Investigating Extratropical Influence on the Equatorial Atlantic zonal bias with Regional Data Assimilation

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ABSTRACT

A reversal of zonal sea surface temperature (SST) gradient in the equatorial Atlantic is a common bias in climate models. Studies to investigate the origin of this bias mainly focused on the tropics itself. Applying the regional data assimilation method in the GFDL CM2.1 model, we investigate the combined and respective influences of the northern and southern extratropics on this bias. It is found that the reversed zonal SST gradient bias is caused to a considerable extent by extratropical atmosphere, especially by the northern extratropics. This extratropical impact on the equator is mainly through influencing the Hadley circulation. Therefore, the ITCZ position in boreal spring in this model most likely determines the dominant role of northern extratropics in the spring equatorial westerly bias and further the zonal SST gradient bias. Due to the cold bias in extratropical atmosphere, the northward shift of ITCZ coupled with the increased meridional SST gradient caused by assimilating the northern extratropics strengthens the cross-equatorial southeasterly wind, thus correcting spring equatorial westerly bias. The strengthened spring equatorial easterlies further steepen the thermocline slope and enhance the eastern upwelling, thus reproducing the summer cold tongue and finally improving the annual-mean zonal SST gradient bias.

1. Introduction

The equatorial Atlantic exhibits a clear zonal asymmetry with a warm pool in the west and a cold tongue in the east, accompanied by a Walker circulation with surface easterly winds along the equator, similar to the equatorial Pacific. Correct simulation of this equatorial zonal SST gradient is important since it is tightly related to the tropical convection and further the atmospheric circulation (Haarsma and Hazeleger 2007; Brandt et al. 2011). Especially, the equatorial Atlantic SST has a great impact on West African summer monsoon (Giannini et al. 2003; Okumura and Xie 2004; Chang et al. 2008; Steinig et al. 2018), the rainfall over the
American continent (Enfield et al. 2001; Wang et al. 2006; Crespo et al. 2019) and the Atlantic hurricane (Goldenberg et al. 2001; Webster et al. 2005; Wang and Lee 2007). However, nearly all the coupled general circulation models (CGCMs) exhibit the bias of a reversed annual-mean zonal SST gradient in the equatorial Atlantic, with SST increasing from west to east (Davey et al. 2002; Richter et al. 2014b). Moreover, this zonal SST gradient bias, has almost been unimproved during decades of model development, from the Coupled Model Intercomparison Project Phase 3 (CMIP3) to CMIP6 (Richter and Xie 2008; Richter et al. 2014b; Richter and Tokinaga 2020). Understanding the causes of this bias is of importance to improve the model simulation.

The origins of the warm SST bias in the eastern tropical Atlantic have been investigated from many aspects. In the ocean component, this warm SST bias is related to the deficient coastal upwelling (Large and Danabasoglu 2006; Grodsky et al. 2012; Kurian et al. 2021) or the vertical mixing (Hazeleger and Haarsma 2005; Exarchou et al. 2018; Deppenmeier et al. 2020). Studies targeting the atmospheric component (Hu et al. 2008; Hu et al. 2011; Toniazzo and Woolnough 2014; Exarchou et al. 2018) link this warm SST bias to the excessive shortwave radiation due to insufficient stratus clouds, though recently the effect of stratus cloud is being debated (Richter 2015; Hourdin et al. 2020). However, some seasonal hindcasts/forecasts (Huang et al. 2007; Voldoire et al. 2019) and sensitivity experiments (Wahl et al. 2011; Voldoire et al. 2019) imply that the warm SST bias in the southeastern tropical Atlantic and that in the eastern equatorial Atlantic is caused by different processes. The former is mainly caused by the surface heat flux, while the latter related to the reversed zonal SST gradient is mainly caused by the wind stress.

The observational annual-mean zonal SST gradient is associated with the boreal summer cold tongue, which is related to the enhanced vertical upwelling and entrainment in the eastern
boundary due to the strong cross-equatorial surface southerly wind with the onset of African monsoon (Philander and Pacanowski 1981; Mitchell and Wallace 1992; Okumura and Xie 2004). Meantime, the reversed annual-mean zonal SST gradient bias is mainly due to the lack of the summer cold tongue in models. According to this hypothesis, the meridional wind bias in the model may result in the disappearance of the summer cold tongue and further the bias of reversed annual-mean zonal SST gradient.

However, some recent studies link the zonal SST gradient bias to the equatorial westerly bias (DeWitt 2005; Richter and Xie 2008; Harlaß et al. 2015, 2018; Voldoire et al. 2019). DeWitt (2005) utilizes an ocean model forced by the observed zonal and meridional wind to find that the observed zonal wind instead of meridional wind produces the correct zonal SST gradient. Moreover, the westerly bias is still present in the atmospheric general circulation models (AGCMs) forced by the observed SST (Richter and Xie 2008; Richter et al. 2012; Richter et al. 2014a), suggesting that the zonal SST gradient bias in CGCMs is attributed to a large extent to the zonal wind bias from atmospheric component. Conceivably, this westerly bias in atmospheric models can further suppress the summer cold tongue in coupled models due to positive atmosphere-ocean feedbacks.

