Wind–Mixed Layer–SST Feedbacks in a Tropical Air–Sea Coupled System: Application to the Atlantic

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ABSTRACT

The ocean–atmosphere feedback associated with the thermodynamic coupling among wind speed, evaporation, and sea surface temperature (SST), called the wind–evaporation–SST (WES) feedback, contributes to the cross-equatorial SST gradient over the tropical oceans. By conducting an eigenanalysis of simple linear air–sea coupled models, it is shown that two additional feedback processes are present when the variable oceanic mixed layer depth (MLD) is considered. The horizontal structures of the leading modes are similar to the WES mode, which shows a meridional dipole in the SST anomalies straddling the equator with cross-equatorial wind anomalies that represent the weakening/strengthening of the trade winds over the warm/cool SST anomalies. The coupling of the variable MLD with winds and SST more than doubles the growth rate of the WES mode and enhances the equatorward propagation of the coupled disturbances. The identified feedbacks operate as follows: The weaker winds associated with warm SST anomalies shoal the mixed layer through suppressed turbulent mixing, which causes the mixed layer to be more sensitive to the climatological shortwave radiation and amplifies the initial positive SST anomalies. Likewise, deepening of the mixed layer due to stronger winds acts to maintain the negative SST anomaly on the other side of the dipole. The MLD anomalies can also be generated by the buoyancy flux anomaly related to the wind-induced latent heat flux anomaly. The antiphase relationship between the SST and MLD anomalies seen in the simple model bears some resemblance to that which is observed in the observations and a state-of-the-art coupled model during the Atlantic meridional mode.

1. Introduction

The tropical climate system hosts ocean–atmosphere coupled processes, some of which give rise to large-scale climate fluctuations. Among others, the wind–evaporation–sea surface temperature (SST) (WES) feedback (Xie and Philander 1994) is known to force or sustain a cross-equatorial SST gradient (hereafter, referred to as a meridional mode) over the tropical oceans (Chang et al. 1997; Xie 1999; Chiang and Vimont 2004; Wu et al. 2008). While the tropical Pacific Ocean variability is dominated by El Niño–Southern Oscillation,

which is closely associated with the zonal gradient, the meridional interhemispheric gradient characterizes the tropical Atlantic variations on interannual to decadal time scales (Nobre and Shukla 1996; Delworth and Mehta 1998; Tanimoto and Xie 1999; Chiang and Vimont 2004). The WES feedback contributes to the anomalous interhemispheric gradient as follows: The meridional SST anomaly gradient is accompanied by an opposite-signed meridional SLP gradient in the overlying atmosphere, which forces the atmospheric surface flow toward the warmer hemisphere; the Coriolis force veers the surface flow to the right (left) in the Northern (Southern) Hemisphere, and hence modulates the easterly trade winds and amplifies the initial SST anomalies by changing the latent heat flux (Fig. 1a).

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FIG. 1. Schematics of (a) the wind–evaporation–SST (WES) feedback, (b) the wind-induced turbulence–mixed layer–SST (WIMS) feedback, and (c) the wind–evaporation–mixed layer–SST (WEMS) feedback.

While the meridional mode does not appear self-sustaining and requires external forcing, the WES feedback plays an important role in generating or sustaining SST anomalies in the deep tropics (Xie and Tanimoto 1998; Saravanan and Chang 2000; Chang et al. 2000; Frankignoul and Kestenare 2005; Czaja et al. 2002; Chiang et al. 2002; Chiang and Vimont 2004; Huang and Shukla 2005; Vimont et al. 2009). As well as affecting the surrounding climate (Vimont and Kossin 2007; Zhang et al. 2019), the tropical meridional modes may also influence the climate at higher latitudes (Watanabe and Kimoto 1999; Okumura et al. 2001) and in other ocean basins (e.g., Ham et al. 2013; Wang et al. 2017). Therefore, understanding the underlying ocean–atmosphere coupled processes is of great importance as they may provide a source of longer-term predictability.

Although the WES concept implicitly assumes that the oceanic mixed layer depth (MLD) is constant, it is known that variations in the MLD play a role in SST variability (e.g., Alexander et al. 2000; Morioka et al. 2010; Doi et al. 2010; Graham et al. 2014; Yamagami and

Tozuka 2015; Kataoka et al. 2017; Ohishi et al. 2017; Tozuka et al. 2018). For example, the shallower (deeper) MLD contributes to the warmer mixed layer temperature (approximately equal to SST) by changing the sensitivity to the positive climatological surface heat flux (i.e., the shortwave radiation; heat flux into the ocean is defined as positive) if everything else remains the same. Such a process is likely more efficient in the tropics as the shortwave radiation tends to be larger at lower latitudes. As explained above, the WES is accompanied by a reduction in wind speed in the warmer hemisphere. This reduced wind speed could contribute to a shallower ocean mixed layer because of the weaker wind stirring. In addition, the reduced wind speed causes a reduction in the latent heat loss, which directly forces the SST anomaly in the WES framework. However, the reduction of the latent heat loss means a buoyancy gain of the ocean and therefore, could make the MLD even thinner. With this in mind, the following two ocean–atmosphere feedback processes could exist in addition to the canonical WES feedback:

- (i) Similar to the WES feedback, the anomalous meridional SST gradient modulates the trade winds; the weaker (stronger) surface winds contribute to the shallower (deeper) ML via suppressed (enhanced) turbulent mixing, which results in an ML that is more (less) sensitive to the warming by the climatological shortwave radiation and thus amplifies the initial SST anomalies (Fig. 1b). In the following, we call this possible feedback mechanism the wind-induced turbulence–mixed layer–SST (WIMS) feedback.
- (ii) The second process is similar to (i) except the latent heat flux is modulated by the surface wind anomalies. The larger (smaller) latent heat flux into the ocean in the warmer (cooler) hemisphere contributes to the shallower (deeper) ML because of the buoyancy gain (loss) of the ocean, which results in an ML that is more (less) sensitive to the climatological shortwave warming and amplifies the initial SST anomalies (Fig. 1c). This possible feedback is referred to as the wind–evaporation–mixed layer–SST (WEMS) feedback mechanism in this paper.

