layer of the Loop Current may become deeper during Pr2: (i) warmer SSTs and increased ocean heat content in the near-surface waters of the Caribbean Sea, which supply the Loop Current; and (ii) increased convergence of the surface layer of the eastern Gulf of Mexico by increased westward wind stress within the Gulf (Chang and Oey 2010). One may quickly discount (ii). By comparing Figs. 2a,b, the increased westward wind stress in Pr2 relative to Pr1 produces at most an Ekman transport of \( \Delta x/\tau^x L \approx 0.03 \text{ m}^2 \text{ s}^{-1} \times 10^6 \text{ m} \) (where \( L = 10^6 \text{ km} \) is the half-width of the Gulf and \( f \) is the Coriolis parameter) \( \approx 0.03 \text{ Sv} \), which is returned eastward along the midlatitudes of the Gulf (Chang and Oey 2010). This is too weak to have a significant effect on the near-surface convergence over the Loop Current. To examine (i), we use monthly ECMWF ocean analysis data (Balmaseda et al. 2008) at \( 1^\circ \times 1^\circ \) from 1959 to 2009. This coarse-resolution global analysis cannot be used to infer eddy shedding. The dataset is useful, however, for the present purpose of inferring large-scale changes in the background fields over the long period. The SSH \( \eta \) is used but other diagnostic variables such as, for example, the depth of the 20°C isotherm give the same conclusions. The SSH difference (Pr2 – Pr1) is all positive in the IAS (Fig. 10b); the value in the eastern Gulf of Mexico is \( \approx 0.05 \text{ m} \) (or an approximately 50 m increase in the upper-layer depth) compared with the averaged SSH value of about 0.1 m in the eastern Gulf of Mexico during Pr1 (Fig. 10a). The ratio of the westward migration speeds of the matured eddy \( C_{Pr2}/C_{Pr1} \approx (\eta_{Pr1} + \Delta \eta)/\eta_{Pr1} \approx 1.5 \) a 50% increase, which is substantial. This increased eddy westward migration speed and the larger drop in \( \Delta Q|_{Pr2ws} \) during Pr2 relative to Pr1 (4) are consistent with the increased shedding frequency during Pr2. Indeed, the eastern Gulf of Mexico SSH has steadily increased since 1994 (Fig. 10c), in general accompanying the steady increase also in AMO. Finally, Fig. 10d shows that the eastern Gulf’s SSH averaged over the winter months from December through the following February has also been steadily increasing since 1994. Such an increase during winter is consistent with the large Loop Current deepening that is a prerequisite for eddy shedding from winter to spring seen in Pr2. In both cases, the SSHA are significantly correlated with AMO, maximum when the AMO leads by 3 years.

\(^7\) Besides \( Q \), the vorticity of the western portion of the inflowing Yucatan Current also controls the extension and retraction of the Loop Current (Chang and Oey 2012). The two covary, however, and to simplify the presentation we lump the vorticity effect into \( Q \).
5. Summary

In this paper, we have used observational data to analyze air–sea coupling between the trade wind, SST, and SST gradients in the Caribbean Sea, at seasonal and longer (interannual and decadal) time scales, with emphasis on the latter. In particular, we focus on the conditions associated with the AMO transition from cold to warm phases around 1994 and also analyze the impact of this warming climate on the Loop Current eddy variability in the Gulf of Mexico. Our conclusions are as follows.

1) There is coupling between the easterly trade wind (i.e., $-U$) and meridional SST gradient (i.e., $\partial T/\partial y$) across a localized region of the southern-central Caribbean Sea at the seasonal, interannual, and decadal time scales.

