Free and Forced Variability of the Tropical Atlantic Ocean: Role of the Wind–Evaporation–Sea Surface Temperature Feedback

SALIL MAHAJAN
Princeton University, Princeton, New Jersey

R. SARAVANAN AND PING CHANG
Department of Atmospheric Sciences, Texas A&M University, College Station, Texas

(Manuscript received 19 June 2009, in final form 22 April 2010)

ABSTRACT

The role of the wind–evaporation–sea surface temperature (WES) feedback in the low-frequency natural variability of the tropical Atlantic is studied using an atmospheric global climate model—the NCAR Community Climate Model, version 3 (CCM3)—thermodynamically coupled to a slab ocean model (SOM). The coupled model is modified to suppress the WES feedback and is compared to a control run. Singular value decomposition (SVD) analysis over the tropical Atlantic reveals that the coupled meridional mode of the Atlantic Ocean is amplified in the presence of the WES feedback. In its absence, the meridional mode still exists, but with a weaker amplitude. A feedback mechanism that involves the near-surface specific humidity is proposed to sustain the weaker Atlantic meridional mode in the absence of the WES feedback. Similar analysis of coupled model integrations when forced with an artificial El Niño–Southern Oscillation (ENSO)-like SST cycle in the Pacific reveals that in the presence of the WES feedback, the meridional mode is the preferred mode of response of the tropical Atlantic to ENSO forcing. In the absence of the WES feedback, the tropical Atlantic response is unlike the meridional mode and the effects of tropospheric warming and subsidence dominate. Regression analysis over the tropical Atlantic reveals that the meridional mode response to ENSO peaks in the spring and begins to decay in the fall in the coupled model in the presence of the WES feedback. The WES feedback also appears to be responsible for the northward migration of the ITCZ during ENSO events.

1. Introduction

The meridional mode of variability in the tropical Atlantic, also known as the interhemispheric mode or the dipole mode, is about as strong as the zonal Atlantic Niño mode (e.g., Ruiz-Barradas et al. 2000). The equally dominant meridional mode is a distinguishing feature of the Atlantic as compared to the neighboring tropical Pacific, where the zonal El Niño–Southern Oscillation (ENSO) mode clearly dominates. While specific cross-equatorial dipole sea surface temperature (SST) patterns have been observed over the tropical Atlantic (Enfield and Mayer 1997; Enfield et al. 1999), they are not believed to characterize the meridional mode. Statistical analysis studies (Houghton and Tourre 1992; Rajagopalan et al. 1998; Enfield et al. 1999; Mehta 1998) reveal that the meridional mode does not manifest itself as a simple cross-equatorial antisymmetry of SSTs at interannual time scales, as was previously suggested (e.g., Moura and Shukla 1981), but exists as cross-equatorial SST gradients (CESGs) that result in dipolelike anomalies of atmospheric variables (e.g., Enfield et al. 1999).

The physical mechanisms of the meridional mode and its associated time scales, however, have remained elusive. Observational studies analyzing only SSTs indicate little correlation between northern and southern tropical Atlantic SSTs on interannual to decadal time scales (Houghton and Tourre 1992; Mehta 1998). However, a recent study finds the leading modes of north and south tropical Atlantic SSTs to be correlated (Tanimoto and Xie 2002) on decadal time scales. A decadal variability of the observed CESGs has also been noted, suggesting a meridional cross-equatorial relation (e.g., Carton et al. 1996; Rajagopalan et al. 1998). A modeling study by Seager

Corresponding author address: Salil Mahajan, AOS Program, Princeton University, Princeton, NJ 08540.
E-mail: smahajan@princeton.edu

DOI: 10.1175/2010JCLI3304.1

© 2010 American Meteorological Society
et al. (2001) using ocean models forced by observed winds shows that the observed variability in the tropical Atlantic SSTs on decadal time scales can be explained primarily by the influence of winds on SST via surface fluxes, with the ocean circulation merely acting to damp the effects of winds. Carton et al. (1996) in an ocean modeling study find that features of off-equatorial interannual and interdecadal tropical Atlantic SST variability are absent when the ocean model surface fluxes are computed using climatological winds. Additionally, studies using simple and complex coupled models raise the possibility of the existence of an oscillatory dipolelike meridional mode of the tropical Atlantic SST on the interannual and decadal time scales caused by the coupled dynamics of an atmosphere–ocean system (Chang et al. 1997; Curtis and Hastenrath 1995; Kushnir et al. 2002; Saravanan and Chang 2000; Giannini et al. 2000; Xie 1999; Huang and Shukla 1997).

An important mechanism contributing to the meridional mode that emerges among various studies is the positive wind–evaporation–SST (WES) feedback, the mechanics of which are the focus of this study.

The boundary layer WES feedback is purely a thermodynamic mechanism, which in the deep tropics works as follows: Consider an anomalous cross-equatorial northward positive SST gradient. This generates anomalous northward cross-equatorial surface winds because of the development of an anomalous southward sea level pressure gradient. The induced northward cross-equatorial winds turn westward (eastward) in the southern (northern) side of the equator because of the Coriolis force, thereby enhancing (reducing) the background south-easterly (northeasterly) trade winds, leading to an increase (decrease) in evaporation, which decreases (increases) the SST further. The interaction between the surface winds, evaporation and SST thus forms a positive feedback amplifying the initial SST gradient (Xie and Philander 1994). In simple dynamic ocean–atmosphere modeling studies, the growth of SST, winds, and evaporation anomalies caused by the WES feedback is limited by oceanic processes, such as the meridional transport of SST anomalies by advection due to mean cross-equatorial currents (Chang et al. 1997, 2001), and poleward SST advection by Ekman flow forced by trade wind anomalies (Xie 1999), leading to self-sustained decadal oscillations when the wind–SST coupling is sufficiently strong. When the air–sea coupling is weak, Chang et al. (2001) suggest that local air–sea feedbacks enhance the persistence of cross-equatorial anomalies.

Here, we study the role of the WES feedback in the low-frequency variability, free and that forced by ENSO, of the tropical Atlantic using an atmospheric general circulation model (GCM)–National Center for Atmospheric Research (NCAR) Community Climate Model, version 3 (CCM3), coupled to a slab ocean model (SOM). The SOM is a static ocean model interacting with the atmospheric model only thermodynamically, allowing us to focus on thermodynamic processes such as the WES feedback. The lack of ocean dynamics in the SOM implies that self-sustained decadal oscillatory dipole features of simple models would be absent, as oceanic advection was found to be dominant in reversal of the dipole pattern in those simple model studies on decadal time scales (Chang et al. 1997, 2001; Xie 1999). However, in the absence of oceanic advection, Kushnir et al. (2002) found that low-frequency variability of the tropical Atlantic still exists in a simple coupled model and is caused by the selective amplification of random forcing—either intrinsic to the tropical Atlantic or from external sources, such as ENSO or the North Atlantic Oscillation (NAO)—by the WES feedback. An analysis of the CCM3–SOM, hence, still allows for the study of variability associated with the WES feedback. The NAO is also believed to influence the north tropical Atlantic (e.g., Xie and Tanimoto 1998; Czaja et al. 2002); however, recent studies find little effect of the NAO on the deep tropical Atlantic (Ruiz-Barradas et al. 2000; Rajagopalan et al. 1998). Here, we choose to ignore the simulated influence of the NAO on tropical Atlantic variability, as the effect of the WES feedback is found to be limited to 10°N in its northward extent in observations (e.g., Czaja et al. 2002).