In the present study, we show that the above two processes, the WIMS and WEMS feedbacks, exist in a tropical ocean–atmosphere coupled system and they are as strong or stronger than the canonical WES feedback. Since it is difficult to distinguish among the above feedback processes in the observations and general circulation models (GCMs), here we construct a simple air–sea coupled model for each feedback process. Because our main focus is more on understanding of each feedback process than how these feedbacks contribute

to the meridional mode growth, we then conduct an eigenanalysis rather than seek the nonnormal amplifications (e.g., Vimont 2010). As we are interested in whether the eigenmodes are accompanied by a cross-equatorial gradient, the eigenmodes are examined under an equatorially symmetric mean state to avoid eigenmodes having a basic-state-origin meridional gradient. For this purpose, the Atlantic conditions during the boreal spring, when the climatology is close to symmetric about the equator (Xie and Carton 2004), are adopted. We note that this is also the season in which the Atlantic meridional mode develops (Fig. A2e).

The rest of the paper is organized as follows. A brief explanation of the observational and reanalysis data and a coupled GCM is given in section 2. In section 3, simple coupled models are constructed. In section 4, the eigenvalue problem for a simple coupled system is solved to investigate the positive feedbacks, and the effects of the negative climatological heat fluxes are examined as well. Implications for the Atlantic meridional mode and the role of oceanic currents are discussed in section 5. The final section summarizes the main findings and discusses the possible roles of processes that were not considered in this study. While the primary aim of this study is to understand the tropical ocean–atmosphere coupled feedback processes, we expect that the processes proposed in this study contribute to tropical meridional modes.

2. Data and coupled general circulation model

We use the Centennial In Situ Observation-Based Estimates of SST, version 2 (COBE-SST2; Hirahara et al. 2014), from 1979 to 2016, the Global Precipitation Climatology Project (GPCP) (Adler et al. 2003) precipitation from 1979 to 2016, the European Centre for Medium-Range Weather Forecasts interim reanalysis (ERA-Interim; Dee et al. 2011) for wind at 10 m for the period 1979–2016, the net surface heat flux and short-wave radiation from the International Satellite Cloud Climatology Project (ISCCP; Zhang et al. 2004) for 1984–2009, and the Mixed Layer Dataset of Argo, Grid Point Value (MILA-GPV; Hosoda et al. 2010), for MLD for 2001–16. The MLD from the MILA-GPV is estimated using the Argo float data and is defined as the depth where the potential density increases by 0.03 kg m^{-3} from the surface.

The ocean–atmosphere coupled GCM (CGCM) used in this study is the Model for Interdisciplinary Research on Climate, version 6 (MIROC6; Tatebe et al. 2018, manuscript submitted to *Geosci. Model Dev. Discuss.*), developed collaboratively by the Japan Agency for Marine–Earth Science and Technology, the University

of Tokyo, and the National Institute for Environmental Studies. The horizontal resolution is T85 ($\sim 1.4^\circ$) for the atmospheric component, with 81 vertical levels up to 0.004 hPa, whereas the oceanic horizontal resolution is 360×256 ($\sim 1^\circ$). The oceanic vertical resolution is 63 levels down to a depth of 6300 m with layer thicknesses of about 2–8 m in the top 50 m (higher resolution closer to the surface). The turbulent mixing in the ocean is represented by the parameterization of Noh and Kim (1999) with a modification for the turbulent Prandtl number according to Noh et al. (2005). The radiative forcing is fixed at the preindustrial conditions and monthly mean outputs of a 50-yr integration after a spinup period of 2000 yr are analyzed. The daily mean potential density is calculated from the daily mean temperature and salinity based on the equation of state (UNESCO 1981) and the MLD is diagnosed with the same threshold as the MILA-GPV. Then, the monthly means of the diagnosed daily MLD are computed. The Atlantic climate simulated by MIROC6 is summarized in the appendix.

Linear trends are removed from the observational and reanalysis data and outputs from the CGCM using a least squares fit. When discussing the MLD, however, we retain trends to avoid artificial inversion layers.

3. Simple model

Since only variations in the meridional direction are essential for the potential ocean–atmosphere coupled processes as well as the canonical WES feedback, a zonally symmetric mean state is considered (e.g., Xie 1999). Hereafter, variables with and without an overbars denote the basic state and the anomaly field, respectively.

a. Atmosphere model

The Matsuno–Gill type model (Matsuno 1966; Gill 1980) is adopted for the atmosphere:

$$\frac{\partial U}{\partial t} + \varepsilon U - \beta y V = 0, \quad (1)$$

$$\frac{\partial V}{\partial t} + \varepsilon V + \beta y U = -\frac{\partial \varphi}{\partial y}, \quad \text{and} \quad (2)$$

$$\frac{\partial \varphi}{\partial t} + \varepsilon \varphi + C_a^2 \frac{\partial V}{\partial y} = -KT, \quad (3)$$

where U and V are the zonal and meridional components of the surface wind velocity anomalies, respectively, φ is the geopotential anomaly, C_a is the atmospheric long-wave speed, T is the SST anomaly, y is the meridional distance from the equator, ε is the Rayleigh damping rate, and K is the thermal coupling

FIG. 2. Meridional profiles of (a) the thermal coupling coefficient K (in $10^{-4} \text{ m}^2 \text{ s}^{-3} \text{ K}^{-1}$) and (b) the observed zonal average climatological shortwave radiation over $30^\circ\text{--}20^\circ\text{W}$ during February–April (black) and the approximation prescribed in the simple model (blue). (c)–(f) As in (b), but for the climatological net surface heat flux, SST, zonal wind speed at 10 m, and oceanic MLD, respectively. Note that the blue solid line in (f) represents the simulated mean MLD distribution in the simple model. Superimposed on (f) are the prescribed mean MLD profile (blue dashed) and the one simulated by MIROC6 (pink).

coefficient. The atmosphere is driven by the ocean through latent heating anomaly, which is parameterized to be proportional to the local SST anomaly. This atmospheric diabatic heating is confined to the equatorial region where the mean surface convergence occurs (Vimont 2010):

$$K = 2 \times 10^{-3} \exp[-(y/y_1)^4],$$

where $y_1 = 1000 \text{ km}$ (Fig. 2a).