FIG. 10. (a) SSH (m) averaged over Pr1 (1989–93) from ECMWF Ocean Analysis (ORA-S3); (b) SSH difference (m) is equal to Pr2 (2004–08) minus Pr1 (dashed CI are 1, 2, 3, 4, and $5 \times 10^{-2}$; solid CI are $6 \times 10^{-2}$ and larger); (c) SSH anomaly (SSHA; m) averaged over the eastern Gulf of Mexico ($23^\circ$–$28^\circ$N, $90^\circ$–$82^\circ$W) from 1959 to 2009, with the 51-yr annual cycle ("steric") removed; (d) eastern Gulf of Mexico SSHA (m) averaged from December of each year through February of the following year. In (c),(d) the AMO (black line) has been shifted to the right by 3 years (i.e., the SSHA lags AMO). Also, correlations and 95% significances are printed. The first set is for the entire 51 years, second set is for 1975–2009.
2) The easterly trade wind also correlates with SST, such that strong trade wind is associated with cool SST and vice versa. However, the coupling is over a larger region covering the entire Caribbean Sea and is much weaker than that between $-U$ and $\partial T/\partial y$. This suggests that the SST–wind coupling is indirect via $\partial T/\partial y$; from summer to fall, in particular, the $\partial T/\partial y$–wind coupling is strong, which may be because of the large-scale warming (cooling) preconditions the atmospheric boundary layer by decreasing (increasing) the static stability for effective (ineffective) coupling of the wind with $\partial T/\partial y$.

3) Contrary to previous presumptions, the strong $\partial T/\partial y$ that is coupled to the trade wind is not caused by wind stress curl–induced Ekman dynamics to the south and north of the easterly wind core; the SVD analysis in fact shows a negative coupling between $\partial T/\partial y$ and the trade wind in the Columbian Basin where strong wind stress curl exists. Instead, the coupled $\partial T/\partial y$ is confined off the Venezuelan coast in the southern-central Caribbean Sea and varies through strengthening and weakening of coastal upwelling.

4) The North Atlantic Oscillation (NAO) or position and intensity of NASH have weak influences on the interannual changes of the Caribbean trade wind, in particular since 1980 when the correlation between the NASH and the trade wind appears to have precipitously degraded.

5) Wind speeds and fluctuations in IAS have increased in the past two decades with a transition near 1994 when AMO shifts from cold to warm phases; the wind stress standard deviation has increased by roughly 40%–50%. In particular, the drop in the trade wind from summer to fall has become larger in the warm climate because of the more effective, positive coupling between the wind and coastal upwelling (i.e., $\partial T/\partial y$), which produces along the Venezuelan coast localized cold and warm anomalies for strong and weak trade winds, respectively. A regional southern Caribbean SST index (Fig. 4c) is found to track the amplitude of weakening of the trade wind from summer to fall. A similar, larger drop during the warm AMO is also observed from winter to spring, though the difference between cold and warm AMO is not as distinct as for the case from summer to fall.

6) These drops in the wind stresses are estimated to produce decreases in the Yucatan transport, which are approximately 60% larger during the warm AMO; in particular, the decrease from summer to fall in the transport is estimated to be as much as $10\text{Sv}$, which makes up nearly the whole ($\sim 80\%$) of the typical westward transport of eddy that is shed from the Loop Current.

7) The upper-thermocline depth in IAS has increased in the past two decades in apparent response to warming associated with the AMO; in particular, in the eastern Gulf of Mexico, the averaged increase is about 50 m, or as much as 50%.

8) During the warm phase of the AMO, the drops in Yucatan transports, from summer to fall and from winter to spring, coupled with the increase in the upper-layer depth, produce favorable conditions for the Loop Current to shed eddies, consistent with the record that eddy shedding has become increasingly biannual and more frequent.