This westerly bias in the AGCMs may be associated with the deficient (excessive) precipitation over Amazon (Congo) (Chang et al. 2007; Richter and Xie 2008; Wahl et al. 2011; Patricola et al. 2012; Richter et al. 2012; Zermeño-Diaz and Zhang 2013). Chiang et al. (2001) points out that the elevated heating released by deep convection may have a greater impact than SST gradients on the equatorial Atlantic zonal winds. Several studies also link this zonal wind bias to the location of the intertropical convergence zone (ITCZ) (Breugem et al. 2008; Richter and Xie 2008; Tozuka et al. 2011; Richter et al. 2014a; Richter et al. 2014b; Richter and Tokinaga 2020). The excessive southward shift of the ITCZ in boreal spring with excessive
precipitation south of the equatorial Atlantic is a common bias even in AGCMs (Biasutti et al. 2006). Richter et al. (2014b) shows a high correlation between the ITCZ position and the strength of equatorial easterly wind among the CMIP5 models. In addition, excessive rainfall in the southeastern tropical Atlantic can cause a spurious fresh barrier layer there (Breugem et al. 2008), which may significantly contribute to a weaker cold tongue by suppressing the entrainment of cold water (Hazeleger and Haarsma 2005).

Most studies so far, however, focused on the tropical processes, with less attention paid to the extratropical influence. This is due partly to the difficulty to isolate the local and remote influences on the tropics in coupled models. Therefore, important questions on the role of extratropics remain wide open. To what extent can the equatorial Atlantic zonal bias be traced back to the extratropics? What are the relative contributions of the northern and southern extratropics? Here, we will investigate these questions using the Regional Data Assimilation (RDA) approach (Lu et al. 2017a; Lu et al. 2017b; Lu and Liu 2018) in the GFDL CM2.1 model. Our study suggests a significant extratropical impact on the equatorial Atlantic zonal bias, especially from the northern extratropics. The paper is organized as follows. Section 2 introduces the RDA method and model experiments. Section 3 shows the improvement of the annual-mean equatorial Atlantic zonal biases. In Section 4, the mechanism of improving the annual-mean zonal SST gradient bias is analyzed. The mechanism of improving seasonal biases is further analyzed in Section 5. Section 6 investigates the mechanism for the dominant role of the northern extratropics in the improvement. A summary is given in Section 7.

2. Methodology and model experiments

a. Model description
The CGCM analyzed in this study is the GFDL CM2.1 (Delworth et al. 2006), which has been used for climate change experiments in the Intergovernmental Panel on Climate Change Fourth Assessment Report (IPCC AR4). The resolution of the atmospheric (AM2.1) component is $2.5^\circ \times 2^\circ$ with 24 vertical levels in a hybrid coordinate. The ocean component is the fourth version of the Modular Ocean Model (MOM4) configured with 50 vertical levels up to 5000 m and $1^\circ \times 1^\circ$ horizontal B-grid resolution, telescoping to $1/3^\circ$ meridional spacing near the equator.

b. Regional Data Assimilation (RDA) method

A data assimilation system based on the ensemble adjustment Kalman filter (EAKF) (Anderson 2001, 2003; Zhang et al. 2007) is implemented in GFDL CM2.1. In the RDA method (Lu et al. 2017a; Lu et al. 2017b; Lu and Liu 2018), one can assimilate the selected state variables in the desired domain and then study the ensemble-mean response in the unassimilated region. In this study, we assimilate extratropical atmospheric temperature (T) and winds (U, V) with 4-times daily reanalysis data from the NCEP/NCAR Reanalysis 1 (Kalnay et al. 1996). This set of reanalysis data is regarded as the observation (OBS) in our study. The assimilation is carried out every 6 hours with the effect radius of 1000 km. The error scales for observational atmospheric temperature (T) and winds (U, V) are assumed to be 0.5 K and 1 m/s, respectively (Zhang et al. 2007). The RDA experiments are run for 40 years from 1969 to 2008.

c. Observational data

The following observational datasets are used for comparison: Hadley Centre Sea Ice and Sea Surface Temperature (HadISST) (Rayner et al. 2003) for SST; NCEP/NCAR Reanalysis 1 (Kalnay et al. 1996) for surface air temperature (SAT), surface winds, surface level pressure (SLP) and surface heat flux; Monthly Climate Prediction Center (CPC) Merged Analysis data.
(CMAP) (Xie and Arkin 1996) for precipitation. Ocean temperature from version 4 of the Met Office Hadley Centre EN series (EN4.2.1) (Good et al. 2013) is used for calculating thermocline depth. Since precipitation data is only available since 1979, the climatology in our study is analyzed based on the 30 years from 1979 to 2008. While the evolution is analyzed from the first month of 1969.

**d. Experiments design**

To investigate the combined and respective influences of the northern and southern extratropical atmosphere on the tropical Atlantic climatology, a control experiment (CTRL) and two groups of RDA experiments are performed. CTRL is a historical run from 1969 to 2008 without data assimilation. The first group of RDA experiments includes ADA20, ADA20N and ADA20S, with atmospheric data assimilation (ADA) activated poleward of 20° in both hemispheres, only north of 20°N and only south of 20°S, respectively. To confirm the robustness of extratropical influence and also to assess the relative contributions of different extratropical belts, the second group is set up including ADA30, ADA30N and ADA30S, which is the same as the first group except that the assimilation boundary is 30°. Each experiment has 12 ensemble members, and the results in our study are the ensemble mean of the 12 members. All experiments use the same ensemble of initial conditions, which are the restart files after a two-year ensemble historical simulation (1967-1968). Compared with CTRL, the improvement of equatorial Atlantic zonal bias in RDA experiments, reveals the contribution of the specific extratropical atmospheric bias to the present equatorial Atlantic bias in the GFDL CM2.1.