The atmospheric model covers both hemispheres, for latitudes up to 4000 km from the equator. Even when the SST anomaly is also coupled with the atmospheric boundary layer (Lindzen and Nigam 1987; Neelin 1989), for example, by replacing the K profile with $K = 2 \times 10^{-3} \exp[-(y/y_1)^4] + 5 \times 10^{-4} \{1 - \exp[-(y/y_1)^4]\}$, the results obtained are qualitatively the same.

b. Ocean model

A motionless ocean mixed layer is considered but its depth varies spatiotemporally:

$$\begin{aligned} \frac{\partial T}{\partial t} &= \frac{\overline{Q} + Q}{\rho_o c_p (\overline{h} + h)} - \frac{\overline{Q}}{\rho_o c_p \overline{h}} \\ &\approx \frac{Q}{\rho_o c_p \overline{h}} - \frac{\overline{Q}}{\rho_o c_p \overline{h}} \frac{h}{\overline{h}}. \end{aligned} \quad (4)$$

Here, ρ_o is the seawater density, c_p is the specific heat of the seawater, h is the MLD, and Q is the net surface heat flux. Since we are primarily concerned with the WES, WIMS, and WEMS, only the latent heat flux anomaly is taken into account for the surface heat flux anomaly, whereas all components will be considered for the climatological surface heat flux:

$$\overline{Q} + Q = \overline{Q}_{\text{sw}} + \overline{Q}_{\text{lw}} + \overline{Q}_{\text{lh}} + \overline{Q}_{\text{sh}} + Q_{\text{lh}}. \quad (5)$$

Assuming that the background surface winds are dominated by strong easterlies, that is, $\overline{U} + U < 0$ and $|\overline{V}/\overline{U}| \ll 1$, the bulk formula for the latent heat flux can be linearized in the following way:

$$\begin{aligned}
\overline{Q_{\text{lh}}} + Q_{\text{lh}} &= -\rho_a LC_E |\overline{U} + U| [q(\overline{T} + T) - \text{RH}q(T_{\text{ref}})] \\
&\approx -\rho_a LC_E (1 - \text{RH}) |\overline{U} + U| q(\overline{T} + T) \\
&\approx \rho_a LC_E (1 - \text{RH}) \left(\overline{U} \overline{q} + U \overline{q} + \frac{L}{R^* \overline{T}^2} \overline{U} \overline{q} T \right).
\end{aligned} \quad (6)$$

The second term on the RHS is the latent heat flux anomaly due to the surface wind speed anomaly, and the third term is that due to the SST anomaly (a Newtonian cooling). Here, the saturated specific humidity is calculated from the Clausius–Clapeyron equation

$$q(\overline{T} + T) = q_0 \exp\{(L/R^*)[1/T_0 - 1/(\overline{T} + T)]\}. \quad (7)$$

Here, R^* is the gas constant, L is the latent heat, RH is the relative humidity, ρ_a is the air density, and C_E is the aerodynamic transfer coefficient. Since the net surface heat flux warms the ocean in the tropics, the MLD is estimated using the Monin–Obukhov depth (Qiu and Kelly 1993):

$$\begin{aligned}
\overline{h} + h &= \frac{\left[m(\overline{U}_* + U_*)^3 + \frac{\alpha g}{\rho_o c_p} \int_{-(\overline{h}+h)}^0 \overline{q_{\text{sw}}}(z) dz \right]}{\frac{\alpha g}{2\rho_o c_p} [\overline{Q} + Q + \overline{q_{\text{sw}}}(-\overline{h} - h)]} \\
&\approx \frac{\left[m(\overline{U}_* + U_*)^3 + \frac{\alpha g}{\rho_o c_p} \overline{Q_{\text{sw}}} \right]}{\frac{\alpha g}{2\rho_o c_p} (\overline{Q} + Q)}.
\end{aligned} \quad (8)$$

Here, α is the thermal expansion coefficient, C_D is the drag coefficient, and m is the wind mixing coefficient. For the second equality, we assume all of the shortwave radiation is absorbed in the ML and the penetrative shortwave radiation at the base of the ML is neglected. In reality, the deepening of the ML occurs even when the net surface heat flux tends to stabilize the upper ocean. However, as will be seen later, Eq. (8) provides a reasonable meridional profile of the MLD climatology in the tropical Atlantic despite its simplicity. Considering the dominant easterlies, the frictional velocity U_* is given by

$$\begin{aligned}
\overline{U}_* + U_* &= \sqrt{\frac{\rho_a C_D (\overline{U} + U)^2}{\rho_o}}, \\
&= -(\overline{U} + U) \sqrt{\frac{\rho_a C_D}{\rho_o}}
\end{aligned} \quad (9)$$

which gives the climatological MLD and its anomaly after linearization of Eq. (8) as follows:

$$\overline{h} \equiv \frac{\left[-m \left(\frac{\rho_a C_D}{\rho_o} \right)^{3/2} \overline{U}^3 + \frac{\alpha g}{\rho_o c_p} \overline{Q_{\text{sw}}} \right]}{\frac{\alpha g}{2\rho_o c_p} \overline{Q}}, \quad \text{and} \quad (10)$$

$$h \approx -\frac{3m\overline{U}_*^2}{\frac{\alpha g}{2\rho_o c_p} \overline{Q}} \sqrt{\frac{\rho_a C_D}{\rho_o}} \overline{U} - \overline{h} \frac{Q}{\overline{Q}}. \quad (11)$$

We note that Liu and Xie (1994) also consider the wind mixing effect. In their model, wind mixing effect may cause the mixed layer temperature change through anomalies in the vertical entrainment of cooler water, but the MLD change is not allowed. More precisely, they estimate the entrainment cooling anomaly from the one-dimensional turbulent kinetic energy balance with an assumption that the MLD is constant, while the MLD anomaly is estimated under the assumption of zero entrainment cooling in the present study.

From the first term on the RHS of Eq. (4) and the second term on the RHS of Eq. (6), the direct effect of the latent heat flux anomaly associated with variations in the wind speed on the SST change, that is, the WES term (Fig. 1a), is given by

$$\frac{1}{\overline{U}} \frac{\overline{Q_{\text{lh}}}}{\rho_o c_p \overline{h}} U \equiv a_1 U. \quad (12)$$

The contribution of the latent heat flux anomaly associated with the response to the SST anomaly is estimated from the first term on the RHS of Eq. (4) and the third term on the RHS of Eq. (6):

$$\frac{L}{R^* \overline{T}^2} \frac{\overline{Q_{\text{lh}}}}{\rho_o c_p \overline{h}} T \equiv -b_1 T. \quad (13)$$

