Multiyear satellite observations of the atmospheric response to Atlantic tropical instability waves

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[1] High-resolution satellite measurements from the Tropical Rainfall Measuring Mission (TRMM), Quick Scatterometer (QuikSCAT), and Special Sensor Microwave Imager (SSM/I) are used to study the variability of sea surface temperature (SST), surface wind velocity, water vapor, cloud liquid water, and precipitation associated with westward moving tropical instability waves (TIWs) in the Atlantic Ocean from 1998 to 2005. Coherent ocean-atmosphere patterns are shown during these 8 years. Southeasterly trades strengthen and water vapor increases over warm SST anomalies associated with TIWs. The opposite is true over cold TIW SST anomalies. The cloud liquid water and rain response to the SSTs follows a similar pattern, appearing generally downstream of SST anomalies in the central tropical Atlantic. The atmospheric response to the TIW SST anomalies extends north of the TIW active region, suggesting a remote response to the TIWs. The atmospheric response to the Atlantic TIWs also exhibits interannual variations. In 1999, owing to the southward movement of the Atlantic Intertropical Convergence Zone (ITCZ), the rainfall response to the TIW SST anomalies is much larger than in other years. When the Atlantic ITCZ moves south, it is more susceptible to TIW influence.


1. Introduction

[2] Cusp-shaped tropical instability waves (TIWs) are observed in the tropical Pacific and Atlantic Oceans as westward propagating oscillations of the temperature front between cold upwelling equatorial water and warmer water to the north [Legckis, 1977; Düngen et al., 1975]. The existence of TIWs is generally attributed to the instability of meridional shears associated with strong equatorial currents flowing in alternating directions [Philander, 1978; Cox, 1980].

[3] While TIWs have oceanic origin, several recent studies have presented evidence that the sea surface temperature (SST) variations associated with TIWs can induce atmospheric variability. Halpern et al. [1988] first indicated a possible relationship between surface wind and SST. Hayes et al. [1989] also found a significant correlation between wind speeds and SST perturbations associated with TIWs from an array of moored sensors. Xie et al. [1998] used satellite data to show that TIW SST variations induce wind perturbations. They further noted that the southeasterly trade winds are enhanced over the warm SST anomalies and reduced over the colder anomalies when the data are filtered to isolate TIW variability. Using satellite observation of visible cloudiness and SST, Deser et al. [1993] found strong association between visible cloud and the SST waves, with enhanced cloudiness in the warm troughs of the waves and reduced cloudiness in the cold crests of the waves. Using data for 1999 from the Tropical Rainfall Measuring Mission (TRMM) and the Quick Scatterometer (QuikSCAT), Hashizume et al. [2001] found covariability of SST, surface wind velocity, column-integrated water vapor, cloud liquid water, and precipitation associated with TIWs. The association of clouds and rain with TIWs indicates that the atmospheric response is not limited to the surface but extends at least through the boundary layer.

[4] Interactions between the ocean and the atmosphere play a fundamental role in determining tropical climate variability. Current research on the response of the atmosphere to SST variation has focused on the El Niño-Southern Oscillation (ENSO) phenomena in the tropical Pacific. ENSO wind anomalies are mostly induced by changes in deep convection, which are limited to high-SST regions. TIWs, on the other hand, induce wind changes over relatively cool SSTs. TIWs are an important aspect of equatorial ocean dynamics and thermodynamics. The horizontal eddy drag of the TIWs on the South Equatorial Current (SEC) and North Equatorial Countercurrent (NECC) has the same order of magnitude as the mean trade wind stress [Hansen and Paul, 1984]. Over the tropical Pacific basin, the TIW-induced oceanic eddy heat flux towards the equator is comparable to the Ekman heat flux away from the equator and the large-scale net air-sea heat flux [Hansen and Paul, 1984; Wang and McPhaden, 1999]. Since TIWs transport a substantial amount of heat and momentum across the equatorial front [Hansen and Paul, 1984; Weisberg, 1984], this high-frequency ocean-atmo-
sphere interaction could have an effect on the tropical climate [Xie et al., 1998].