The validation of the RDA scheme is exhibited in the model bias in zonal- (60°W–10°E) and annual-mean temperature in the Atlantic in Fig. 1. The overall effect of the assimilation on global zonal-mean temperature bias is similar (not shown). A cold bias up to 4°C appears...
widely in the extratropical air temperature in CTRL (Fig. 1a). After the assimilation, air temperature bias in the assimilated extratropical region is reduced dramatically to ~0.5°C (except over the Antarctic continent) (Fig. 1b–g), consistent with the error used for observation. It should be noted that, there is a basin wide cold bias for ~1°C in both the zonal-mean SST and SAT in the tropical Atlantic in CTRL (Fig. 1a), which is also mentioned by Li and Xie (2012). This basin wide cold bias disappears and even turn to a modest warm bias when either hemispheric extratropics is assimilated (Fig. 1b–g). This change of zonal-mean temperature in the tropical Atlantic is likely caused by zonal-mean heat transport as shown previously in ideal experiments (Liu and Yang 2003) and is not the focus of this paper.

3. The improvement of the annual-mean equatorial Atlantic zonal biases

In the annual mean, the observational equatorial Atlantic shows a great zonal asymmetry with the west warmer than the east by ~1.5°C in SST (Table 1 and black line in Fig. 2a) and by over 1°C in SAT (black line in Fig. 2b), accompanied by a strong surface easterly (black line in Fig. 2d). However, CTRL shows a severe bias of a reversed zonal SST gradient as most CGCMs (Richter et al. 2014b), with the west colder than the east by ~0.5°C (Table 1 and red line in Fig. 2a). The reversed zonal gradient is also present in the oceanic section (37°W~10°W) of SAT (red line in Fig. 2b), albeit with a smaller amplitude. Coupled with the reversed SST/SAT gradient, the equatorial surface easterly over the oceanic section in CTRL is weaker than the OBS by ~0.6 m/s (red line in Fig. 2d).

When the extratropical atmosphere poleward of 20° is assimilated in ADA20, the equatorial biases are reduced substantially. The reversed zonal SST gradient is corrected, with the west (50°W) warmer than the east (10°W) by ~0.6°C (Table 1 and blue solid line in Fig. 2a), indicating the improvement degree reaches 55% (Table 1). The equatorial easterly over the oceanic section is nearly enhanced to the observational level (blue solid line in Fig. 2d).
Besides, the zonal SAT gradient (blue solid line in Fig. 2b) over the central oceanic section is improved to be almost the same as the OBS, although there is not much change in the SAT gradient of the western and eastern sections due to the poor simulation of continental SAT.

A comparison between ADA30 and ADA20 further shows that these extratropical impacts originate largely from poleward of 30°. The improvement in the zonal gradient of SST/SAT is also clear in ADA30 (blue dotted line in Fig. 2a,b) with the improvement degree in zonal SST gradient reaching 38% (Table 1), which is also accompanied by a significant enhancement in the equatorial easterly (blue dotted line in Fig. 2d). Of course, as expected, the biases are reduced slightly more in ADA20 than in ADA30 (Fig. 2a,b,d, blue solid vs. blue dotted), suggesting the additional role of extratropics between 20° and 30° in the equatorial Atlantic zonal biases.

Furthermore, a comparison between ADA20N (ADA30N) and ADA20S (ADA30S) suggests that the extratropical impact on the equatorial Atlantic is dominated by northern extratropical impact. It can be found that the zonal gradient of SST/SAT and the equatorial easterly wind (Fig. 2a,b,d) are almost the same between ADA20N (ADA30N) (orange line) and ADA20 (ADA30) (blue line), and between ADA20S/ADA30S (green line) and CTRL (red line), implying the little role of southern extratropics in these biases.

It is interesting to compare the zonal SST gradient in the equatorial Atlantic with that in the equatorial Pacific in the model. In the equatorial Pacific (Fig. 3a), the zonal SST gradient in CTRL and RDA experiments (except for ADA30) is nearly the same as that in OBS (~4.5°C), although with a stronger surface easterly wind than OBS (Fig. 3b). One interpretation for this robust consistent zonal SST gradient in the equatorial Pacific is that it has reached a quasi-saturation state that is about one-quarter of the equilibrium SST difference between the tropics and mid-latitude, as suggested by Liu and Huang (1997). If this is true, the zonal SST gradient
in the equatorial Atlantic is far below the saturation state and can be influenced substantially by extratropics. Another possibility is that the narrow tropical Atlantic is affected substantially by the land processes, while the tropical Pacific is overwhelmingly determined by the ocean-atmosphere process.