In addition to transport, Oey et al. (2003b) have suggested that eddy shedding can be influenced by perturbations coming through the Yucatan Channel from the Caribbean Sea, the energetic state of which may also be strongly biannual dependent on the wind especially from summer to fall (Jouanno et al. 2012; Chang and Oey 2013). This possibility may be a worthwhile topic for further study. We also found that the easterly trade wind ($-U$; also $\partial T/\partial y$) correlates with the tropospheric vertical wind shear $U_s$. Low $U_s$ are believed to be necessary for tropical cyclogenesis (DeMaria 1996), which suggests that the period of warm AMO with a stronger drop in the trade wind from summer to fall may generally coincide with the period of more frequent tropical cyclogenesis in the Caribbean Sea. It is interesting that the same environmental conditions of anomalously weakened trade wind from summer to fall, and of warm AMO that preconditions a more unstable atmosphere, may all contribute to increased activity of mesoscale “storms” in both the ocean (Loop Current warm eddies) and the atmosphere (tropical cyclones). On the other hand, while the role of southern Caribbean Sea coastal upwelling in producing the coupled response between the trade wind and meridional SST gradient is clear, the relevant dynamics as well as connection with the large-scale atmospheric and oceanic circulation remain to be explored. A study of the connection of the proposed coupled process in SCARSS with eastern-tropical Pacific air–sea conditions that give rise to MSD, for example, and/or the effects of coupling on (and interaction with) NASH, will provide improved understanding of possibly interrelated mechanisms. As to the latter, Rodwell and Hoskins (2001) have shown how the Asian monsoon heating may extend eastward as both poleward Sverdrup flow and equatorial Kelvin wave similar to the eastern portion of the solution from Gill (1980), as far east as the North Atlantic. The influence of $\partial T/\partial y$ over the SCARSS may be similar, though of a much smaller amplitude, which nonetheless may affect the NASH. The resolution of present-day climate models [e.g., the Geophysical
Fluid Dynamics Laboratory (GFDL) Earth System Model (ESM2) with 2.5° × 2° atmosphere and 1° × 1° up to 5/8° at equator; see Dunne et al. (2012)] is still inadequate to resolve the finescale coupling of coastal upwelling and trade wind described herein. It is also too coarse to resolve the Loop Current eddy-shedding process, which typically would require a grid resolution of approximately 10 km (e.g., Oey et al. 2005). Further research using a regional high-resolution coupled model, together with observations, is necessary.

**Acknowledgments.** Support from the U.S. Bureau of Offshore Energy Management; and the Taiwan Foundation for the Advancement of Outstanding Scholarship, the National Science Council, the Ministry of Education, and the National Central University are gratefully acknowledged.

**APPENDIX A**

**Dependence of SST Gradient–Induced Winds on ABL Thickness and Mixing**

The modified model from Feliks et al. (2004) (see SM) model gives the (nondimensionalized) wind amplitude as

\[
A_U = \left(\frac{|\Delta T^*|}{\theta_o} d^{-1}\right) \left(\frac{e}{Fr}\right) \left(1 + k + k^2/2\right) \left(1 + k^2/4\right) k,
\]

where \(\Delta T^*\) is SST change (K) across the (nondimensionalized) distance \(d\), \(\theta_o^{-1} (=1/300 \text{ K}^{-1})\) is the thermal expansion coefficient, \(e\) is the Rossby number, \(Fr\) is the Froude number that varies as inverse atmospheric boundary layer (ABL) height \(\delta_E^{-1}\), and \(k > 0\) is a parameter that measures how deep the ocean influence (i.e., SST) penetrates into the troposphere, such that large \(k^{-1}\) models more ABL mixing. In addition to being dependent on SST gradient, that is, \([|\Delta T^*|/\theta_o]d^{-1}\) in (A1), wind amplitude is stronger for thicker (larger \(\delta_E\)) and more mixed (larger \(k^{-1}\)) ABL, as well as in lower latitudes (larger \(e\) and \(\delta_E\)).