Changes in the surface wind stirring cause the MLD anomaly, which alters the sensitivity of the ML to the surface heat flux. The sensitivity change effect on the anomalous SST differs among the climatological surface heat fluxes because the shortwave radiation warms the ocean, whereas the other fluxes (i.e., $\overline{Q_{\text{lw}}}$, $\overline{Q_{\text{lh}}}$, and $\overline{Q_{\text{sh}}}$) tend to cool it. For example, given the background easterlies, the westerly wind anomaly forced by a positive SST anomaly results in a shallower MLD, which enhances the warming by the climatological shortwave radiation. Therefore, the climatological shortwave radiation together with the MLD shoaling associated with the weaker wind stirring could act as a positive feedback. Such a positive air–sea coupled process, that is, the WIMS term (Fig. 1b), may be derived from the second

term on the RHS of Eq. (4) with $\overline{Q}$ replaced by $\overline{Q}_{\text{sw}}$ and the first term on the RHS of Eq. (11):

$$\frac{3m\overline{U}_*^2}{\frac{\alpha g}{\rho_o c_p h^2}} \sqrt{\frac{\rho_a C_D}{\rho_o}} \frac{\overline{Q}_{\text{sw}}}{\overline{Q}} U \equiv a_2 U. \quad (14)$$

On the other hand, since the other climatological heat fluxes cool the ocean, the MLD sensitivity change to them works as a negative feedback, which will be considered in section 4b.

The other feedback process we propose in this study, the WEMS (Fig. 1c), also involves the MLD anomaly, but it forms because of the latent heat flux anomalies related to variations in the surface wind speed. The positive feedback term related to the climatological shortwave radiation is constructed using the second term on the RHS of Eq. (4) with $\overline{Q}$ replaced by $\overline{Q}_{\text{sw}}$, the second term on the RHS of Eq. (6), and the second term on the RHS of Eq. (11), giving

$$\frac{\overline{Q}_{\text{sw}}}{\rho_o c_p h} \frac{\overline{Q}_{\text{lh}} \overline{U}}{\overline{Q}} U \equiv a_3 U. \quad (15)$$

Again, we note that the sensitivity change of the ML to the other heat fluxes ($\overline{Q}_{\text{lw}}$, $\overline{Q}_{\text{lh}}$, and $\overline{Q}_{\text{sh}}$) works as a negative feedback, and this process is discussed in section 4b.

The latent heat flux anomaly driven by the SST anomaly also generates the MLD anomaly. This MLD anomaly also changes the sensitivity to the climatological shortwave radiation; for example, given a warm SST anomaly, the latent heat release increases and the corresponding buoyancy loss causes a deepening of the ML, which dilutes the climatological shortwave warming. Hence, this SST–evaporation–mixed layer–SST effect acts to oppose the WEMS-related SST anomaly, which reflects the duality of the latent heat release similar to b_1 for a_1 . The second term on the RHS of Eq. (4), the third term on the RHS of Eq. (6), and the second term on the RHS of Eq. (11) combine to give

$$\frac{\overline{Q}_{\text{sw}}}{\rho_o c_p h} \frac{L}{R^* \overline{T}^2} \frac{\overline{Q}_{\text{lh}}}{\overline{Q}} T \equiv -b_2 T. \quad (16)$$

The negative climatological surface heat fluxes ($\overline{Q}_{\text{lw}}$, $\overline{Q}_{\text{lh}}$, and $\overline{Q}_{\text{sh}}$) will be considered in section 4b.

The above five processes (at most) yield the following thermodynamic equation for the SST change from Eq. (4)

$$\frac{\partial T}{\partial t} = aU - bT, \quad (17)$$

$$a \equiv a_1 + a_2 + a_3, \quad \text{and} \quad (18)$$

$$b \equiv b_1 + b_2. \quad (19)$$

The oceanic model extends for latitudes up to 2500 km from the equator in both hemispheres, but almost the same results are obtained even when the oceanic model covers the same domain as the atmospheric one (cf. black dashed line in Fig. 8 and lightest blue line in Fig. 9).

c. Basic states

Since our model is an anomaly model, the basic states need to be specified. Because we are interested in whether WIMS and WEMS feedbacks are related to the interhemispheric SST gradient, an equatorially symmetric mean state is adopted in order to avoid an asymmetry in the basic state leading to asymmetric eigenmodes. For this purpose, we specify a mean state that mimics the Atlantic in the late boreal winter to early spring [February–April (FMA)], when the intertropical convergence zone (ITCZ) and highest SSTs are located close to the equator and therefore, the Atlantic mean climate comes closest to being the equatorially symmetric (Xie and Carton 2004; Chiang et al. 2002):

$$\overline{Q}_{\text{sw}} = \begin{cases} 245 - 15 \cos(2\pi y/|y_0|), & |y/y_0| < 0.5 \\ 230 + 30 \cos[\pi(|y/y_0| - 0.5)], & |y/y_0| \geq 0.5, \end{cases}$$

$$\overline{Q} = 40 + 35 \cos(0.5\pi y/|y_0|),$$

$$\overline{T} = 23 + 4 \cos(0.5\pi y/|y_0|), \quad \text{and}$$

$$\overline{U} = \begin{cases} -4.5 + 0.5 \cos(2\pi y/|y_0|), & |y/y_0| < 0.5 \\ -3.5 - 1.5 \cos[\pi(|y/y_0| - 0.5)], & |y/y_0| \geq 0.5. \end{cases}$$

Meridional profiles of these fields are described in Figs. 2b–e together with the observations. The simulated mean MLD in this simplified model is shown in Fig. 2f. Since our model lacks the ML deepening process, it tends to underestimate the MLD, especially poleward of 20°N. As will be seen in the following sections, however, oceanic variations are mainly confined within 10°, where our model captures characteristics such as minimum MLD at the equator despite its simplicity, giving some confidence to use the simple ocean model. We note that qualitatively the same results are obtained even when we prescribe a more realistic mean MLD distribution, for example $\overline{h} = 35 - 15 \cos(\pi y/|y_0|)$ (blue dashed line in Fig. 2f).

As previous studies using simple models were able to provide fundamental features of the tropical coupled variability, such as stability (e.g., Philander et al. 1984; Yamagata 1985; Hirst 1986; Jin 1997; Xie et al. 1999; Vimont 2010), we believe that our linear model can offer a good insight into the hypothesized WIMS and WEMS feedback processes. The simple coupled model is discretized with a spacing of 100 km using a staggered grid with free-edge boundary conditions for the ocean

TABLE 1. Abbreviations and notations used in this study.