4. Mechanism of improving the annual-mean zonal SST gradient bias

The change of zonal SST gradient should be forced by either the surface heat flux or the ocean dynamics. However, the analysis shows that the surface heat flux cannot be the cause for the observed zonal SST gradient, the model bias of reversed zonal SST gradient and the improvement of zonal SST gradient in RDA experiments. In OBS, surface heat flux increases from west to east (black line in Fig. 2c), which alone would force a warmer SST in the east, opposite to the observed zonal SST gradient (black line in Fig. 2a). This distribution of surface heat flux is mainly due to the less downward shortwave radiation and more upward latent heat flux in the west (not shown). In CTRL, the surface heat flux decreases from 30°W to the east (red line in Fig. 2c), which would force a zonal SST gradient opposite to CTRL (red line in Fig. 2a). Furthermore, in ADA20 and ADA30 (blue line in Fig. 2c), the surface heat flux increases from 30°W to the east, which cannot interpret the correction of reversed zonal SST gradient (blue line in Fig. 2a).

The failure of surface heat flux in forcing the zonal SST gradient implies the importance of the ocean dynamics, which is consistent with the previous work (Richter and Xie 2008). Here in the GFDL CM2.1, the reversed zonal SST gradient in CTRL and ADA20S/ADA30S is also accompanied by a weaker surface easterly and a flatter thermocline than OBS (red and green line in Fig. 2d,e). Furthermore, the improved zonal SST gradient in ADA20/ADA30 and ADA20N/ADA30N is also accompanied by the enhanced easterly and the steepened thermocline relative to CTRL (blue and orange line in Fig. 2 d,e). This suggests the vital role
of surface zonal wind in the zonal SST gradient in this model. Considering the importance of meridional wind in the observational zonal SST gradient, we also check this component. A slightly stronger cross-equatorial southerly wind appears in the eastern boundary both in ADA20/ADA30 and ADA20N/ADA30N (blue and orange line in Fig. 2g), which may also contribute to a relative colder SST in the east by enhancing upwelling. However, the annual-mean upper oceanic upwelling remains almost unchanged in ADA20/ADA30 compared with CTRL (blue vs. red in Fig. 2f). A possible suggestion is that the upwelling needs to be examined from the seasonal perspective. This seasonal perspective is further motivated by the fact that the observed annual-mean zonal SST gradient is related to the summer cold tongue, which most CGCMs fail to capture.

5. Mechanism of improving the seasonal biases

The eastern equatorial Atlantic in OBS (Fig. 4a) shows a clear annual cycle with the cooling of SST from boreal spring to summer, namely the appearance of summer cold tongue, accompanied by the enhanced southeasterly wind with the onset of West African monsoon. In CTRL (Fig. 4b), the reversed zonal SST gradient bias in annual mean is attributed to both a warm SST bias in the east in boreal summer (June-July-August, JJA) and a cold SST bias in the west in summer and fall (June-November). This warm (cold) SST bias in the east (west) in summer, or namely the disappearance of summer cold tongue, is preceded by a westerly bias in boreal spring (April and May, AM) and further a deeper (shallower) thermocline bias in the east (west) in May-June-July (MJJ). It is this spring westerly bias that contributes almost entirely to the annual-mean westerly bias (Fig. 2d). The cold SST bias (Fig. 4b) in the west is also accompanied by a stronger cross-equatorial southerly bias in the west. While in OBS (Fig. 4a), the cross-equatorial southerly in the west is weaker than that in the east due to the onset of West African monsoon.
When the extratropical atmosphere is assimilated in ADA20, all seasonal biases are reduced significantly (Fig. 4c). Hence the improvement of annual-mean zonal SST gradient bias in ADA20 could be caused by two mechanisms. First, the strengthened easterly in boreal spring forces a steeper thermocline in MJJ, and in turn, a relative colder (warmer) SST in the east (west), thus reproducing the summer cold tongue and correcting the annual-mean zonal SST gradient. Second, the weakened southerly wind in the west in summer and fall produces a warmer SST in the west by inhibiting latent heat loss, therefore improving annual-mean zonal SST gradient.

Further analysis indicates that the first mechanism is the leading mechanism, which can be seen in a comparison of the improved seasonal biases between ADA20N (Fig. 4e) and ADA20S (Fig. 4f). Here the extratropical impact on the equatorial Atlantic seasonal biases is approximately a linear superposition of each extratropical impact (Fig. 4d vs. Fig. 4c) without regard to a difference in equatorial-mean SST. Furthermore, the improved biases in spring surface zonal wind, MJJ thermocline slope and summer cold tongue are caused by the northern extratropics (Fig. 4e), while the improved meridional wind bias in the west in summer and fall is caused by the southern extratropics (Fig. 4f). As discussed in Fig. 2, the annual-mean zonal SST gradient bias is corrected in ADA20N, not in ADA20S. Therefore, it is the improvement of the spring surface westerly bias and further the summer cold tongue bias caused by the northern extratropics that play a vital role in correcting the reversed annual-mean zonal SST gradient bias. The improvement in ADA30/N/S (Fig. A1) are very similar with that in ADA20/N/S, despite with a slight smaller magnitude.

In addition to steepening the thermocline, the strengthened spring equatorial easterly also enhances the eastern upper oceanic upwelling in MJJ, contributing to the reproduce of the summer cold tongue. In Fig. 5, a stronger upwelling occurs in the eastern upper ocean in
ADA20 and ADA20N (Fig. 5b,c) due to a stronger equatorial easterly than CTRL (Fig. 4c,e). In contrast, a weaker upwelling occurs in ADA20S (Fig. 5d) caused by a weaker southeasterly wind (Fig. 4f).