Regression of ENSO indices on SST over the Atlantic reveals a CESG pattern similar to the meridional mode (Enfield and Mayer 1997; Ruiz-Barradas et al. 2000; Saravanan and Chang 2000). About 30% of the variability of the Atlantic meridional mode can be explained by ENSO (Ruiz-Barradas et al. 2000; Enfield and Mayer 1997), suggesting that the meridional mode is the preferred mode of response of the tropical Atlantic to ENSO events (Enfield and Mayer 1997). The understanding of the precise mechanisms of the response of ENSO on the Atlantic through atmospheric teleconnections and the associated potential predictability are subjects of ongoing research. Several observational studies have found significant warming (cooling) in the north tropical Atlantic in response to El Niño (La Niña) (Curtis and Hastenrath 1995; Enfield and Mayer 1997; Klein et al. 1999). Much of the warming during El Niño events has been attributed to a reduction in latent heat fluxes over the north subtropical and tropical Atlantic, because of the reduction in trade winds in a number of studies (Curtis and Hastenrath 1995; Enfield and Mayer 1997; Klein et al. 1999; Saravanan and Chang 2000), caused by the Pacific–North America (PNA) teleconnection (e.g., Wallace and Gutzler 1981; Nobre and Shukla 1996) and the modification of the Walker and Hadley circulations (Klein et al.
1999; Saravanan and Chang 2000). Alexander and Scott (2002) also find significant contributions of sensible heat flux, also affected by winds, and shortwave heat flux in the response of the northern tropical Atlantic to ENSO.

Air–sea differences in temperature and specific humidity have also been found to be responsible for the increase in SST over the northern tropical Atlantic during El Niño (Saravanan and Chang 2000; Chikamoto and Tanimoto 2005). Entire tropical tropospheric warming during ENSO warm events (e.g., Charney 1963; Schneider 1977; Held and Hou 1980; Sobel and Bretherton 2000) causes a warming of the SSTs over the tropical Atlantic by boundary layer adjustment of humidity, which modulates surface evaporation (Chiang and Sobel 2002). This mechanism is referred to as the tropospheric temperature (TT) mechanism (Chiang and Sobel 2002). There are also air–sea coupled Bjerknes feedback-like dynamical interactions excited over the tropical Atlantic by ENSO, which complicate the remote ENSO influence, as shown by Chang et al. (2006).

Modeling studies suggest a strong role of local thermodynamic air–sea feedbacks over the Atlantic. Suppressing the thermodynamic local air–sea coupling in the models has been found to substantially reduce tropical Atlantic variability (Carton et al. 1996; Wu and Liu 2002). Over the north tropical and equatorial Atlantic, it has been proposed that CESG anomalies, generated by local processes or externally induced, grow to reach observed amplitudes through amplification by the thermodynamic WES feedback (Enfield and Mayer 1997; Xie and Carton 2004; Kushnir et al. 2002; Sutton et al. 2000). However, only a weak response associated with the WES feedback is observed over the equatorial south tropical Atlantic (Enfield and Mayer 1997).

Since its formal introduction by Xie and Philander (1994), the WES feedback has been hypothesized to contribute significantly to tropical climate variability. In observations and modeling studies, the existence of the WES feedback is suggested by correlating and covarying patterns of SSTs, surface heat fluxes, and winds (e.g., Carton et al. 1996; Chang et al. 1997; Saravanan and Chang 2000). It finds further support in modeling studies, where a difference in the response between atmospheric models forced with climatological SSTs and atmospheric models coupled with interactive oceans (e.g., Saravanan and Chang 2000; Chang et al. 1997) suggests the presence of feedbacks in the atmosphere–ocean coupled system. While most studies invoking the WES feedback to explain a physical mechanism are supported by strong theoretical arguments, few explicitly demonstrate the presence of the WES feedback. The goal of this investigation is to clearly isolate the role of the WES feedback from concomitant mechanisms in the free and forced variability of the tropical Atlantic by comparing GCM simulations with and without the WES feedback.

The next section describes the model setup and the approach to switch off the WES feedback in the CCM3–SOM. An analysis of the role of the WES feedback in the free meridional mode of the Atlantic as simulated in the CCM3–SOM using singular value decomposition (SVD) analysis is presented in section 3. A detailed investigation of the role of the WES feedback in the response of the Atlantic to a forced idealized ENSO cycle is carried out in section 4. We conclude by summarizing and discussing our results in section 5.

2. Model

To better understand the effect of the WES mechanism over the Atlantic Ocean, we employ the NCAR CCM3 a general circulation atmospheric model (Kiehl et al. 1998) coupled to a slab ocean model with spatially varying but temporally constant ocean mixed layer thickness. The SOM lacks ocean dynamics and only interacts with the atmosphere above thermodynamically, allowing for a clear analysis of thermodynamic interactions such as the WES feedback. Sea ice in the model is made noninteractive and is prescribed to follow current climatological mean state sea ice extent. The model integrations are conducted at a spectral resolution of T42, equivalent to a spatial resolution of about 2.81° × 2.81°. A monthly climatology of heat adjustment, called Q, flux, is prescribed to the SOM to account for absent mean ocean dynamics (Kiehl et al. 1996).

The latent and sensible heat fluxes over the ocean surface in the CCM3 are computed via bulk aerodynamic formulations, which incorporate the stability of atmosphere and turbulent scales in exchange coefficients (Kiehl et al. 1998). The bulk formulations can be simplified for our purposes here as

\[ Q_{\text{sh}} = -(u^*C_e\Delta q)\rho L_{\text{vap}} \]  \hspace{1cm} (1)

\[ Q_{\text{sh}} = -(u^*C_d\Delta T)\rho C_p, \]  \hspace{1cm} (2)

where \( Q_{\text{sh}} \) and \( Q_{\text{sh}} \) are the latent and sensible heat fluxes, respectively, defined to be positive for an upward transfer from the ocean to the atmosphere; \( u^* \) is a product of the wind speed at the lowest atmospheric level and a neutral momentum exchange coefficient; \( \Delta q \) is the difference \( (q_1 - q_0) \) in the specific humidity \( q \) of the lowest atmospheric surface and the surface saturation specific humidity \( q_0 \); \( C_e \) is a neutral tracer exchange coefficient; \( \Delta T \) is the difference \( (T_a - T_s) \) in the potential temperature of the lowest atmospheric surface \( T_a \) and the ocean
surface temperature \((T_s)\); \(C_d\) is a neutral heat exchange coefficient; \(\rho\) is the density of air at the lowest atmospheric surface; \(C_p\) is the specific heat capacity of moist air; and \(L_{\text{vap}}\) is the latent heat of the vaporization of water.