Abbreviation	Description	Coefficient	Feedback sign
WES	Wind–evaporation–SST	a_1	Positive
WIMS	Wind-induced turbulence–mixed layer–SST	a_2	Positive
WEMS	Wind–evaporation–mixed layer–SST	a_3	Positive
—	Latent heat release anomaly due to SST anomaly	b_1	Negative
—	SST–evaporation–mixed layer–SST	b_2	Negative

and no-mass-flux boundary conditions for the atmosphere. By specifying particular term(s), say a_1 for the WES, we can isolate the effect of each process, and we look for the leading eigenmode of each system in the following sections. Abbreviations and coefficients for each feedback are summarized in Table 1 and the basic parameters are listed in Table 2.

4. Leading eigenmodes

a. Positive feedback processes

We first investigate the positive feedback: the WES, WIMS, and WEMS processes. This is achieved by specifying a_1 , a_2 , or a_3 separately for the thermal forcing coefficient a with $b = 0$, and by analyzing the leading eigenmode of each coupled system.

Consistent with previous studies (Xie 1999; Chang et al. 2001; Vimont 2010; Martinez-Villalobos and Vimont 2017), the WES mode has the structure of a meridional temperature gradient, where larger loadings are confined to the deep tropics ($\sim 10^\circ$ from the equator; Fig. 3a), with low and high surface pressure anomalies in the warmer and cooler hemispheres, respectively. In association with the surface pressure anomalies, westerly (easterly) anomalies are seen over warm (cool) SST anomalies, where the former in turn reinforces the latter through changes in the latent heat flux. Because the thermal forcing coefficient a_1 is larger closer to the equator (Fig. 4), the trade wind anomalies closer to the equator excite SST anomalies more efficiently. Also, the surface wind anomaly maxima are located slightly equatorward of the SST anomaly maxima (Xie 1999). On the other hand, opposite-signed wind anomalies are seen on the poleward sides of the SST anomaly maxima, which damp the SST anomalies and eventually force the opposite-signed SST anomalies. As a result, the coupled disturbances propagate equatorward (e.g., Xie 1999; Vimont 2010).

TABLE 2. List of parameters.

ε^{-1}	2 day
C_a	45 m s^{-1}
R^*	$462 \text{ J kg}^{-1} \text{ K}^{-1}$
L	$2.5 \times 10^6 \text{ J kg}^{-1}$
C_E	1×10^{-3}
RH	0.75
q_0	2.6×10^{-2}
T_0	303 K
ρ_a	1.2 kg m^{-3}
ρ_o	$1 \times 10^3 \text{ kg m}^{-3}$
C_D	1.4×10^{-3}
c_p	$4 \times 10^3 \text{ J kg}^{-1} \text{ K}^{-1}$
m	1.0
α	$2.5 \times 10^{-4} \text{ K}^{-1}$
ε_o^{-1}	2 day

The growth rate of the WES mode is 0.15 month^{-1} (an e -folding time scale of about 6–7 months) and the period is 49.3 months, both of which are much longer than the length of a season (3 months). Since a boreal-spring-like mean state is specified, the results should not be interpreted over periods longer than a season. Therefore, we may conclude that the “pure” canonical WES mode is a neutral mode (e.g., Kimoto et al. 2001) in the boreal spring in this simple model framework.

In the WIMS and WEMS modes, anomalous surface winds flow from the cooler hemisphere to the warmer hemisphere, which increases the magnitude of the trade winds in the cooler hemisphere and reduces the magnitude of the trade winds in the warmer hemisphere similar to the WES mode (Figs. 3b,c). The reduced (increased) trade winds generate shallower (deeper) MLDs over the warm (cool) SST anomalies because of the weaker (stronger) wind stirring for the WIMS mode and reduced (enhanced) surface evaporation for the WEMS mode, which feed back to the warm (cool) SST anomalies because of the higher (lower) sensitivity of the ML to the climatological shortwave radiation. The WIMS and WEMS modes also share the equatorward-propagating and deep-tropic features seen in the WES mode. This may seem intuitive because the essence of this study is in a one-dimensional extension of the ocean at each location (i.e., allowing temporal variations of the MLD at each grid point). However, we note that the fact that the WIMS and WEMS modes share many similarities with the canonical WES mode is not obvious because this one-dimensional extension leads to changes in the spatial distribution of the thermal forcing coefficients (Fig. 4). Thus, oceanic responses differ even under the same atmospheric forcing.

Regarding the growth rate, it is found that the WIMS and WEMS modes are stronger processes than the

FIG. 3. (a) One cycle of the oscillatory component of the SST (shading), surface winds (vectors), and geopotential (contours) anomalies for the WES ($a = a_1$, $b = 0$). The growth rate (in month^{-1}) and the period (in month) are indicated at the top. Westerly anomaly is represented as the rightward vector. Vectors are drawn in the region between 10°S and 10°N for visual aids. (b) As in (a), but for the SST (shading), surface winds (vectors), and MLD (contours) anomalies for the WIMS ($a = a_2$, $b = 0$). (c) As in (b), but for the WEMS ($a = a_3$, $b = 0$). The color and vector scales are the same in every panel and are arbitrary. The contour interval is the same in (b) and (c) and is arbitrary. Solid contours denote positive values and dashed contours denote negative values.

FIG. 4. Meridional distribution of the thermal forcing coefficient a_1 (WES; black), a_2 (WIMS; red), and a_3 (WEMS; blue) in 10^{-7} K m^{-1} .

canonical WES feedback. Their growth rates are 0.82 and 0.48 month^{-1} , which correspond to an e -folding time scale of about 1.2 and 2.1 months, respectively. Therefore, the “pure” WIMS and WEMS modes may be interpreted as weakly unstable modes. In addition, their periods are relatively short, and so the oscillatory component could contribute to the amplification via transient growth in boreal spring.