The relationships among the variables are quantified. The largest biases in CTRL (red line) as well as the largest improvement in ADA20/ADA30 (blue line) and ADA20N/ADA30N (orange line) appear in April-May-June (AMJ) surface zonal wind (Fig. 6a), MJJ thermocline slope (Fig. 6b) and JJA zonal SST gradient (Fig. 6c), consistent with Fig. 4. Furthermore, the strength of AMJ surface zonal wind, the MJJ thermocline slope, also, the intensity of MJJ eastern upwelling, and finally the JJA zonal SST gradient exhibit robust linear relationships (Fig. 6d–h), which is reminiscent of the relations in the CMIP5 ensemble (Prodhomme et al. 2019). As such, an ~1 m/s increase in surface easterly forces an ~1°C increase in zonal SST gradient in this model (Fig. 6d). It should be noted that, the JJA zonal SST gradient in ADA20/ADA20N is still somewhat less than that in OBS, in spite of the almost same strength of surface easterly wind (Fig. 6d and Fig. A2), indicating a deficiency in the ocean component, perhaps related to a weaker vertical mixing (Hazeleger and Haarsma 2005; Deppenmeier et al. 2020) or a weaker upwelling in the east than OBS.

As suggested by previous studies, although the meridional wind associated with monsoon development may control the appearance of summer cold tongue in the observation, the anomalous equatorial zonal wind in boreal spring can also lead to the variability of summer cold tongue (Keenlyside and Latif 2007; Lübbecke et al. 2018; Planton et al. 2018; Martín-Rey and Lazar 2019). This occurs because the zonal wind anomaly generates equatorial Kelvin waves, which can propagate eastward and alter the thermocline slope in about two months (Servain et al. 1982; Zebiak 1993; Doi et al. 2007; Polo et al. 2008; Hormann and Brandt 2009; Richter et al. 2014b). The importance of the spring wind bias to the summer zonal SST gradient
bias here is also consistent with Richter et al. (2012), in which the JJA eastern warm SST bias is reduced by 33% when forced with observed wind stress in MAM, but is only reduced by 3% when forced with the observed wind in JJA.

6. The mechanism for the leading role of the northern extratropics in the improvement

Discussions above indicate that it is the northern extratropics that predominantly reduces the spring zonal wind bias, thus reproducing the summer cold tongue and finally improving the annual-mean zonal SST gradient. Therefore, investigating why the northern extratropics can correct the spring westerly bias is the key to understand the leading role of the northern extratropics in the zonal SST gradient bias in this model.

We firstly analyze the model bias and the improvement of RDA experiments in the tropical Atlantic in boreal spring. In March-April-May (MAM), the biases in CTRL (Fig. 7b,e) are very similar with those in CMIP5 multi-model mean. In the tropical Atlantic off the equator, the cold SST/high SLP (warm SST/low SLP) bias in the west (east) is accompanied by the strengthened (weakened) trade winds, consistent with the Wind-Evaporation-SST (WES) feedback (Xie and Philander 1994). In the equatorial Atlantic, besides the surface westerly bias, there exists excessive (insufficient) precipitation bias south (north) of the equator (contours in Fig. 7b) as in most climate models (Richter and Xie 2008; Richter et al. 2014b; Richter and Tokinaga 2020), namely the bias of excessive southward shift of the ITCZ in spring. When extratropics is assimilated in ADA20 (Fig. 7c,f), all biases in MAM are improved substantially without regard to a warmer SST in the whole tropics. This extratropical impact on the tropical Atlantic is approximately the linear superposition (Fig. 7i,l) of each extratropical influence. The significant changes in ADA20N (Fig. 7g) expands from northern extratropics to the south

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of equator with the strengthened cross-equatorial southeasterly wind, while those in ADA20S (Fig. 7h) are mainly confined within the southern tropics. Therefore, the equatorial improvement is dominated by the northern extratropics. In ADA20N, it is likely the northward shift of ITCZ (contours in Fig. 7g) (Gill 1980) coupled with the increased cross-equatorial meridional SST gradient (shading in Fig. 7g) (Lindzen and Nigam 1987) that drives the cross-equatorial northwestward pressure gradient force (shading in Fig. 7j), thus strengthening the equatorial southeasterly wind and correcting the spring westerly bias. Our speculation is consistent with previous studies linking the spring westerly bias to either the excessive southward shift of ITCZ (Richter et al. 2014a; Richter et al. 2014b) or the meridional SST gradient bias (Wang et al. 2014; Song et al. 2015).

To further investigate the causes of the northward shift of ITCZ along with the increased meridional SST gradient in ADA20N, we examine the evolution of the changes (relative to CTRL) in Hadley circulation (HC) and SST in the tropical Atlantic.