**APPENDIX B**

**Effects of NASH and South American Continental Land Heating**

Figure 2c shows that, from 0° to 30°N, the most conspicuous cooling (Pr1) and warming (Pr2) is in the southern Caribbean Sea. We examine if such a localized feature may be forced solely by NASH. We first note that near the NASH center (30°–40°N, 40°–20°W), the correlation map of Wang (2007) between CLLJ and SLP gives weak or insignificant values for Corr(CLLJ,SLP). We next take the NAO index as a measure of NASH variations (e.g., Wang 2007), and compare the regression coefficients (Regr) between the monthly zonal wind in the CLLJ region (i.e., where \(U = -\text{CLLJ}\) with NAO and \(U\) with ECI(−\(\partial T/\partial y\)). Table B1 summarizes the results. The correlation between \(U\) and ECI(−\(\partial T/\partial y\)) is significant, as discussed previously in Figs. 6 and 7. The corresponding regression shows a change of approximately 0.5–0.74 m s\(^{-1}\) per standard fluctuation of \(\partial T/\partial y\); the trade wind weakens or strengthens with the SST gradient. In contrast, the correlation between \(U\) and NAO is weak and insignificant. Lastly, we use NCEP data from 1948 to 2010 to compute the 10-yr running correlation between \(\partial P/\partial y\) in the CLLJ region and the SLP averaged within 25°–45°N, 48°–18°W surrounding the NASH center. The correlation is significant and quite high (average \(\approx 0.5\)) for the early period from 1960 to 1980. However, consistent with Table B1, the correlation becomes insignificant after 1980. These results suggest that the influence of NASH on the Caribbean winds and SST is weak, and that the observed pressure gradient changes in the Caribbean are likely to be regionally controlled by \(\partial T/\partial y\) as we have detailed in the main text.

To understand the potential effects of the land heating on the proposed coupled process, NCEP reanalysis data are used to examine the coupled response of land and sea with wind using SVD (Fig. 8c), and this is then compared with Fig. 8b: the sea-only SVD analysis using GHRSST and CCMP. For the sea-only analysis, the

<table>
<thead>
<tr>
<th>ECI(−(\partial T/\partial y))</th>
<th>NAO</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Corr</strong></td>
<td><strong>Regr</strong></td>
</tr>
<tr>
<td>Whole Sep–Oct</td>
<td>Jan–Feb</td>
</tr>
<tr>
<td>0.58</td>
<td>0.61</td>
</tr>
</tbody>
</table>

Table B1. Corr and Regr coefficients between zonal wind averaged in the CLLJ region (12.5°–17.5°N, 80°–70°W) with \(U = -\text{CLLJ}\) and ECI(−\(\partial T/\partial y\)) from SVD, and \(U\) and the NAO index. The wind is based on CCMP from 1988 to 2009. Italicized entries are below (i.e., not significant) the 95% significances. The regression has a unit of m s\(^{-1}\) per one std dev of either ECI(−\(\partial T/\partial y\)) or NAO. Positive Regr(\(U\), ECI(−\(\partial T/\partial y\))) means weakening (strengthening) of the easterly trade wind per weakening (strengthening) of the SST-gradient—that is, changes in \(U\) and \(\partial T/\partial y\) are of opposite signs. On the other hand, negative Regr(\(U\), NAO) means strengthening (weakening) of the easterly trade wind per strengthening (weakening) of the NAO—that is, changes in \(U\) and NAO are also of opposite signs. Note that positive Regr(\(U\), NAO) as for September–October is physically implausible based on our intuitive understanding of the effect of NAO (NASH) on the Caribbean trade wind (see SM).
coupled response is such that both $EC1(\partial T/\partial y)$ and $EC1(U)$ are biannual as discussed before. The zero-lag correlation between monthly (not composite) $EC1(\partial T/\partial y)$ and $EC1(U)$ is 0.78 (see Fig. 6d). With land included, the $EC1(\partial T/\partial y)$ is annual while $EC1(U)$ is biannual, although it becomes much less distinct with weak amplitude. The winter-to-spring $EC1(U)$ cycle is in fact barely significant, and the zero-lag correlation between monthly $EC1(\partial T/\partial y)$ and $EC1(U)$ now drops to 0.46. The results suggest that the contribution of the $\partial T/\partial y$ from the ocean coastal-upwelling process is more tightly coupled with the wind because both tend to vary biannually, than the $\partial T/\partial y$ contribution from land–sea temperature contrast, which is strongly annual and therefore tends to be out of sync with the biannual wind.

REFERENCES