We note that while the ocean-to-atmosphere coupling is common between the WES, WIMS, and WEMS models, the relative strength of the atmosphere-to-ocean coupling among them, and thus the relative amplitude of their growth rates, might be subject to a choice of the parameter. We confirm that, however, the inequality relation of the growth rates among the three feedback processes do not change for different values of the empirical parameter, the wind mixing coefficient m , covering the range from 0.5 to 1.5 (not shown).

b. Effective feedbacks: Introduction of damping terms and the negative climatological surface heat fluxes

In the previous subsection, we investigated the characteristics of the positive feedbacks. As discussed in section 2, the WES and WEMS feedback processes have corresponding damping terms, b_1 and b_2 , respectively. Moreover, the MLD change associated with the WIMS and WEMS modes modulates the contribution from the negative climatological surface heat fluxes in addition to the climatological shortwave radiation. The effect of the ML sensitivity change to the negative surface heat flux

can be easily introduced by replacing $\overline{Q_{sw}}$ with $\overline{Q}$ in a_2 , a_3 , and b_2 :

$$\frac{3m\overline{U}_*^2}{2\alpha g h^2} \sqrt{\frac{\rho_a C_D}{\rho_o}} U \equiv a'_2 U. \quad (14')$$

$$\frac{1}{\overline{U}} \frac{\overline{Q_{lh}}}{\rho_o c_p \overline{h}} U \equiv a'_3 U, \quad \text{and} \quad (15')$$

$$\frac{L}{R^* \overline{T}^2} \frac{\overline{Q_{lh}}}{\rho_o c_p \overline{h}} T \equiv -b'_2 T. \quad (16')$$

The inclusion of the negative climatological surface heat fluxes leads to smaller WIMS and WEMS terms ($a'_2 < a_2$ and $a'_3 < a_3$), because $\overline{Q} < \overline{Q_{sw}}$ and therefore weakens the positive feedbacks. On the other hand, it results in a smaller damping for WEMS ($b'_2 < b_2$), that is, it dilutes the damping effect. In this subsection, we analyze and compare the WES, WIMS, and WEMS modes with consideration of the effect of their respective damping terms and the negative climatological surface heat fluxes. In this case, the thermal forcing and damping coefficients of the WEMS mode become identical to those of the WES mode ($a_1 = a'_3$ and $b_1 = b'_2$). We note that this stems from the model assumptions: the entrainment cooling is assumed to be zero; and the penetrative shortwave radiation at the base of the ML is neglected to derive Eq. (8). These assumptions lead to the identical coefficients between the WES and WEMS, and thus, they are unlikely to be identical in reality.

In the presence of the damping terms and the negative climatological surface heat fluxes, the WES and WEMS feedback processes exist as least damped modes (Figs. 5a,c) consistent with the previous studies on the WES mode (e.g., Xie 1999). For the WES mode, while the introduction of the damping term leads to the smaller growth rate and slightly shorter period, the physical relationship between the atmosphere and ocean remains essentially the same. Thus, this mode is still accompanied by the cross-equatorial SST gradient interacting with zonal winds, which is confined to the deep tropics. The equatorward-propagating feature is observed as well. The zonal wind anomalies forced by the meridional SST gradient feed back onto the initial SST anomalies through the wind speed–related latent heat flux anomalies. For the WEMS mode with the damping term and the negative climatological surface heat fluxes, this wind speed–related latent heat release anomaly generates an MLD anomaly, which alters the sensitivity to the climatological net surface heat flux and feedback to the initial

FIG. 5. As in Fig. 3, but for (a) the WES mode with the damping term ($a = a_1$, $b = b_1$), (b) the WIMS mode with the negative climatological heat fluxes ($a = a'_2$, $b = 0$), and (c) the WEMS mode with the damping term and the negative climatological heat fluxes ($a = a'_3$, $b = b'_2$).

SST anomalies. Because of the diluted positive feedback ($a'_3 < a_3$; the efficiency of an MLD anomaly to generate a SST anomaly is reduced), T lags behind h more than in the “pure case” (Figs. 3c and 5c).

The evolution of the WIMS mode in the presence of the negative climatological heat fluxes is again confined

to the deep tropics with a cross-equatorial gradient (Fig. 5b). Zonal wind anomalies associated with the meridional SST gradient reduce (enhance) the wind speed over the warmer (cooler) hemisphere. The reduced (enhanced) trade winds then generate the shallower (deeper) ML because of the weaker wind stirring and feed back to the positive (negative) SST anomalies through the enhanced (diluted) warming by the climatological net surface heat flux. The equatorial maximum of the atmospheric thermal forcing coefficient a'_2 (not shown) and the equatorward phase shift of the zonal wind speed anomalies relative to the SST anomalies, together with the opposite-signed wind speed anomalies over the higher latitudes, lead to the equatorward propagation of the coupled disturbances. However, its period of about 29 months is much longer than the length of a season. Also, although the WIMS mode remains a weakly unstable mode, the growth rate is relatively small (an e -folding time scale is about 4 months) given that our analysis is based on the boreal spring conditions. Therefore, the WIMS mode with the negative climatological surface heat fluxes may be interpreted as a neutral mode. We note again that the inequality relation of the growth rates between the three feedbacks remains the same for different wind mixing coefficient m ranging from 0.5 to 1.5 (not shown).

5. Implications for the AMM

The eigenanalysis of the simple air–sea coupled systems revealed that the WIMS and WEMS feedback, which are accompanied by anomalous cross-equatorial SST gradients, do exist, and they are stronger than or as strong as the canonical WES feedback. Are they observed in reality? To address this question, the same analysis is performed again but with all of the components of the simple model retained ($a_1 = a_1 + a'_2 + a'_3$ and $b_1 = b_1 + b'_2$) as these processes are expected to occur simultaneously in reality. We note that although how the eigenvectors associated with the three feedback coupling interfere and how the transient growth is produced are interesting topics, more processes, such as low-level cloud feedback (e.g., Evan et al. 2013), should be included for a better simulation of the Atlantic meridional mode (AMM). Therefore, they are beyond the scope of this study and we still focus on the eigenmodes of the system. As expected, the leading mode has a cross-equatorial temperature gradient interacting with U and h as seen in the previous section (Fig. 6). While it is a weakly unstable mode, its growth rate is small and the period is long relative to a season, so that this mode may be considered a neutral mode. Figure 7a shows a snapshot of the oscillatory component of T at its peak

FIG. 6. As in Fig. 3b, but for the “mixed” mode ($a = a_1 + a'_2 + a'_3$, $b = b_1 + b'_2$).

and h 2 months before (when h peaks). An antiphase relationship between T and h can be seen. This feature bears some resemblance to those observed during the Atlantic meridional mode from the observations and the CGCM (Figs. 7b,c). In fact, recent studies based on in situ observations showed the importance of MLD anomalies in generating SST anomalies in the tropical North Atlantic during the AMM (Foltz et al. 2012; Rugg et al. 2016). Therefore, it is likely that the coupled feedback processes proposed in this study operate in addition to the canonical WES feedback, although the meridional width in the simple model is about 3 times smaller.