[5] At present there are two main hypotheses concerning the relationship between SST and surface winds over the tropical oceans. In the first hypothesis, SST affects sea level pressure, which then changes the surface wind [Lindzen and Nigam, 1987]. Lindzen and Nigam [1987] postulate that SST-induced air temperature anomalies can change the sea level pressure (SLP) by changing the hydrostatic balance. Pressure gradients lead to low-level convergence over warm SST and divergence over cold SST. The mechanism is similar to sea breeze circulations, which are also driven by surface heating gradients. In this scenario, warm SST is associated with low SLP and cool SST is associated with high SLP. In the tropics, surface winds tend to flow down the SLP gradient. Hashizume et al. [2001] found some evidence to support this hypothesis. In the second hypothesis, SST is coupled with wind through stability changes in the atmospheric boundary layer [Hayes et al., 1989; Wallace et al., 1989]. When the southeasterly trade winds flow from the colder to the warmer waters, the atmosphere becomes more buoyant, mixing increases, and wind shear is reduced in the boundary layer. As a result, surface winds increase. That is, increased boundary layer mixing brings high momentum air down to the surface. Support for this hypothesis has been reported by Liu et al. [2000], Hashizume et al. [2001], and Chelton et al. [2000].

[6] Compared with the tropical Pacific, our current understanding of TIW air-sea feedback mechanisms in the tropical Atlantic is still in the beginning stage. Since TIWs are undulations of the SST cold front in the ocean field, TIW activity is closely related to the evolution of the equatorial cold tongue. In the Atlantic Ocean, the cold tongue develops rapidly due to geographic features of the Atlantic basin [Mitchell and Wallace, 1992]. As a consequence, TIWs also develop rapidly, beginning in early June, reaching maximum amplitude in early August, and then decaying as they propagate westward. In the Atlantic, TIWs are active in the region between the Intertropical Convergence Zone (ITCZ) and the equator, where the SST gradient is large.

[7] Wu and Bowman [2007] showed that TIW SST interannual variability in the Atlantic Ocean is related to the interannual Atlantic equatorial mode (or Atlantic Niño). It is thus interesting to investigate how the atmospheric response to the TIW SST anomalies varies from year to year. Hashizume et al. [2001] showed that significant variability of wind, water vapor, cloud liquid water, and precipitation can be observed in the Atlantic ITCZ, where direct TIW forcing is absent. They hypothesized that TIWs can have a remote influence on the ITCZ. Nevertheless, the authors used only a short time series of satellite data from 1999, which prevents study of interannual variations. Caltabiano et al. [2005] studied tropical Atlantic instability waves with 5-year TRMM data, but they did not discuss interannual variations of atmosphere response to the Atlantic instability waves nor specify the atmosphere remote response to TIW SST anomalies in the Atlantic Ocean.

[8] Some important questions about the atmospheric response to TIW in the Atlantic Ocean are, What is the interannual variability of the atmospheric response to the TIW SST anomalies? Is the atmospheric remote response to TIW SST anomalies in the Atlantic Ocean shown by Hashizume et al. [2001] for 1999 typical of other years? To what extent do these SST anomalies have an impact on rainfall variability in the ITCZ? To address these questions, we use 8 years of TRMM data, 6 years of QuikSCAT data, and available SSM/I data.

2. Data and Methods

[9] SST data for this study are from the TRMM Microwave Imager (TMI). The lowest frequency channel of the TMI penetrates nonraining clouds with little attenuation, giving a clear view of the sea surface except near coasts and in regions of strong precipitation. This is a distinct advantage over traditional infrared SST observations, which require a cloud-free field of view [Wentz et al., 2000]. Because the atmosphere is nearly transparent to microwave radiation under nonraining conditions, the TMI provides an essentially uninterrupted record of the westward propagation of the SST signatures of TIWs [Chelton et al., 2000]. Most missing TMI data occur north of the TIW region in the ITCZ, where rain rates are largest. There are only a few small gaps in the TIW region. TMI SST data used in this study are Version 3a from Remote Sensing Systems (RSS) with a spatial resolution of 0.25° latitude by 0.25° longitude, spanning the period January 1998 to December 2005. Daily maps of these data sets are averaged in 3-day running means. SST data are recorded as the average of 3 days ending on the file date. Remaining gaps in the 3-day mean data are filled by linear interpolation in the zonal direction.