The control run of the thermodynamically coupled model is termed as CCM3–SOM integration in the analysis that follows. We also integrate a modified version of CCM3–SOM, termed as the WES-off–SOM run, where we suppress the effect of wind variations on both the latent and sensible heat surface fluxes over global oceans. This is achieved by replacing model computed \(u^*\) with prescribed climatological values of \(u^*\) in the above formulations [Eqs. (1) and (2)]. The replacement is carried out only over ocean grid points. The prescribed value of \(u^*\) in the WES-off–SOM run for the computation of surface heat fluxes is assigned to be the monthly climatology of the control CCM3–SOM \(u^*\) at each ocean grid point. The value of \(u^*\) is interpolated from these prescribed monthly CCM3–SOM climatological values at each model time step, and it is used only in the computation of surface latent and sensible heat fluxes. The suppression of the effect of fluctuations in the winds on surface heat fluxes switches off the thermodynamic feedbacks associated between winds and surface heat fluxes (both latent and sensible heat flux), including the WES feedback, over global oceans in the WES-off–SOM run. Latent heat fluxes dominate over the tropics; hence, the WES feedback is the dominant effect that is modified in the WES-off runs.

It should be noted that the surface winds, and hence \(u^*\), in the WES-off–SOM run is still modulated by the SST, in exactly the same physical manner as in the control integration, and there is no discrepancy between surface winds and model-computed \(u^*\) in the WES-off–SOM run. The differences in the two runs arise only from the modifications made in the computation of the bulk aerodynamic formulations over global ocean grid points, which uses prescribed \(u^*\) in the WES-off–SOM run. The usage of control CCM3–SOM integration monthly \(u^*\) climatology as the prescribed \(u^*\) in the bulk formulations of WES-off–SOM run also ensures a similar climate mean state of the two integrations. A detailed comparison of the CCM3–SOM and the WES-off–SOM integrations can be found in Mahajan et al. (2009).

To understand the influence of ENSO on the Atlantic, we carry out separate runs with a forced ENSO-like SST cycle in the tropical Pacific. To isolate our analysis from ENSO multiyear variability, we use a prescribed idealized 4-yr ENSO-like SST cycle over the tropical Pacific. An artificial ENSO cycle is generated by adding a weighted cosine function with an amplitude equal to one standard deviation of the observed Niño-3 index and a period of 4 yr, to the observed SST climatology over each grid point of the CCM3 over the tropical Pacific. The weights of the cosine functions are taken to be the loadings of the first EOF of observed SSTs over the tropical Pacific derived from the Carton–Geise Simple Ocean Data Assimilation (SODA) dataset available at the International Research Institute for Climate Prediction (IRI) data library (Carton et al. 2000). The generated ENSO SST cycle can be represented as

\[
T_{i, j} = \tilde{T}_{i, j} + e_i \cos \left( \frac{2\pi t}{48} - m \right),
\]

where \(T_{i, j}\) is the SST at gridpoint \(i\) and time \(t\) in months, such that \(t = 0\) corresponds to January of the first year of simulation; \(\tilde{T}_{i, j}\) is the climatology of SST at gridpoint \(i\) for the month number of \(j = t \mod 12\); \(e_i\) is the loading of the first EOF at gridpoint \(I\); \(A\) is the amplitude of the cosine function, that is, of the artificial ENSO cycle; and \(m\) is the phase of the cosine function, whereby \(m = 11\) implies that ENSO peaks in December, as applied in this study. SST anomalies prescribed in the above manner ensure a spatial pattern of ENSO forcing for the CCM3 that is similar to the real-world ENSO pattern. The integration forced with the prescribed ENSO cycle in the control CCM3–SOM model is termed as CCM3–SOM–ENSO and that in the modified WES-off model is termed as WES-off–ENSO. All integrations are run for 75 yr. The analysis presented here is based on 70 yr of integrations after discarding the first 5 yr as spinup time.

As a caveat, our experiment design removes the influence of surface winds on surface heat fluxes and associated feedback processes over global oceans in the WES-off integrations. Hence, the simulated tropical Pacific climate itself, having remote influences on the Atlantic, is also affected in the WES-off–ENSO runs. The response of net surface heat flux over the tropical Pacific is found to be weaker in the WES-off–ENSO run. However, switching the WES feedback off only weakly diminishes the simulated convective precipitation over the tropical Pacific—the diabatic atmospheric heat source—in response to ENSO SST forcings, as is seen in Fig. 1, which shows the regression of January–March averaged convective precipitation on the January Niño-3 index for the two runs. The weaker response of WES-off–ENSO convective precipitation, nonetheless, suggests that by experiment design, the tropical Pacific atmospheric response to prescribed ENSO SST forcings in the WES-off–ENSO run is a little weaker, and thus the atmospheric response over the Atlantic would also be expected to be slightly weaker.

Additionally, when removing the influence of winds on the surface heat fluxes to remove the local thermodynamic feedbacks between the winds and the SST, our experimental design also removes the influence of
externally forced winds on the SST. We discuss this further in the following sections in context of the Atlantic meridional mode and the Atlantic response to ENSO.

3. Free tropical Atlantic variability

a. SVD analysis

The thermodynamic coupled variability over the tropical Atlantic is analyzed by performing an empirical SVD analysis of SST and $Q_{lh}$ following Chang et al. (1997), who suggested the presence of WES feedback in the tropical Atlantic performing a similar analysis on observational data. Applying SVD matrix operation on the cross-covariance matrix of two spatiotemporal fields allows for the identification of pairs of leading spatial patterns of the two fields with maximized temporal covariance (Bretherton et al. 1992). SVD was performed on the cross-covariance matrix of monthly SST and $Q_{lh}$ anomalies, after removing the annual cycle, over the tropical Atlantic (20°S–20°N). Figure 2a shows the anomalies of SST and $Q_{lh}$ associated with the first leading SVD mode of variability over the tropical Atlantic for the CCM3–SOM integration. The spatial patterns presented are obtained by scaling the singular vectors of each field of the SVD mode with the standard deviation of the corresponding expansion coefficient time series.

The first SVD mode accounts for 42% of the squared covariance between SST and $Q_{lh}$, and their expansion coefficient time series exhibits a maximum simultaneous correlation of 0.6, suggesting strong covariability of the leading patterns. A cross-equatorial dipole pattern is observed in the SST field with basinwide northern and southern poles extending from the equator to about 20° latitudes in the CCM3–SOM run. A dipole in $Q_{lh}$ is also observed; however, the spatial extent of its poles is limited to the deep tropics from the equator to about 10° latitudes. Poleward of 10° latitudes, $Q_{lh}$ variability is small, and in fact it seems to oppose the development of SST anomalies in those regions. Over the deep tropical Atlantic (10°S–10°N) in CCM3–SOM integration, heterogeneous correlation maps of SST and $Q_{lh}$—obtained by
correlating gridpoint times series of one field with the expansion coefficient time series of the other—account for 20% and 8% of domain-integrated variability, respectively further indicating the covariability of the leading patterns.