The ocean–atmosphere coupled feedbacks examined in the present study involve only one-dimensional vertical processes for the ocean. However, in reality, oceanic horizontal currents may be invoked when an atmospheric surface wind forcing is applied. Therefore, it is interesting to investigate the possible role of oceanic advection. Under the assumption of a zonally symmetric state, we can consider only the Ekman flow because the contributions from geostrophic currents disappear because of the absence of zonal temperature and pressure gradients. The contribution of the Ekman flow can be introduced by including the following two terms on the RHS of Eq. (17) (e.g., Xie 1999):

$$-\frac{f}{f^2 + \varepsilon_o^2} \frac{\rho_a C_D \overline{U}^2}{\rho_o \overline{h}} \frac{\partial T}{\partial y}, \quad \text{and} \quad (20)$$

$$-\frac{f}{f^2 + \varepsilon_o^2} \frac{2\rho_a C_D \overline{U}}{\rho_o \overline{h}} \frac{\partial \overline{T}}{\partial y} U, \quad (21)$$

where ε_o is the mechanical damping rate. The former term represents the mean advection of the anomalous

FIG. 7. (a) A snapshot of T at its peak ($t = 9$; black; left y-axis label) and h at $t = 7$ (when h peaks; blue; right y-axis label) from the “mixed” mode in Fig. 6. Units are arbitrary, but the relative amplitudes between T and h are conserved. (b) Zonally averaged anomalies of SST in April (in $^{\circ}\text{C}$; black; left y-axis label) and MLD in February (in m; blue; right y-axis label) over the Atlantic Ocean regressed onto the normalized March–May–mean (MAM–mean) AMM index. The AMM index is defined by the difference in the SST anomalies between the northern (5° – 15°N , 50° – 20°W) and southern (5° – 15°S , 20°W – 10°E) tropical Atlantic Ocean (Doi et al. 2010). (c) As in (b), but for MIROC6. The standard deviations of the MAM–mean AMM index used for the normalization in (b) and (c) are 0.60° and 0.37°C , respectively. Note that different vertical scales are used for the MLD from the observations and MIROC6.

SST gradient, whereas the latter is the anomalous advection of the mean SST gradient.

As shown by Xie (1999), the mean advection of the anomalous temperature gradient opposes the equatorward propagation of the SST anomalies and therefore, offers a favorable condition for the ocean–atmosphere coupling. As a result, the period becomes longer and the growth rate gets larger. In addition, the meridional structure of the anomalous ocean temperature becomes wider (Fig. 8). Also, the relative amplitudes of T and h are more similar to those during the observed and simulated AMM. However, the meridional width of a lobe is still small compared to the observations and the CGCM. Since the previous studies show that the ocean–atmosphere coupling window influences the variability of the system (Xie 1996; Okajima et al. 2003; Vimont 2010; Martinez-Villalobos and Vimont 2016), the same analysis but with wider coupling windows is conducted (Fig. 9). It is found that although it is not a monotonic increase, the meridional structure tends to be wider with a broader coupling window. The meridional width of a lobe increases up to about 17° when $y_1 = 4000$ km, but is still narrower than the observations and the CGCM, indicating that processes not included in our simple model, such as low-level clouds (Okumura et al. 2001; Tanimoto and Xie 2002; Evan et al. 2013; Myers et al. 2018) and ocean subsurface dynamics (Doi et al. 2010), also play a role in the AMM evolution.

6. Conclusions and discussion

We have constructed simple ocean–atmosphere coupled models for the tropical climate and analyzed their stability under equatorially symmetric conditions, which

mimic the tropical Atlantic during boreal spring (February–April). The key findings can be summarized as follows:

- In addition to the canonical wind–SST–evaporation (WES) feedback, two ocean–atmosphere coupled feedback processes, which involve temporal variations of oceanic MLDs, exist in the tropical Atlantic: Wind-induced turbulence–mixed layer–SST (WIMS) feedback and wind–evaporation–mixed layer–SST (WEMS) feedback.
- Wind-induced turbulence–mixed layer–SST (WIMS) feedback: The anomalous meridional SST gradient is accompanied by an opposite-signed meridional SLP

FIG. 8. As in Fig. 7a, but for the case with the Ekman flow is superimposed (dashed lines). Note that the oscillatory component vanishes when the Ekman flow is included (the growth rate is 0.66 month $^{-1}$).

FIG. 9. As in Fig. 7a, but for the cases with the Ekman flow under a various width of the coupling window. Note that only T profiles are shown and that the lightest blue line is the same as the black dashed line in Fig. 8 except that the larger ocean domain is used here.

gradient in the overlying atmosphere, which forces the atmospheric surface flow toward the warmer hemisphere; the Coriolis force veers the surface flow to the right (left) in the Northern (Southern) Hemisphere, and hence modulates the trade winds; the weaker (stronger) surface winds contribute to the shallower (deeper) ML through suppressed (enhanced) turbulent mixing, which results in a greater (smaller) sensitivity of the ML to the climatological shortwave radiation and amplifies the initial warm (cool) SST anomalies (Fig. 1b).

- Wind–evaporation–mixed layer–SST (WEMS) feedback: The meridional SST gradient anomaly is accompanied by an opposite-signed meridional SLP gradient in the overlying atmosphere, which forces the atmospheric surface flow toward the warmer hemisphere; the Coriolis force veers the surface flow to the right (left) in the Northern (Southern) Hemisphere, and hence modulates the trade winds and the latent heat flux; the larger (smaller) latent heat flux into the ocean contributes to the shallower (deeper) ML because of the buoyancy gain (loss) of the ocean, which results in a greater (smaller) sensitivity of the ML to the climatological shortwave radiation and amplifies the initial warm (cool) SST anomalies (Fig. 1c).
- Without the damping and negative climatological surface heat fluxes, the WIMS and WEMS are stronger than the WES feedback in terms of stability, but

are still weakly unstable (Fig. 3). In the presence of a damping term and the negative climatological surface heat fluxes, the growth rate of the WEMS feedback is identical to that of the WES feedback in this simple model framework and that of the WIMS feedback is the largest (Fig. 5). Since their growth rates are close to zero and the corresponding periods are longer than the length of a season (the boreal spring mean state is assumed), all of these feedback processes may be considered as neutral modes. Similar to the WES feedback, the coupled disturbances associated with the WIMS and WEMS propagate toward the equator.