[10] The surface vector winds analyzed here are from the SeaWinds scatterometer [Hoffman and Leidner, 2005] launched on 19 June 1999 on board the QuikSCAT satellite. The geophysical data record began on 15 July 1999. QuikSCAT is designed to provide accurate ocean surface winds in all conditions except moderate to heavy rain. The QuikSCAT scatterometer infers wind stress magnitude and direction from measurements of microwave radar backscatter received from a given location on the sea surface at multiple antenna look angles. The wind retrievals are calibrated to the so-called neutral stability wind at a height of 10 m above the sea surface, that is, the wind that would exist if the atmospheric boundary layer were neutrally stable. QuikSCAT data extend from 89.875°S to 89.875°N with a spatial resolution of 0.25° latitude by 0.25° longitude. Daily maps of these data sets are averaged in 3-day means. The SeaWinds data are Version 3, spanning July 1999 to December 2005.

[11] The TRMM 3B42 rain rate analysis is used to study the response of rain to TIWs. The 3B42 algorithm merges microwave and infrared precipitation estimates. These grid-ded estimates have a 3-hour temporal resolution and 0.25° × 0.25° spatial resolution in a global belt extending from 50°S to 50°N latitude. Here the 3B42 data are averaged to daily temporal resolution to match the SST data resolution.

[12] The TMI also measures column-integrated water vapor and cloud liquid water. These variables are poorly sampled by a single microwave imager due to high-frequency variability. We combine the TMI and four Version 5 SSM/I instruments (F-11, F-13, F-14, F-15) that were in space simultaneously with TRMM. For data from 1998 to 1999, we combine the TMI, F-11, F-13, and F-14.
For data from 2000 to 2005, we combine the TMI, F-13, F-14, and F-15. The combination of these five microwave imagers greatly improves the sampling of water vapor and cloud liquid water. TMI, SSMI, and QuikSCAT data are processed and made available by RSS (http://www.ssmi.com).

[13] Monthly mean Reynolds SST and monthly mean outgoing longwave radiation (OLR) are used to study the interannual variability of the ITCZ location in the Atlantic Ocean. Monthly mean SST covers the period from 1982 to 2005 with 1° grid resolution. Monthly mean OLR data are from 1974 to 2005 with 2.5° grid resolution.

[14] In order to isolate TIW variability from other temperature signals, such as the seasonal cycle, a filtering scheme is used. The filter is designed using prior knowledge of the approximate range of the period and wavelength of the waves. We apply a spatial filter of 5° to 12° (~600 km to 1300 km) in longitude and 20 to 40 days in time. The filter works well to extract TIW signals. We use this filter to derive TIW anomaly fields of SST, surface wind, water vapor, cloud liquid water, and rain.

[15] Our aim is to investigate the covariability of the geophysical fields measured by the satellite. Following Hashizume et al. [2001], a linear regression technique is used to map the spatial structure of the TIWs. Using the grid point with maximum TIW SST variability as a reference point (x₀, y₀), a linear relation between the atmospheric fields at each point and SST at the reference point is computed using least squares. It seeks a linear relation between one of the filtered atmospheric fields F(x, y, t) and the filtered SST at the reference point T₀ = T(x₀, y₀, t). Since the mean of the filtered SST anomalies is approximately equal to 0, the regression relation F = bT instead of F = a + bT₀ is used.

[16] The least squares method leads to the regression coefficients,

\[ b(x,y) = \frac{\sum_{n=1}^{N} F_n(x,y,t_n) T(t_n)}{\sum_{n=1}^{N} T^2(t_n)} \]

The regression method is useful to identify the relationship between an atmospheric variable and a TIW SST index at reference point [Hashizume et al., 2001].