The leading SVD spatial patterns resemble the observed coupled meridional mode of variability (e.g., Chang et al. 1997). In the absence of coupled dynamics, thermodynamical modes could be expected to be dominant. Hence, the appearance of the meridional mode, which is believed to be thermodynamically driven, as the leading SVD pattern in the CCM3–SOM is not surprising. A detailed analysis of the simulated Atlantic meridional mode in the CCM3–SOM and a comparison with observations can be found in Biasutti (2000). Briefly, the modeled meridional mode pattern displays some differences with the observed pattern in its poleward extent. In an SVD analysis of observed SST, winds, and net surface heat flux, strong air–sea thermodynamic coupling is also seen around 15°S and 15°N in addition to the deep tropics (Chang et al. 1997). However, the model results are consistent in the northern tropical Atlantic with that observed in reanalysis data by Frankignoul and Kestenare (2005), who, using rotated maximum covariance analysis, find that $Q_{lh}$ damps the SST north of 10°N. In the southeastern tropical Atlantic, SST–low cloud negative feedback is found to be dominant in observations (Xie and Carton 2004), and $Q_{lh}$ is found to damp the SST anomalies, similar to that noted in the CCM3–SOM simulation.

Lagged cross correlations between the expansion coefficients time series of the SVD mode for SST and $Q_{lh}$ are shown in Fig. 3a. Symmetric lagged correlations about zero lag suggest a presence of positive feedback between SST and $Q_{lh}$ with the SST and $Q_{lh}$ leading patterns reinforcing each other, similar to observations (Chang et al. 1997). The collocation of $Q_{lh}$ and SST centers of variability suggests a positive feedback between the two, with warming (cooling) by $Q_{lh}$ occurring where the SSTs are warm (cold), over the deep tropics (e.g., Carton et al. 1996). A regression of zonal and meridional winds on the expansion coefficient time series of SST of the SVD mode is also shown in Fig. 2a. Regressed wind anomalies are plotted only for grid points where the wind anomalies are estimated to be statistically significant at the 95% level, from a two-tailed $t$ test. CCM3–SOM wind anomalies resemble the observed C-shaped wind anomalies (Chang et al. 1997), with southeasterlies (southerlies, southwesterlies) appearing over the southern (northern) deep tropics, enhancing (weakening) the trade winds in the regions where the SSTs are cold (warm) and $Q_{lh}$ has a cooling (warming) effect, and hence they are suggestive of the WES feedback in the CCM3–SOM.

The role of the WES feedback over the deep tropics is further elucidated by comparing the modes of variability of the control run to that of the WES-off–SOM run. Figure 2b shows the spatial pattern of the anomalies of SST and $Q_{lh}$ associated with the leading SVD mode of SST and $Q_{lh}$ for the WES-off–SOM run. A substantial weakening of the variability of the SST dipole pattern in the SVD mode is observed compared to the CCM3–SOM integration over the deep tropics between the equator to 10° latitudes. Figures 3b,c show the spectrum of the expansion coefficient time series of both SST and $Q_{lh}$ of the leading modes of the WES-off–SOM and CCM3–SOM runs. A sharp reduction in the variability of the meridional mode in the WES-off–SOM run on the interannual time scales as compared to the CCM3–SOM run is observed in both spectrum plots, clearly indicating that the WES feedback is responsible for the
low-frequency variability of the meridional mode. An increase in variability on decadal–multidecadal time scales is also observed in the WES-off–SOM run. However, the robustness of the decadal variability is suspect in our analysis of only 70 yr of model integration, and the result could very well have been caused by sampling errors. We plan to investigate the decadal variability of the WES-off–SOM integration in a future analysis of longer simulations.

As a caveat, our experiment design also removes the influence of externally forced winds on the SST in addition to switching off the wind–SST feedbacks in the WES-off runs. In the absence of ENSO, the dominant external forcing on the tropical Atlantic in the CCM3–SOM run would be the NAO. However, the meridional mode in the CCM3–SOM is found to be located in the equatorial region, whereas NAO’s influence under realistic air–sea thermodynamic coupling is confined to the subtropical North Atlantic (e.g., Chang et al. 2001). Additionally, as mentioned in the introduction, observational studies reveal that the NAO and the Atlantic meridional mode act independently (e.g., Ruiz-Barradas et al. 2000). Hence, the reduction in the variability of the meridional mode, as seen in the WES-off–SOM run, most likely results from the lack of the WES feedback.

In the WES-off–SOM experiment, the leading SVD mode of SST and \( Q_{lh} \) still explains 48% of their squared covariance, and their expansion coefficient time series exhibit a simultaneous correlation of 0.8, suggesting that the two fields are still strongly coupled, although with considerably weakened amplitudes. In addition, the lagged cross-correlation structure is also symmetric around zero lag (Fig. 3a), similar to the CCM3–SOM integration, suggesting the presence of a positive feedback between SST and \( Q_{lh} \) even in the absence of wind-induced fluctuations. The increased lagged correlations between the expansion coefficient time series of SST and \( Q_{lh} \) in the WES-off–SOM run, as compared to the CCM3–SOM run, is probably caused by the filtering of the influence of high-frequency atmospheric noise induced by the winds on SST and \( Q_{lh} \) in the WES-off–SOM run, as is noted in the reduction of interannual variability of the expansion coefficient time series of SST and \( Q_{lh} \) (Figs. 3b,c). By experiment design, the winds can still respond to CESGs in the WES-off–SOM run, and this response is depicted in the regression of surface winds on the expansion coefficient time series of SST of the leading SVD mode (Fig. 2b).

b. Near-surface humidity feedback

The above results imply that although the WES feedback amplifies the meridional mode in the deep tropics, it is not solely responsible for the generation of the meridional mode, since a weaker version of the mode is still noted in the WES-off–SOM run. What causes this residual dipole pattern in the WES-off–SOM run?

To answer this question, we partition \( Q_{lh} \) anomalies associated with the dipole mode. The SVD analysis reveals that the correlation between SST and \( Q_{lh} \) is strong only over the deep tropical regions, which is a smaller domain to the domain used in the SVD analysis. Here, we use a dipole SST index based on the SST difference between the cross-equatorial centers exhibiting the largest covariability in the SVD analysis. The dipole index is defined as the monthly area-averaged SST over 5°–15°N, 50°–20°W minus that over 5°–15°S, 35°–10°W. Similar indices have also been used in other studies (e.g., Carton et al. 1996; Chang et al. 1997). A dipole pattern is seen in the \( Q_{lh} \) anomalies, similar to that observed in the SVD analysis, with poles in the deep tropical Atlantic extending from the equator to about 10° latitudes. The \( Q_{lh} \) anomalies, if linearly partitioned, appear to come from two terms as follows:

\[
Q'_{lh} = -[u^{\prime\prime}(\Delta q) + \overline{u^\prime}(\Delta q)]B, \tag{4}
\]

where the primes denote the monthly anomalies and the bars denote climatological means, and \( B \) represents the climatological values of \( C_p\rho L_{vap} \). The contribution of each of these terms is shown in Figs. 4c,d, where the terms are individually regressed on the SST dipole index. A dipole pattern analogous to that observed in \( Q_{lh} \) is observed for the term. A dipole pattern, albeit with opposite polarity to that of the \( Q_{lh} \) pattern, is observed for the \( u^{\prime\prime}(\Delta q) \) term. The coherence of the pattern of the \( u^{\prime\prime}(\Delta q) \) term with the \( Q_{lh} \) anomalies over the deep tropics indicates that the term dominates in contribution to \( Q_{lh} \), whereas \( \overline{u^\prime}(\Delta q) \) acts to damp the anomalies, consistent with the result of Breugem et al. (2007). So, whereas the wind anomalies work to increase the SST on the northern side during the positive phase of the dipole mode, the anomalous development of \( \Delta q \) acts to cool down the SST, probably caused by the increase in saturation specific humidity above the ocean surface, which is linked to SST via the Clausius–Clapeyron equation.