The northward shift of ITCZ is likely caused by the MSE advection. From the evolution of the changes in HC over the tropical Atlantic in ADA20N (Fig. 8), we can find that ADA20N causes a weakened northern HC along with the northward shift of ITCZ, thus the significant changes are nearly confined within the northern HC of CTRL. The changes can be understood naturally from perspective of energetics: Due to the cold bias in the extratropical atmosphere in this model (Fig. 1a), the assimilation in ADA20N increases atmospheric moisture static energy (MSE) in the northern extratropics (shading in Fig. 8). This decreased energy contrast between tropics and northern extratropics calls for a weakened northern HC due to the leading role of HC in the atmospheric energy transport (AET) from tropics to mid-latitude (Held 2001). Furthermore, this injected MSE in the NH leads to an interhemispheric energy imbalance and calls for a compensating cross-equatorial AET from NH to SH. Since the direction of AET is
the same as the upper branch of HC and the strength of cross-equatorial AET is linear with ITCZ latitude (Chiang and Bitz 2005; Broccoli et al. 2006; Kang et al. 2008; Kang et al. 2009; Frierson and Hwang 2012; Donohoe et al. 2013; Frierson et al. 2013; Bischoff and Schneider 2014; Marshall et al. 2014; Schneider et al. 2014; Adam et al. 2016), this desired cross-equatorial AET from NH to SH calls for a northward shift of ITCZ. However, energetics framework is a fundamental diagnostic tool and hence does not offer physical insight into the extratropical impacts. Here we offer a speculation to the physical process: On the one hand, the descending branch of HC is located in the assimilated region. The strength of descending movement can be modified directly towards observation, thus adjusting the whole northern HC (vector in Fig. 8). On the other hand, the advection of MSE causes the northward shift of ITCZ. The injected MSE poleward of 20°N (shading in Fig. 8) can be advected to the tropics by the equatorward lower branch of CTRL’s HC. Meanwhile, the poleward lower branch of anomalous HC (ADA20N minus CTRL; vectors in Fig. 8) can reduce the advection of climatologic cold dry air from extratropics. These two ways lead to the increase of MSE in the northern HC, causing the northward shift of ITCZ (e.g., Fig. 8c,f).

By checking the evolution of the changes in surface variables over the tropic Atlantic, firstly, we find that the dominant role of the northern extratropics in the spring westerly bias is determined by the ITCZ location in spring in this model. Consistent with the changes of HC in Fig. 8, the changes of surface variables in ADA20N (ADA20S) in Fig. 9 are confined to the north (south) of the CTRL’s ITCZ, with the anomalous meridional precipitation dipole crossing the CTRL’s ITCZ. Therefore, with the seasonal shift of ITCZ in CTRL, the influence scope of each extratropics is different. In boreal spring (MAM), due to the bias of excessive southward shift of ITCZ, the equatorial Atlantic is predominantly influenced by the northern extratropics. Secondly, the increased meridional SST gradient is mainly driven by the wind-induced latent
heat flux changes, most likely caused by the changes of HC. From boreal autumn to next spring, the warm SST anomaly in ADA20N (minus CTRL) expands from the northern tropics to the equator (Fig. 9a), following the weakened northeasterly winds and the downward latent heat anomaly. This weakened easterly would drive a weakened poleward Ekman transport (e.g. contour in Fig. 8a,b) to bring less heat and then to produce a cold SST anomaly, thus the ocean dynamics cannot explain the warming of tropical SST. Instead, the changes of latent heat flux should be the main cause. We further assess the effect of changed surface wind to the changed latent heat flux and then to the warm SST, by replacing the surface wind of ADA20N with that of CTRL and by replacing all variables except for surface wind of ADA20N with those of CTRL. The results show that this weakened trade winds related to the weakened northern HC, no doubt dominates the increase of anomalous downward latent heat flux (Fig. 10a) and further causes the warming of tropical SST from autumn to next spring (Fig. 10b), thus increasing the cross-equatorial meridional SST gradient in boreal spring.

Discussions above indicate that the northern extratropics influences the spring equatorial westerly bias most likely by changing the HC. Here the WES feedback is also a potential contributing mechanism for northern extratropical impact on the equator (Liu and Xie 1994; Chiang and Vimont 2004; Chiang and Bitz 2005; Vimont et al. 2009; Wang 2010; Zhang et al. 2014; Amaya et al. 2019) although a detailed analysis requires sensitivity experiments. Under the balance of the pressure gradient force, Coriolis force and friction, the improved warm SST (Fig. 9 a) in the line of 20°N will produce an anomalous southwesterly wind at its south, weakening the northeasterly trade winds and decreasing the latent heat loss here, thus producing a new warm SST anomaly at its south, finally leading to the equatorward propagation of warm SST and weakened trade winds (Liu and Xie 1994). Therefore, the improved warm SST coupled with the weakened trade winds (Fig. 9 a) propagate equatorward.
in ADA20N can also be supported by WES feedback, thus increasing cross-equatorial meridional SST and precipitation gradient, further correcting the spring westerly bias.

7. Summary

The bias of the reversed zonal SST gradient in the equatorial Atlantic has been a puzzling problem for two decades among climate models. Most previous work has focused on the effect of local processes in the tropics on this bias. Our RDA experiments in the GFDL CM2.1 provide a new perspective: this zonal SST gradient bias can be caused substantially by extratropical atmosphere, especially by northern extratropics. The major mechanism for the extratropical influence on this equatorial Atlantic zonal bias is summarized in Fig. 11.