[17] The correlation coefficients (r²) of the regression measure the fraction of the total variance of the dependent variable explained by the SST. To investigate how much variation of the atmospheric field F(x, y) is explained by the TIW SST T, we use the same regression model, F = bT. With the same least squares method, we obtain the r² value:

\[ r^2(x,y) = S^2_{f}(x,y)/\left(S_\theta(x,y)S_{ff}(x,y)\right) \]

where,

\[ S_\theta(x,y) = \sum_{n=1}^{N} \left(F(x,y,t_n) - F(x,y,\bar{t})\right)T(t_n) \]

\[ S_f(x,y) = \sum_{n=1}^{N} T^2(t_n) \]

\[ S_{ff}(x,y) = \sum_{n=1}^{N} \left(F(x,y,t_n) - F(x,y,\bar{t})\right)^2 \]

3. Results

3.1. Atmospheric Response

[18] June, July, and August (JJA) are the active period for TIWs in the Atlantic Ocean. Figure 1 shows a 1-day snapshot of TIWs in the Atlantic Ocean. These cusp-shaped waves develop primarily north of the equator, reach maximum amplitude between 15°W and 20°W, and then decay westward.

[19] Figures 2 and 3 show the time-mean large-scale characteristics of the SST, surface wind, and precipitation fields during the TIW season. Figures 2a and 3a are the climatological maps for 1998–2005. Because 1999 had some unusual characteristics, separate maps for that year (to be discussed later) are included in Figures 2 and 3. In the tropical Atlantic, the ITCZ is collocated with maximum SSTs. During the boreal summer a cold SST tongue develops along the equator, and the ITCZ reaches its northernmost extent. On average during the years 1998 to 2005, the northern boundary of the ITCZ in the Atlantic Ocean...
Figure 2. Time-averaged SST (color) and TIW SST variance (contours in °C^2) for June, July, and August (JJA) (a) 1998 to 2005 and (b) 1999.

Figure 3. (a) Time-averaged rainfall and wind velocity (vectors) during JJA. Rainfall is for the period 1998 to 2005; winds are for 1999 to 2005. (b) Time-averaged JJA rainfall for 1999.
during JJA lies at $\sim 10^\circ N$ (determined subjectively by the 5 mm day$^{-1}$ isopleth), although the location of the ITCZ varies from year to year. The southern boundary of the ITCZ is located near $4^\circ N$, except in 1999, when the southern boundary is at $2^\circ N$ (Figures 2b and 3b).

The contours in Figure 2 represent the temporal variance of the filtered SST. This is the temporal variance at each point of the bandpass filtered SSTs. The largest filtered SST variance is found to the south of the ITCZ, where there are large meridional SST gradients. The SST variance can be used as an index of the TIW activity. TIWs are most active in the latitudes between $1^\circ N$ and $5^\circ N$. Weak TIW activity is also visible to the south of the equator (Figure 1).

Figure 3a presents the time-mean precipitation during JJA for 1998 to 2005 and the time-mean surface wind fields for 2000 to 2005. In the Tropics, rainfall is primarily associated with deep convective activity. The region of enhanced rainfall (ITCZ) constitutes the upward branch of the Hadley circulation. As the southeast trades in the Atlantic basin flow across the cold tongue, they gain speed and turn northward between $0^\circ$ and $10^\circ N$. The southeast and northeast trades converge around $9^\circ N$ over the warmest water. The mean wind speed in the TIW active region between $0^\circ$ and $5^\circ N$ is about 6.5 m/s, and the maximum seasonal-mean wind speed is about 8 m/s. As stated above, Figure 3b shows that the warmest SST and the ITCZ lie further south in 1999 than is typical of the 8-year climatology.

Figure 4 shows regressions of SST and wind for each individual year. In each panel the reference point for the regression is indicated by a plus sign, where TIW SST variance is maximum according to Figure 2a. Choosing a different reference point may change the response amplitude, but not the wave pattern. These regression maps, created using filtered SSTs, show clear horizontal structure in SST and wind velocity. As previous studies have shown [Liu et al. 2000; Hashizume et al., 2001], the typical TIW wavelength is about $10^\circ$ of longitude. Phase lines tilt southwest to northeast. The SST anomalies are approximately antisymmetric about the equator with much weaker anomalies south of the equator. The TIW-induced SST variations are $0.3 \sim 0.6 K$ in magnitude in our selected band.