- It is likely that the WIMS and WEMS feedback operate during the AMM because an antiphase relationship between the SST and MLD anomalies, a characteristic feature of the WIMS and WEMS feedback in the simple model, is seen during the observed AMM and that simulated in a coupled GCM (Fig. 7).

In the present study, the linear stability of a simple tropical air–sea coupled system has been investigated. For this purpose, nonlinear terms are omitted. One of the processes ignored is the change in the sensitivity of the ML to the anomalous heat flux. Since a shallower ML is more sensitive to the same amplitude of the latent heat flux anomaly than a deeper ML, the amplitude of positive SST anomalies should be larger than those of negative SST anomalies in the other hemisphere in the WEMS when this nonlinear process is included. Thus, asymmetry between the hemispheres would arise in the variability even under the equatorially symmetric mean state. Another nonlinear process, the entrainment of cooler water below the ML, is also not considered. When the ML deepens, the entrainment tends to cool the ML, which supports the cooling effect associated with the diluted warming by the climatological heat flux. We therefore speculate that omitting the entrainment cooling does not change our conclusions qualitatively, though a quantitative assessment is beyond the scope of this study.

Also neglected is the buoyancy flux due to precipitation, despite the fact that the model atmosphere is driven by large-scale condensation. The freshwater flux anomaly associated with the precipitation anomaly may offer another possible feedback process: anomalously warm SST causes a positive precipitation anomaly, which shoals the MLD because of freshening of the ML, and feeds back to the initial SST anomaly through enhancement of the warming by the climatological surface heat flux. Such an ocean–atmosphere coupled process is a possible topic for future studies.

Finally, we note that both the WIMS and WEMS feedback processes have a similar spatial SST pattern to

FIG. A1. (a) Climatological annual-mean SST (in $^{\circ}\text{C}$) along the equator averaged between 5°S and 5°N from the observations. (c) Longitude–time section of the observed climatological SST (in $^{\circ}\text{C}$) averaged between 5°S and 5°N . The contour interval is 0.5°C . (e) Time–latitude section of the observed climatological SST (in $^{\circ}\text{C}$; contour) and precipitation (in mm day^{-1} ; shading) averaged between 30° and 20°W . The contour interval is 0.5°C and 2.5 mm day^{-1} , respectively. The 27°C isotherms are drawn in a thick line. (b),(d),(f) As in (a),(c),(e), but for MIROC6.

FIG. A2. (a) Regressed anomalies of the observed SST (in $^\circ\text{C}$; shading) and surface wind (in m s^{-1} ; vector) onto the normalized AMM index. Vectors are drawn when the wind speed exceeds 0.2 m s^{-1} for visual aids. (c) As in (a), but for precipitation (in mm day^{-1}). (b),(d) As in (a),(c), but for MIROC6. (e) Monthly standard deviation of the normalized AMM index (black; COBE-SST2 and pink; MIROC6). The standard deviations used for the normalization are indicated in (e).

that of the WES feedback as discussed above. Moreover, they are also accompanied by reduced (strengthened) surface winds over warm (cool) SST anomalies. This indicates that analyses focusing only on surface wind speeds and heat fluxes and SSTs may hide the potential role of the WIMS and/or WEMS feedback. Therefore, it may be worth revisiting phenomena in which the WES feedback plays an important role (e.g., Chiang and Vimont 2004; Wu et al. 2008) with the MLD variations taken into account.

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FIG. A3. SST anomalies on 1 Apr (in $^{\circ}\text{C}$; contour) and MLD anomalies on 1 Feb (in m; shading) regressed onto the normalized MAM-average AMM index from (a) the observations and (b) MIROC6. Contour intervals are 0.2°C . Note that the different color scales are used between the observations and MIROC6.

APPENDIX

Tropical Atlantic Climate Simulated by MIROC6

In this appendix, the ability of MIROC6 to reproduce the mean tropical Atlantic climate and the AMM is briefly summarized. We begin by calculating the annual-mean SST along the equator because most state-of-the-art climate models have difficulty in simulating this (Richter et al. 2014, 2016). MIROC6 also fails to reproduce the observed annual-mean zonal SST gradient (Figs. A1a,b). This is because the simulated cold tongue that develops around the boreal fall is detached from the eastern boundary and shifted westward compared to the observations (Figs. A1c,d). It is, however, found that the seasonal warming of the SST during the boreal spring is well simulated by MIROC6. Regarding the meridional structure of the SST, the model realistically simulates the seasonal cycle (Figs. A1e,f). Although the simulated SST maximum is located northward of the observations, the model reproduces the seasonal movement of the maxima onto the equator around the boreal spring. In association with this, the model is also successful in reproducing the seasonal migration of the ITCZ toward the equator in this season, though the modeled ITCZ migrates far southward in comparison with the observations, which is a common bias in GCMs: even atmospheric GCMs forced with the observed SST show this bias (Biasutti et al. 2006; Richter et al. 2014). Finally, MIROC6 simulates reasonably well the meridional distribution of the MLD in spring, but it exaggerates the equatorial minima and the deepening at higher latitudes (Fig. 2f).

Regarding the AMM, MIROC6 simulates reasonably well spatial patterns including the anomalous SST meridional gradient, ITCZ displacement, and the associated surface wind anomalies toward the warmer

hemisphere (Figs. A2a–d). One noticeable discrepancy is that their amplitudes are modest compared to the observations. Also, the precipitation anomalies are shifted southward, likely because of the above-mentioned bias in the mean ITCZ position. While most CGCMs appear to have difficulty in reproducing the seasonal intensification of the interhemispheric SST gradient variability (Amaya et al. 2017), the seasonal phase locking in MIROC6 is comparable to the observations (Fig. A2e). The peak in the boreal spring, however, is not as sharp as that in the observations. As shown in section 5, the zonal averages of the SST anomalies are preceded by those of the opposite-signed MLD anomalies during the AMM, indicating the importance of the WIMS and WEMS feedback proposed in this study (Fig. 7). Therefore, it is interesting to look at their spatial patterns (Fig. A3). We can see that the SST anomalies around the peak of the AMM tend to be anticorrelated with the MLD anomalies two months before both in the observations and MIROC6, though the observed MLD pattern is noisy because its period is somewhat short and it is sparse. We also note that the amplitude of the simulated MLD anomaly is smaller. In summary, MIROC6 reasonably well simulates the tropical Atlantic climatology during the boreal spring, which gives us confidence using this model for the current study.

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