Figure 4b shows the regression of \( Q_{lh} \) against the SST dipole index for the WES-off–SOM run. The dipole structure is still seen, with a comparable magnitude to that of the CCM3–SOM run. The anomalies in \( Q_{lh} \) in the WES-off–SOM are caused only by the second term in Eq. (4) and hence, all of the fluctuations of the \( Q_{lh} \) are brought about by changes in \( \Delta q \). Figure 4e shows the regression of \( \Delta q \) against the dipole index for the CCM3–SOM integration. A dipole pattern coherent with the contribution of the \( \overline{u^\prime}(\Delta q) \) term, as per the sign convention in Eq. (1), is observed. However, in the WES-off–SOM run, opposite polarities of the \( \Delta q \) pattern (Fig. 4f)
are observed as compared to the control run (Fig. 4e). The polarity of $\Delta q$ in the WES-off–SOM run is in accord with the $Q_{lh}$ pattern seen in Fig. 4b.

It appears then in the WES-off–SOM run that humidity and SST are reinforcing each other, with an increase in $\Delta q$ associated with an increase in SST. Hence, in the absence of wind-induced warming, the above suggests that the positive feedback exhibited between SST and $Q_{lh}$ materializes through the modulation of the air–sea humidity difference. Air–sea humidity difference has also been found to be important in other studies of tropical Atlantic variability (Chikamoto and Tanimoto 2005; Saravanan and Chang 2000). Chikamoto and Tanimoto (2005) analyzing reanalysis data and in situ observations find that an increase in atmospheric humidity induces 64% of the latent heat flux anomaly during warm ENSO events over the Caribbean Sea in the month of January, when the wind anomalies are weak. Their result is similar to our finding that over the equatorial Atlantic, surface humidity can dominate when the wind anomalies are constrained to zero by experiment design.

Moist convection is stronger over warmer equatorial SSTs. A cross-equatorial SST dipole would hence result in increased moist convection over the warmer hemisphere. The boundary layer specific humidity, which is tied to moist convection to maintain deep convective quasi-equilibrium, would thus increase with the increase in moist convection (Brown and Bretherton 1997). Over the cooler side, where deep convection decreases, the associated anomalous subsidence decreases the boundary
layer humidity. The dipolar pattern of \( \Delta q \) associated with the SST dipole in the WES-off–SOM run supports the argument. While the boundary layer specific humidity also increases (decreases) accordingly in the CCM3–SOM run where SST increases (decreases), the amplified SST increase (decrease) in CCM3–SOM also results in exponential increases (decreases) in \( q_s \), resulting in the opposite polarity of \( \Delta q \) (defined as \( q - q_s \)) as compared to the WES-off–SOM run. In the WES-off–SOM run, however, cross-equatorial \( q_s \) changes are limited because of the weaker SST dipole, and \( q \) is allowed to play a dominant role in \( \Delta q \) and hence \( Q_{lh} \). The above mechanism, while plausible, suggests little interaction between the two hemispheres in the generation of anomalous SST and the \( Q_{lh} \) dipole over the tropical Atlantic.

The cross-equatorial dipolar pattern of \( \Delta q \) in the WES-off–SOM run is, however, also suggestive of a cross-equatorial advective tendency of moisture. A regression of monthly meridional moisture advection (\( \nu q \), where \( \nu \) represents meridional wind speed) of near-surface humidity on the monthly SST dipole index (Fig. 5) shows a large northward transport of surface moisture across the equator. An additional mechanism, involving this cross-equatorial advection of moisture, is hypothesized here to explain the positive feedback between \( Q_{lh} \) and SST in the WES-off–SOM run, where \( q \) anomalies rather than \( q_s \) anomalies are allowed to dominate \( \Delta q \) anomalies over the tropical Atlantic. An initial cross-equatorial SST dipole anomaly would result in anomalous cross-equatorial winds and an initial anomalous dipole in \( \Delta q \), possibly generated as a response of moist convection to the SST dipole. The anomalous cross-equatorial winds (\( \nu' \)) would drive boundary layer moisture to the warmer hemisphere by anomalous advection at the rate of \( \nu'q \), increasing (decreasing) \( q \) in the warmer (cooler) hemisphere. The anomalous amplified \( q \) dipole results in an amplified dipole in \( \Delta q \) in the WES-off–SOM run, independent of the convective adjustment. The resulting dipole of anomalous \( Q_{lh} \) further enhances the cross-equatorial SST dipole, completing the positive feedback loop. By the above mechanism, the cross-equatorial meridional mode can thus also be generated by humidity in the absence of the WES feedback, albeit with a weaker amplitude.

The near-surface humidity feedback mechanism described above would be most effective when the axis of the SST dipole is concurrent with the ITCZ, under which lies the spatial maximum of near-surface mean specific humidity over the tropical Atlantic. The anomalous cross-equatorial winds would then advect moisture from under the ITCZ into the warmer hemisphere generating large dipolar anomalies. The positive feedback mechanism described above is illustrated as a schematic in Fig. 6. As a caveat, the anomalous advective tendency into the warmer hemisphere does not necessarily imply that the anomalous increase in \( q \) would be sustained long enough for the slab ocean to reach equilibrium, as moisture could be lost to atmospheric sinks like convection at time scales shorter than the response time of the slab ocean, which is about a few weeks.

**4. ENSO-forced Atlantic variability**

**a. SVD analysis: Tropical Atlantic**

The meridional mode of variability has been hypothesized to be the chosen mode for the response of the tropical Atlantic to ENSO forcing (e.g., Ruiz-Barradas et al. 2000). Figure 7a shows the spatial pattern of the leading SVD mode of monthly SST and \( Q_{lh} \) anomalies, after removing the annual cycle, over the tropical Atlantic for the CCM3–SOM–ENSO experiment, which explains 44% of the squared covariance. The regression of winds on the expansion coefficient time series of SST of the leading mode is also shown. A coherent dipole pattern of SST and \( Q_{lh} \) and winds similar to that of the CCM3–SOM integration is observed. The lagged correlation between the expansion coefficient’s time series of SST and \( Q_{lh} \) of the leading SVD mode exhibits a symmetric structure about lag zero (not shown), and it is indicative of a positive feedback between SST and \( Q_{lh} \). The coherent structure of SST, \( Q_{lh} \), and winds and the positive feedback between the SST and \( Q_{lh} \) in the CCM3–SOM–ENSO experiment supports the role of WES feedback in the meridional mode response of the tropical Atlantic to the Pacific ENSO forcing.