The near-surface wind anomalies are generally easterly or southeasterly over the warm SST anomalies, and the reverse is true over cold anomalies. The maximum magnitude of the effect is $\sim 0.4$ to 0.6 m s$^{-1}$ K$^{-1}$ in those areas. In spite of the variability of surface winds and the difficulty of estimating derived quantities, the convergence and divergence patterns can be clearly seen in the satellite observations in Figure 5b. Figure 5b clearly shows the convergence and divergence of wind field in Figure 5a. Convergence tends to occur to the northwest of the warm SST anomalies, while divergence occurs to the northwest of the cold anomalies (Figure 5 shows 2000 as an example). Because the convergence and divergence are out of phase with the SST anomalies, this result suggests that vertical mixing...
dominates the relationship between SST and surface wind, as proposed by Wallace et al. [1989]. The Lindzen and Nigam mechanism, by contrast, would lead to convergence and divergence patterns that are in phase with respect to the SST anomalies. Although SST anomalies disappear north of 5°N, wind anomalies are observed even at the northern edge of the plot at 8°N, which suggests that there may be a link between TIWs and the ITCZ. In some years, the wind anomalies at 8°N are even larger than the local response (e.g., 2002 and 2005). The wind anomalies south of the equator can be quite large, even though the SST anomalies are small, but this response may not be real, as the patterns are not consistent from year to year.

[24] Local SST anomalies can affect the boundary-layer water vapor, and enhanced turbulence over warm SST can mix moist air upward. Figure 6 shows the regression of water vapor with SST at the reference point. While there are interannual variations in magnitude, the vapor anomalies are generally in phase with the SST anomalies in the northern hemisphere in the sense that water vapor increases above warm SST anomalies and decreases above cold SST anomalies. The water vapor anomalies are generally positive to the north and west of the positive SST anomalies, where there is convergence. The water vapor anomalies extend to 8°N, well beyond the extent of the TIW SST anomalies. The water vapor anomalies in 2004 are notably weaker than other years. Large water vapor anomalies can also be found in the southern hemisphere in some years (e.g., 1999 and 2001), although the SST anomalies are weak. Also, water vapor anomalies do not appear to be as antisymmetric about the equator as the SSTs. TIW SST variations explain about 5 to 15% of the total bandpass-filtered water vapor variations.

[25] The cloud liquid water (Figure 7) and rain (Figure 8) regressions are similar. The atmospheric anomalies appear downstream of the SST anomalies in the central tropical Atlantic. Even though the TIWs are active in latitudes between the equator and 5°N, the cloud liquid water and rain anomalies range from ~2° to 8°N. The cloud liquid water anomalies are accompanied by precipitation changes in the ITCZ.

[26] In most years the southern boundary of the Atlantic ITCZ is located at around 4°N. Compared with the Pacific ITCZ, this more southerly position of the Atlantic ITCZ makes it more susceptible to TIW variability. These results confirm the findings of Hashizume et al. [2001] that during 1999 the TIW variability had a large impact on the deep convection associated with the ITCZ. The results also confirm the suggestion of Xie et al. [1998], based on a
Figure 6. Regression maps of water vapor during JJA (colors). The plus sign indicates the reference point (15°W, 2°N) for the regressions. The contours are SST regression with intervals of 0.4 K.

Figure 7. Regression maps of cloud liquid water during JJA (colors). The plus sign indicates the reference point (15°W, 2°N) of regressions. The contours are SST regression with intervals of 0.4 K.
GCM simulation, that TIWs might induce a remote response in the atmosphere and affect the ITCZ downstream from the direct SST forcing. The cloud liquid water variability induced by TIW SST anomalies explains ~5 to 10% of the total variation, while the TIW SST-induced precipitation anomalies account for ~10 to 20% of the total variation. One distinct feature of Figure 8 is that during 1999 the rainfall anomalies are much larger than in other years and extend farther south. During 1999 the regression coefficients for rainfall reach as high as 4 mm day$^{-1}$ K$^{-1}$. At latitudes between 2° and 4°N, the TIW-induced precipitation anomalies account for ~20 to 30% of the total variation, which is larger than other years. Further analysis of this year is presented in the next section.