The reversed annual-mean zonal SST gradient bias in this model is mainly associated with the disappearance of the summer cold tongue, which is primarily caused by the equatorial surface westerly bias in boreal spring. In our RDA experiments, we identify the northern extratropics as the major extratropical forcing that corrects the spring equatorial westerly bias, thus improving the annual-mean zonal SST gradient bias. The dominant role of northern extratropics in zonal SST gradient bias is most likely determined by the ITCZ position in spring in this model. Since the influence of the northern (southern) extratropics is nearly confined within the northern (southern) HC of CTRL, in boreal spring, the equatorial Atlantic is predominantly controlled by the northern extratropics due to the bias of excessive southward shift of ITCZ. Therefore, the northward shift of ITCZ coupled with the increased meridional SST gradient caused by the northern extratropics drives the strengthened equatorial easterly, thus correcting the spring westerly bias. This enhanced spring equatorial easterly steepens the equatorial thermocline and enhances the eastern upwelling, thus reproducing the summer cold tongue and finally correcting the reversed annual-mean zonal SST gradient bias.

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We suggest that the mechanism for correcting the spring westerly bias is the northward shift of ITCZ, which may be related to the cold bias of extratropical atmosphere. Many precious studies have linked the shift of ITCZ to the extratropical forcing from energetics perspective. However, the physical process through which extratropical atmosphere affects the tropics still remain unclear. In our study, besides giving an explanation from energetics framework, we also attempt to interpret this remote forcing in terms of advection of MSE and/or the WES feedback. Much more rigorous analysis, however, remains to be studied in the future to fully understand this remote impact. It is also interesting to speculate if the conclusion derived from our model can also apply to some other climate models. We analyzed the climatologic air temperature bias of 11 CMIP5 models that are widely used currently, and found that the severe cold bias exists in the extratropical atmosphere of 10 models. We tend to believe that our conclusion may also apply, to some extent, to these models, even though at present we cannot explain the general cold bias in the extratropics in climate models.

It should also be noted that, however, the tropics has a great impact on the extratropical climate and variabilities. Therefore, the assimilated extratropical atmosphere in our RDA experiments which actually contains some observed tropical information that is not necessarily coherent with the model simulation. This fully interactive nature of tropical-extratropical climate makes it difficult to cleanly attribute the origin of the tropical bias to remote and local processes. Further studies are needed to better understand this issue.

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Data Availability Statement.

NCEP Reanalysis data and CMAP precipitation data are provided by NOAA/OAR/ESRL PSD from their website http://www.esrl.noaa.gov/psd/. HadISST sea surface temperature and EN4 (EN4.2.1) ocean subsurface temperature are provided by Met Office Hadley Centre from their website https://www.metoffice.gov.uk/hadobs/index.html. Reasonable request for the code of RDA and data of the experiments can be made to the corresponding authors.

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### TABLE 1. The zonal SST gradient in the equatorial Atlantic (SST difference between 50°W and 10°W; Units: °C) in the observation (OBS), CTRL and the RDA experiments. The middle column shows the biases of zonal SST gradient in CTRL and the RDA experiments (minus OBS; Units: °C). The last column shows the improvements (minus CTRL; Units: °C) of the RDA experiments. The percentages in parentheses quantify the degree of improvement measured by the ratio of improvements to the model bias of CTRL.

<table>
<thead>
<tr>
<th></th>
<th>Zonal SST gradient</th>
<th>Bias</th>
<th>Improvement</th>
</tr>
</thead>
<tbody>
<tr>
<td>OBS</td>
<td>1.46</td>
<td></td>
<td></td>
</tr>
<tr>
<td>CTRL</td>
<td>-0.51</td>
<td>-1.97</td>
<td></td>
</tr>
<tr>
<td>ADA20</td>
<td>0.57</td>
<td>-0.89</td>
<td>1.07(55%)</td>
</tr>
<tr>
<td>ADA20N</td>
<td>0.46</td>
<td>-1</td>
<td>0.96(49%)</td>
</tr>
<tr>
<td>ADA20S</td>
<td>-0.63</td>
<td>-2.09</td>
<td>-0.12(-6%)</td>
</tr>
<tr>
<td>ADA30</td>
<td>0.23</td>
<td>-1.23</td>
<td>0.74(38%)</td>
</tr>
<tr>
<td>ADA30N</td>
<td>0.16</td>
<td>-1.3</td>
<td>0.67(34%)</td>
</tr>
<tr>
<td>ADA30S</td>
<td>-0.35</td>
<td>-1.81</td>
<td>0.16(8%)</td>
</tr>
</tbody>
</table>
FIG. 1. The climatologic bias of atmospheric and upper oceanic temperature in the Atlantic (60°W–10°E) (a) in CTRL and the RDA performance (minus observation) of (b) ADA20, (c) ADA20N, (d) ADA20S, (e) ADA30, (f) ADA30N and (g) ADA30S. The thin dashed lines represent atmospheric isentropic surfaces. The thick dashed lines and dotted lines represent the assimilation boundaries of 20°N/S and 30°N/S.
FIG. 2. Annual-mean (a) SST, (b) SAT, (c) surface heat flux (positive value means downward), (d) surface zonal wind, (e) thermocline depth (depth of 20°C isotherm), (f) upper ocean vertical velocity (0–100 m average) and (g) surface meridional wind along the equatorial Atlantic (5°S–5°N) in OBS (black), CTRL (red), ADA20 (blue solid), ADA20N (orange solid), ADA20S (green solid), ADA30 (blue dotted), ADA30N (orange dotted) and ADA30S (green dotted). To show more details, the biases of zonal wind and thermocline depth in experiments instead of their original value are show in (d) and (e), respectively. The colored lines corresponding to the left axis in (d)(e) mean the biases of experiments, and the black line corresponding to the right axis in (d)(e) means the observational value. The corresponding colors and line styles will be applied to the full text. The shadings in line graphs show the spread of 12 ensemble members of each experiment.
FIG. 3. Annual-mean (a) SST, (b) surface zonal wind along the equatorial Pacific (5°S–5°N) in OBS (black), CTRL (red), ADA20 (blue solid), ADA20N (orange solid), ADA20S (green solid), ADA30 (blue dotted), ADA30N (orange dotted) and ADA30S (green dotted). The shadings in line graphs show the spread of 12 ensemble members of each experiment.
FIG. 4. Climatologic seasonal cycle of surface wind (vectors), SST (shading) and thermocline depth (purple contours, solid and dashed lines represent positive and negative values, respectively, contour interval is 2 m) (a) in OBS, (b) the biases (minus OBS) in the climatologic seasonal cycle of CTRL, and the changes (minus CTRL) in the climatologic seasonal cycle of (c) ADA20, (e) ADA20N, (f) ADA20S compared with CTRL along the equatorial Atlantic (5°S–5°N). (d) is the linear superposition of (e) ADA20N-CTRL and (f) ADA20S-CTRL. Only the significant changes based on T test at a 95% significance level is drawn.