In the absence of the WES feedback, the response of the Atlantic in the WES-off–ENSO run, from a similar analysis of the leading SVD mode, is found to exhibit...
little coupled cross-equatorial structure (Fig. 7b) in the SST and $Q_{in}$ field as compared to the CCM3–SOM–ENSO run. The dipole pattern, which is a free mode of variability even in the absence of WES feedback, ceases to exist in the WES-off–ENSO experiment. The effect of humidity, capable of generating a dipole mode in the absence of external forcings, appears to be overshadowed by the effects of ENSO forcing. In response to ENSO, the boundary layer humidity increases across the tropical Atlantic (section 4). While tropospheric subsidence causes the drying of the boundary layer over the equatorial Atlantic following ENSO (Klein et al. 1999), the TT mechanism suggests that in response to tropospheric warming, the surface humidity increases in convective regimes (Chiang and Sobel 2002). In the absence of the WES feedback, these two forcings appear to damp out any dipolar anomalies of boundary layer humidity, preventing the near-surface humidity feedback to act to amplify any dipolar SST anomalies.

In the context of the role of winds in the response of the tropical Atlantic to ENSO forcings, it should be noted that our experiment design, in addition to shutting off the WES feedback, also prevents the SST from responding to externally forced winds. Hence, the lack of a wind-induced meridional mode in the WES-off–ENSO run is caused by both the lack of deep tropical WES feedback and the lack of the forcing of north tropical Atlantic SST through the modulation of trade winds, for example, by the PNA extension into the North Atlantic (Nobre and Shukla 1996). While we cannot distinguish between the two mechanisms here, our experiment clarifies the importance of the role of winds in the tropical response of the Atlantic to ENSO, confirming the results of other studies (e.g., Curtis and Hastenrath 1995; Alexander and Scott 2002). The observed meridional mode response to ENSO thus cannot be reproduced without the influence of winds, at least in the CCM3 coupled to a SOM.

A statistical analysis of the CCM3–SOM–ENSO and WES-off–ENSO runs is performed to explore the role of the WES feedback in the response of the tropical Atlantic to ENSO in the following sections.

b. Regression analysis: CCM3–SOM–ENSO run

The mechanism of the meridional mode response of the tropical Atlantic to ENSO events is studied using lagged regression analysis of SST and atmospheric variables against the model Niño-3 index. A regression of Atlantic SST averaged against the Niño-3 index for January shows statistically significant warm responses in all of the northern tropical Atlantic up to 20°N, strengthening progressively following the mature phase of ENSO and displaying a maximum response in April–June, peaking.
at about 0.4 K off the coast of West Africa (Figs. 8a,e). The simulated Atlantic response of CCM3–SOM compares well with observations in the tropical Atlantic (Lau and Nath 2001; Wallace et al. 1990; Deser and Blackmon 1993).

A strong meridional SST gradient response slightly south of the equator is observed in Fig. 8e, with warming in the northern equatorial Atlantic and a mild cooling in the southern side even in the absence of ocean dynamics. An anomalous SST response in the slab ocean model, a result only of changes in the surface heat fluxes, lags the surface heat fluxes by a few weeks because of the thermal capacity of the mixed layer depth. Figure 9a shows the regression of the net surface heat flux over the tropical Atlantic against the January Niño-3 index for the January–March period, which results in the SST anomalies in the April–June period (Fig. 8e). A decomposition of the contribution of individual surface heat flux components—latent, sensible, shortwave, and longwave heat fluxes—to the net heat flux during January–March reveals that \( Q_{lh} \) (Fig. 9b) is dominant in the Atlantic. Sensible and radiative heat fluxes (shortwave and longwave heat fluxes) play a secondary role over the tropical Atlantic in response to ENSO (Figs. 9c,d).

A dipole pattern in \( Q_{lh} \) anomalies over the deep tropical Atlantic is observed as suggested by the SVD analysis (section 3a). The anomalous surface pressure developments and their associated anomalous flow during warm ENSO events (Figs. 8b,f) decrease surface wind speeds over most of the north tropical Atlantic, working
FIG. 8. Regression of (a),(c) SST (contour interval: 0.1 K), (b),(f) surface pressure (contour interval: 5 Pa) and winds, (c),(g) surface humidity (contour interval: 0.1 g kg$^{-1}$), and (d),(h) convective precipitation (contour interval: 0.5 mm day$^{-1}$) averaged over (a)–(d) January–March and (e)–(h) April–June, against the January Niño-3 index for the CCM3–SOM–ENSO integration. Shaded areas represent statistically significant responses at the 95% confidence level based on a two-tailed $t$ test. Only wind anomalies statistically significant at the 95% level based on a two-tailed $t$ test are shown in (b) and (f).
to reduce $Q_{lh}$ and to increase the wind speed over the equatorial south Atlantic, increasing the $Q_{lh}$ (Fig. 9b), suggestive of the existence of the WES feedback as noted in other studies (e.g., Alexander and Scott 2002). A partition of $Q_{lh}$ anomalies [Eq. (3)] as shown in Figs. 9e,f also reveals that $u^*$ is actually dominant in contributing to $Q_{lh}$ anomalies over the tropical Atlantic, consistent with other studies (Enfield and Mayer 1997; Alexander and Scott 2002), with anomalies in $\Delta q$ largely opposing the effect of changes in $u^*$ on $Q_{lh}$.

A regression of the simulated convective precipitation averaged over January–March and April–June against the Niño index (Figs. 8d,h) exhibits the development of a strong cross-equatorial dipole, with increased (decreased) precipitation in the northern (southern) side, consistent with the CESG response. This anomalous dipole of precipitation in the boreal spring following ENSO is also seen in observations (e.g., Chiang et al. 2002; Saravanan and Chang 2000) and is noted in other GCM studies (e.g., Saravanan and Chang 2000). The northward shift of convergence during El Niño events has been linked to CESGs via anomalous cross-equatorial winds in observations (Nobre and Shukla 1996; Chiang et al. 2002), and it is also supported in our simulation, as seen in the cross-equatorial wind response (Figs. 8b,f). The ITCZ, modulated by the CESGs, displays the largest variability during this part of the year (reaching its southernmost extreme in April and northernmost extreme in August in

**FIG. 9.** Regression of January–March averaged (a) net surface heat flux, (b) $Q_{sh}$, (c) $Q_{o}$, and (d) net surface radiative flux against the January Niño-3 index for the CCM3–SOM–ENSO integration. The $Q_{lh}$ anomalies caused by fluctuations in (e) $u^*$ and (f) $\Delta q$ in the January–March period during ENSO events for the CCM3–SOM–ENSO integration. Shaded areas represent statistically significant responses at the 95% confidence level based on a two-tailed $t$ test. Negative contours indicate warming of the ocean mixed layer. Contour interval is 1.5 W m$^{-2}$. 

Downloaded from http://journals.ametsoc.org/jcli/article-pdf/23/22/5958/3971175/2010jcli3304_1.pdf by guest on 21 September 2020
observations; e.g., Xie and Carton 2004), and hence responds to the ENSO-induced CESG sensitively. The reduction in convective precipitation, as seen over the south equatorial Atlantic in our simulation, following El Niño, is also believed to be caused by subsidence associated with the anomalous Walker circulation (e.g., Saravanan and Chang 2000; Chiang et al. 2002).

A northward amplifying gradient is also seen in the near-surface humidity response (Figs. 8c,g). Previous studies suggest that the tropical Atlantic surface humidity response to ENSO could be caused by the TT mechanism (Chiang and Sobel 2002) and by the modified Walker circulation associated subsidence (Saravanan and Chang 2000; Klein et al. 1999). In the northern side, where the ITCZ is located, the TT mechanism—believed to be most effective in a convective environment—appears to dominate, causing an increase in surface humidity, whereas in the southern side, subsidence overshadows the effect of the TT mechanism on boundary layer humidity. Anomalous cross-equatorial winds could also generate a cross-equatorial gradient of near-surface humidity (Figs. 8c,g) by anomalous surface moisture advection ($\nu'q$).