[27] Wu and Bowman [2007] found that the Atlantic air-sea coupled mode influences TIW activity in the Atlantic Ocean. The relationship is similar to that between ENSO and TIWs in the Pacific Ocean. TIWs are stronger during cold phases and weaker during warm phases of the coupled mode. The root mean square of the atmospheric response coefficient $b$ in the region from 2°N to 8°N and 12°W to 22°W is used as a measurement of the interannual variability of the strength of the air-sea coupling (Figure 9). The results indicate the amplitude of atmosphere response to the SST variations at the reference point. Unlike the TIW SST signature in Wu and Bowman [2007], no clear trend is detected, except perhaps for clouds, but results for 1999 are outliers in all three cases. The unusual response of rain

Figure 8. Regression maps of precipitation during JJA. The plus sign indicates the reference point (15°W, 2°N) of regressions. The contours are SST regression with intervals of 0.4 K.

Figure 9. Root mean square of the atmosphere in response coefficient $b$ in the region between 2°N and 8°N and 12°W and 22°W as a function of the Atlantic coupled mode index (ATL3) showing (a) water vapor, (b) cloud liquid water, and (c) rain. Bold cross represents the year 1999.
is particularly noticeable (Figure 9c). This indicates that the interannual variations of the atmospheric response are more complex than the interannual variation of the TIW SST signature itself.

3.2. Year 1999

[28] Wu and Bowman [2007] found the wave activity in 1999, when the coupled mode is in a warm phase, is the weakest of the 8 years from 1998 to 2005. However, the rainfall anomalies during 1999 associated with TIWs are much larger than those in all of the other years, particularly between 2°N and 4°N (Figure 8), suggesting that conditions other than the TIW activity must have influenced the atmospheric response to the TIW SSTs.

[29] Figures 2 and 3 show that in 1999 the southern Atlantic ITCZ boundary in JJA moves ~2° further south. This southward movement of ITCZ makes 1999 very different from the other years in the latitude zone of TIW activity. Figure 10 shows the latitudinal distribution of zonal-mean SST, water vapor, cloud liquid water, and rain for JJA in the tropical Atlantic. The zonal-mean SST in the TIW active region (from the equator to 5°N) during 1999 is about 0.5°C to 1°C higher than other years and the peak SSTs occur several degrees south of the typical position. The zonal mean water vapor shows little variability from 1998 to 2005, although water vapor at 2°N in 1999 is higher than in other years. Owing to the sharp meridional gradient, the zonal-mean cloud liquid water at 2°N in 1999 is about 0.08 mm, while in the other 7 years, the zonal-mean cloud liquid water is 0.03 to 0.04 mm. The zonal-mean rain rate at 2°N in 1999 is about 3 mm/day, while in the other 7 years, zonal-mean rain rates range from 0.1 to 0.7 mm/day. Both the zonal-mean cloud liquid water and the rain rate in 1999 are several times larger than other years. We consider that the southward movement of the ITCZ makes it more susceptible to TIW influences.

3.3. ITCZ Interannual Variability

[30] In this section we consider how the Atlantic ITCZ location might influence the atmospheric response to the oceanic TIWs. Gu and Adler [2006] show that the ITCZ strength and total rainfall amount in the tropical Atlantic basin are significantly modulated by the Pacific El Niño and the Atlantic Niño, particularly during boreal spring and summer, whereas the impact of the Atlantic interhemispheric mode is considerably weaker. El Niños influence the tropical Atlantic in two distinct ways [Saravanan and Chang, 2000; Giannini et al., 2001; Chiang et al., 2002]: (1) by modulating SST in the tropical North Atlantic through the Pacific-North American Pattern (PNA) several months after the tropical Pacific anomalies and (2) by suppressing rainfall in the equatorial Atlantic via an anomalous Walker circulation and increased stability through widespread tropospheric warming in the tropics, which occurs with a much smaller time lag. The Atlantic Niño itself can directly influence both rainfall and latitudes of the ITCZ [Delecluse et al., 1994; Latif and Grötzner, 2000].

[31] We use 32 years of OLR (from 1974 to 2005) and 24 years of Reynolds SST data (from 1982 to 2005) to study the variability of the Atlantic ITCZ. Figure 11 shows the Atlantic ITCZ as a function of latitude in the period of JJA with OLR and SST data. During the JJA period of 1999, OLR is colder than any of the other years and SST is as...
warm or warmer than any other years in the TIW active region (0° and 5°N). By chance, the year that Hashizume et al. [2001] analyzed was extreme in terms of Atlantic SSTs and the position of the ITCZ and should not be taken as representative of average conditions during recent years.