In the April–June period, when ENSO-associated SST anomalies peak, the net surface heat flux anomalies (Fig. 10a) work to damp the tropical Atlantic SST anomalies developed during the previous months, resulting in a weaker SST response over the July–September period (Fig. 11). A decomposition of the net heat flux reveals

![Diagram of heat fluxes](image-url)
the meridional mode. Forcing of the Atlantic Niño, dominant in the summer until winter (Okumura and Xie 2006), by ENSO events (Chang et al. 2006), is also absent in the SOM. Furthermore, Lee and Wang (2008) suggest that the weakening of the meridional mode is related to the strengthening of the equatorial thermocline slope, a relation missing in the SOM. The meridional mode hence is probably sustained in the region in the SOM during ENSO because of the lack of such ocean dynamics, allowing the thermodynamic interaction of atmosphere-ocean system to dominate and generate its own modes of variability.

The amplification of cross-equatorial gradients of tropical SST and atmospheric response after the mature phase of ENSO suggests that the meridional mode, mechanized by the WES feedback, is the preferred mode of response of the tropical Atlantic to ENSO. In the next section, we analyze the response of the tropical Atlantic to ENSO in the WES-off–ENSO run, to isolate the role of the WES feedback in the meridional mode and to understand how other mechanisms manifest themselves in its absence.

c. Regression analysis: WES-off–ENSO run

Figure 12 shows the response of the tropical Atlantic Ocean to the artificial ENSO forcing in the Pacific in the WES-off–ENSO experiment in the January–March and April–June periods. Peak anomalous warming of about 0.2 K is observed in tropical Atlantic in the April–June period (Fig. 12e). The response is weaker than that observed in the CCM3–SOM–ENSO experiment over the tropical Atlantic. The weak response of tropical Atlantic SST as compared to the CCM3–SOM–ENSO experiment is due to the weaker net surface fluxes in the experiment, as seen in Fig. 13a.

The weak tropical SST response and the weaker response of the surface latent heat flux (Fig. 13b), surface winds, near-surface humidity, and precipitation (Fig. 12) lacks the cross-equatorial gradient as observed in the CCM3–SOM–ENSO experiment, suggesting that the response bears little relation to the meridional mode in the absence of the role of winds on surface fluxes as suggested by the SVD analysis (section 3a). Hence, in the absence of the WES feedback, and the lack of the influence of modulated North Atlantic trade winds on SST, the response of ENSO clearly does not manifest as a meridional mode in the tropical Atlantic. However, the tropical North Atlantic trade winds response itself to ENSO is weaker in the WES-off–SOM run as compared in the CCM3–SOM. This weaker trade wind response could be due to the weaker atmospheric response of the WES-off–SOM run by experiment design and by the lack of amplification of the trade winds by the WES feedback. As shown in Fig. 1, the atmospheric response of WES-off–ENSO should be

![Figure 11](https://example.com/figure11.png)
only slightly weaker than that of the CCM3–SOM–ENSO run. Studies (Enfield and Mayer 1997; Xie and Carton 2004; Sutton et al. 2000) suggest that ENSO-modulated trade winds in the tropical Atlantic region are amplified by the WES feedback. Also, our experiment design does not specifically constrain the winds, implying that the ENSO-forced wind response minus the effects of the WES feedback on the winds should be the same in the WES-off–ENSO

FIG. 12. As in Fig. 8, but for the WES-off–ENSO integration.
and CCM3–SOM–ENSO integrations. Hence, a large part of the weakening of the trade wind response in the WES-off–SOM run appears be caused by the lack of amplification by the WES feedback.

The surface latent heat flux plays a stronger role in the net surface flux over the tropical Atlantic in the WES-off–ENSO run (Fig. 13b) with $Q_{sh}$ and radiative heat fluxes playing a weaker role (Figs. 13c,d) in the January–March period. Since we disallow the interannual variance of wind speed to influence $Q_{lh}$, the change in $Q_{lh}$ in response to ENSO events is caused almost completely by a change in $D_q$. An increase in boundary layer humidity throughout the tropical Atlantic is observed (Figs. 12c,g), with the largest response in the north equatorial regions where climatological deep convection occurs. The TT mechanism (Chiang and Sobel 2002) maintains that free tropospheric warming forced by warm ENSO events modulates boundary layer humidity via convection to homogenize moist static energy in the tropospheric column (Brown and Bretherton 1997). The change in boundary layer humidity then modulates surface temperature by evaporation. The TT mechanism is believed to be most effective in convective regimes (Chiang and Sobel 2002). The near-surface humidity response in the WES-off–ENSO experiment appears to be consistent with that predicted by the TT mechanism. The surface temperature response over the tropical Atlantic is also consistent with the change in near-surface humidity. Thus, in the absence of the meridional mode, which generates cross-equatorial gradients, the TT mechanism appears to play a dominant role over the tropical Atlantic in response to ENSO and generates a more meridionally symmetric response. Chikamoto and Tanimoto (2006) also show that in observations, the near-surface humidity anomalies over the tropical Atlantic associated with ENSO tend to induce an equatorially symmetric SST response. Convective precipitation response over the Atlantic (Figs. 12d,h) is primarily that of decline caused by the subsidence associated with the change in Walker circulation (e.g., Klein et al. 1999).

The collocation of oceanic–atmospheric anomalies in the CCM3–SOM–ENSO integration and the weaker coupled response in the WES-off–ENSO response conclusively indicates that the WES feedback is indeed directly amplifying the SST and atmospheric response to ENSO events in the tropical Atlantic in the CCM3–SOM–ENSO simulations in the January–June period.

5. Summary and discussion

A SVD analysis of the CCM3 coupled to a slab ocean model over the tropical Atlantic indicates a cross-equatorial dipole pattern of oceanic and atmospheric anomalies as seen in observational studies (Chang et al.

---

**FIG. 13.** Regression of January–March averaged (a) net surface heat flux, (b) $Q_{sh}$, (c) $Q_{sh}$, and (d) net surface radiative flux against the January Niño-3 index for the WES-off–ENSO integration. Shaded areas represent statistically significant responses at the 95% confidence level based on a two-tailed $t$ test. Negative contours indicate warming of the ocean mixed layer. Contour interval is 1.5 W m$^{-2}$. 

---
1997; Ruiz-Barradas et al. 2000), even in the absence of remote forcings such as ENSO. This implies that thermodynamic interactions between the atmosphere and the ocean have the potential to organize the intrinsic atmospheric variability to arrange anomalies in a pattern that resembles the meridional mode. This result is similar to the findings of Chang et al. (2001), that in the presence of moderate air–sea coupling and atmospheric noise, it is possible to generate a dipole-like decadal oscillation in a hybrid coupled model. Decadal oscillations are not identifiable in the CCM3–SOM integrations because of the lack of ocean dynamics as mentioned in the introduction. We plan to address the issue of the role of the WES feedback in decadal oscillations in the tropical Atlantic by conducting a similar study using an atmospheric GCM coupled to a dynamical ocean model. Here, we test the hypothesis that the arrangement into a dipole-like pattern of the tropical Atlantic anomalies is caused by the WES feedback as suggested by other studies (Chang et al. 1997, 2000) and also suggested by the collocation of coherent SST, latent heat flux, and wind anomalies in the CCM3–SOM integrations. A comparison between a CCM3–SOM integration where the WES feedback is switched off and the control run indicates that the free meridional mode of tropical Atlantic variability is indeed enhanced in the presence of the WES feedback at low frequencies. However, the WES feedback is found not to be essential in the generation of the dipole mode itself. Specific humidity difference between the atmosphere and the ocean surface (Δq) is capable of independently generating a dipole mode but with a substantially lower interannual variance through a feedback mechanism in the WES-off–SOM run. However, in the presence of the WES feedback, the mechanism appears to fail because of stronger SST anomalies that result in strong changes in saturation specific humidity, nullifying the effect of anomalous advection of moist air and in fact working to damp the SST anomalies, as is seen in the CCM3–SOM simulation. The result here thus indicates a stronger role for humidity in tropical Atlantic climate variability as suggested by Breugem et al. (2007), albeit in the absence of the influence of winds on SST.

It should be noted that intrinsic tropical Atlantic atmospheric variability in our CCM3–SOM simulations includes the influence of NAO. This might imply a fair chance for the CCM3–SOM Atlantic meridional mode to have been generated as a forced response to the NAO rather than be a part of natural tropical Atlantic variability. And, that the weaker meridional mode in the WES-off–SOM is a result of the absence of the influence on NAO-induced winds on North Atlantic trade winds. However, recent studies have found little evidence of NAO influencing deep tropical Atlantic variability (Okumura et al. 2001; Rajagopalan et al. 1998; Ruiz-Barradas et al. 2000), suggesting that most likely it is the absence of the WES feedback that leads to the reduced magnitude of the Atlantic meridional mode in the WES-off–SOM integration.

Under forced ENSO conditions, the role of WES feedback in modulating tropical Atlantic meridional mode variability becomes evident in our CCM3–SOM experiments forced by an ENSO-like SST cycle in the tropical Pacific. A comparison of leading SVD patterns of SST and latent heat flux over the tropical Atlantic, between ENSO-forced runs where the WES feedback is allowed to exist and where the WES feedback is switched off, reveals that the coupled response of the Atlantic resembles the meridional mode only in the presence of the WES feedback. Otherwise, no discernible coupled response pattern is observed. In the presence of remote forcings like ENSO, humidity alone fails to sustain the meridional mode and tropical variability appears to be governed by other processes in the WES-off–ENSO runs. Penland and Matrosova (1998) find that ENSO disrupts the Atlantic dipole. Our modeling analysis suggests that the disruption occurs if the Atlantic dipole is solely governed by humidity. In the presence of the WES feedback, however, the response of the Atlantic to ENSO events occurs through the meridional mode, even enhancing the mode, as is found in other studies (e.g., Ruiz-Barradas et al. 2000; Saravanan and Chang 2000).

Regression of the ENSO index on Atlantic SSTs alone reveals meridional gradient (Fig. 8e) and not a clear meridionally antisymmetric response. This result is consistent with the study of Enfield and Mayer (1997), who found only a mild asymmetry in observations of Atlantic SSTs rather than a cross-equatorial dipole. The appearance of a dipole-like pattern in coupled variability of free and forced tropical Atlantic climate (e.g., Ruiz-Barradas et al. 2000; Chang et al. 1997) and a lack of the same in independent analysis of SSTs (e.g., Mehta 1998; Enfield et al. 1999) is consistent with other studies, as coupled dipolar configurations only reveal the bivariate relationships between SSTs and atmospheric variables (Enfield et al. 1999). The development of anomalous meridional SST gradients in the CCM3–SOM–ENSO simulations over the tropical Atlantic in response to ENSO is found to be critically dependent on the WES feedback, in the absence of which the meridional gradient is found to be much weaker, consistent with the SVD analysis. The anomalous cross-equatorial flow associated with the SST meridional gradient also causes the ITTCZ to move northward over the Atlantic during ENSO events in the CCM3–SOM–ENSO run, consistent with observations and modeling studies (e.g., Moura and Shukla 1981; Saravanan and Chang 2000; Chiang et al. 2002) but...
not in the WES-off–SOM run. As a caveat, our experiment design also removes the influence of winds forced by the PNA teleconnection to influence the SST in the north tropical Atlantic, which also contributes to the CESGs, in the WES-off–ENSO integration. However, tropical Atlantic anomalies are still generated in the WES-off–ENSO integration by other mechanisms; but these anomalies form a weaker meridional gradient.

Our results hence support the hypothesis that the WES feedback amplifies the cross-equatorial SST and atmospheric gradients over the tropical Atlantic during ENSO events proposed in other studies (Enfield and Mayer 1997; Chiang et al. 2002; Sutton et al. 2000). Without the feedback, the tropical Atlantic surface response is more symmetric about the equator. In the absence of the WES feedback, which forces the generation of cross-equatorial gradients, the TT mechanism (Chiang and Sobel 2002) manifests itself more clearly. The global free tropical troposphere warms following El Niño events (e.g., Sobel and Bretherton 2000). The TT mechanism maintains that the moist static energy exchange between the boundary layer and the warm free tropospheric column above results in an increase in boundary layer humidity during El Niño events, and that it is most effective in the deep convective regions. The increased boundary layer humidity warms the SST below caused by the reduction in surface evaporation. In the WES-off–ENSO run, the boundary layer humidity increases across the tropical Atlantic in response to warm ENSO events, with the strongest anomalies occurring over the deep convective regions, clearly indicative of an active TT mechanism.

The variability of the CESGs that is associated with the meridional mode has widespread societal effects influencing rainfall in northeast Brazil (e.g., Moura and Shukla 1981) and northwestern Africa (e.g., Folland et al. 1986). Recent evidence also suggests the influence of the CESGs on the NAO (e.g., Rajagopalan et al. 1998) and the tropical Pacific (e.g., Zhang and Delworth 2005; Wu et al. 2007). The oscillatory nature of the meridional mode provides potential predictability (Chang et al. 1997), the tapping of which requires an improved understanding of the mechanisms governing the variability. Further efforts are needed to represent these mechanisms correctly in GCMs, which demonstrate large variability in their representation of air–sea feedbacks (Wang and Carton 2003) and hence in the variability of the Atlantic, for improved seasonal predictions of the region.

Acknowledgments. This work was supported by research grants from NOAA’s CLIVAR Program (Project NA05AR4311136) and NSF’s Climate Dynamics Program (Grant ATM-0337846). We are grateful for the suggestions by three anonymous reviewers.

REFERENCES


Biasutti, M., 2000: Decadal variability in the tropical Atlantic as simulated by the Climate System Model and the CCM3 coupled to a slab ocean model. M.S. thesis, Dept. of Atmospheric Sciences, University of Washington, 43 pp.


Doi, T., T. Tozuka, and T. Yamagata, 2009: Interannual variability of the Guinea Dome and its possible link with the Atlantic