4. Conclusions

[32] We use high-resolution satellite measurements from TRMM, QuikSCAT and SSM/I to study coupled ocean-atmosphere variability in the equatorial Atlantic from 1998 to 2005. Bandpass filtering and linear regression are used to extract the characteristics of the atmospheric response to the SST anomalies created by tropical instability waves (TIWs). The highly coherent patterns from our analysis indicate that current satellite technologies are capable of detecting intra-seasonal changes in atmospheric variables, including precipitation, even within the noisy ITCZ.

[33] Coherent ocean-atmosphere patterns appear throughout the study period. Although TIWs are largely confined to a narrow latitudinal zone just north of the equator, their atmospheric response extends farther north and can influence the ITCZ. The observations show that the southeastern trades strengthen over warm TIW SST anomalies and weaken over cold SST anomalies. Warm anomalies have an amplitude of ~0.4 to 0.6 m s⁻¹ K⁻¹ at 2°N. The water vapor variability induced by TIW SST anomalies explains ~5 to 15% of the total variation. Cloud liquid water and rainfall responses to the TIW SST follow similar patterns, appearing to downstream of the SST anomalies. The TIW-induced cloud liquid water anomalies account for ~5 to 10% of the total variation and precipitation anomalies account for ~10 to 20% of the total precipitation variation.

[34] Atmospheric responses to TIW SSTs extend to 8°N, which is north of the principal TIW activity zone, suggesting a remote downstream response to the TIW SST anomalies. The atmospheric responses to these TIWs show interannual variability. In 1999 the rainfall response to TIWs in the latitude between 2°N and 4°N is much larger than in other years, which we attribute to the anomalously warm SSTs at 2°N and the southward movement of Atlantic ITCZ in that year. The atmospheric response in 1999 [Hashizume et al., 2001] does not appear to be typical of other years. The comparatively warm temperatures in the TIW region in 1999 apparently increased the precipitation and made the atmosphere more sensitive to TIW variations. When the ITCZ takes a more southerly position, the synoptic variability in precipitation can be larger than its monthly mean. As the Atlantic ITCZ moves south, it is more susceptible to TIW influences.

[35] This paper suggests that vertical mixing dominates the relationship between SST and surface wind, as proposed by Wallace et al. [1989] because the convergence and divergence are out of phase with the SST anomalies. We cannot conclude, however, that both mechanisms are not operating. Using a regional model in the Pacific, Small et al. [2003] found vertical mixing, horizontal advection, and the perturbation pressure gradient are all induced by the SST changes. Horizontal advection leads to the occurrence of the air temperature and moisture extrema downwind of the SST extrema. Previous studies [Hashizume et al., 2001; Liu et al., 2000; Calabiano et al., 2005] have used the relative phase between atmosphere response and SST to test the relationship between surface wind and SST. However, this method may be unreliable because of the thermal and moisture advection by the mean wind as suggested by Small et al. [2003]. Further exploration of air-sea coupling mechanism of TIW in the Atlantic Ocean is underway with numerical models.

[36] How the atmosphere responds to intraseasonal SST changes is an important aspect of climate variability in the Atlantic Ocean. Current research on large-scale ocean-atmosphere interaction has focused on ENSO phenomena in the tropical Pacific, a process limited to high-SST regions. Most of the ocean in the world is not warm enough to support deep convection. Thus TIWs provide an opportunity for studying the atmospheric response in cooler SST regimes. One obvious question associated with TIWs is whether these TIW-induced perturbations of wind field have a significant feedback onto the ocean circulation. Pezzi et al. [2004] showed that the wind perturbation have a negative feedback that tends to reduce the SST and meridional velocity signatures of TIW. This, in turn, reduces equatorward eddy heat flux associated with TIW, thereby resulting in intensified cooling of equatorial cold tongue. This negative feedback has potential important implications about the role of TIW in climate variability. Numerical models should prove useful for investigating the mechanisms of ocean-atmosphere interaction associated with TIW.

References


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