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FIG. 5. Upper oceanic vertical velocity along the equatorial Atlantic (5°S–5°N) in May-June-July (MJJ) in (a) CTRL, (b) ADA20, (c) ADA20N and (d) ADA20S. The solid and dashed lines represent the equatorial mixed layer and thermocline, respectively.
FIG. 6. The relationships among the surface zonal wind (averaged between 5°S–5°N and 50°W–0°), thermocline slope (difference of thermocline depth between 50°W and 0°W, averaged 5°S–5°N), eastern upper oceanic upwelling (averaged between 20°W–0° and 5°S–5°N, 0–50 m average) and zonal SST gradient (SST difference between 50°W and 10°W, averaged 5°S–5°N). (a)(b)(c) are the seasonal cycles of the (a) surface zonal wind, (b) thermocline slope and (c) zonal SST gradient. (d)–(h) are the relationships between (d) AMJ surface zonal wind and JJA zonal SST gradient, (e) AMJ surface zonal wind and MJJ thermocline slope, (f) MJJ thermocline slope and JJA zonal SST gradient, (g) AMJ surface zonal wind and MJJ eastern upper oceanic upwelling and (h) MJJ eastern upper oceanic upwelling and JJA zonal SST gradient. In (d)–(h), the correlation coefficient is marked with ** if it is significant at a 99% significance level, and blue lines show the regression.
FIG. 7. The climatologic biases in March-April-May (MAM) in SST (shading in the first and third rows), SLP (shading in the second and bottom rows), surface wind (vectors) and precipitation (green contours, green solid lines and purple dash-dot lines represent positive and negative values, respectively, contour interval is 2 mm/day) and the changes of RDA experiments. (a)(d) the observational climatology in MAM, (b)(e) the bias of CTRL, and the changes of RDA experiments in (c)(f) ADA20, (g)(j) ADA20N, (h)(k) ADA20S. (i)(l) is the linear superposition of (g)(j) ADA20N-CTRL and (h)(k) ADA20S-CTRL. Only the significant changes based on T test at a 95% significance level is drawn.
FIG. 8. The evolution of the changes (minus CTRL) in Hadley circulation (vector), atmospheric moisture static energy (shading in the atmosphere), ocean temperature (shading in the ocean) and meridional ocean current (black contour) in the tropical Atlantic basin (60°W-10°E) in ADA20N. (a)-(f) are the changes in the 1st month, 2nd month, 5th month, 9th month 125th month and the changes in climatology, respectively. Solid and dashed contours in the ocean in (a)-(f) represent anomalous northward and southward ocean current, respectively. The contour interval is 0.5 cm/s in (a)-(e) and 0.2 cm/s in (f). In (a)-(f), green dashed lines represent the ITCZ location (the maximum precipitation location) in CTRL, grey dashed lines indicate the assimilation boundary of 20°N and the purple dotted lines represent the location of equator. The vertical axis in the ocean uses logarithmic coordinates to show the mixing-layer oceanic changes more clearly. Only the significant changes based on T test at a 95% significance level is drawn.
FIG. 9. The evolution of zonal-mean changes (minus CTRL) in SST (shading), surface wind (vector), precipitation (colored contour) and latent heat flux (grey contour) in the tropical Atlantic basin (60°W-10°E) in (a)(b) ADA20N and (c)(d) ADA20S. (a)(b) are the changes of ADA20N (a) in the initial 36 months and (b) the changes in the climatology, respectively. (c)(d) are the changes of ADA20S (c) in the initial 36 months and (d) the changes in the climatology, respectively. Green solid contours and magenta dashed contours in (a)-(d) represent positive and negative precipitation anomaly, respectively. The colored contour interval of precipitation anomaly is 1 mm/day. Grey solid contours in (a)-(d) represent positive (downward, warming the ocean) latent heat flux anomaly (only positive value shown here due to the too much information on this figure and unimportance of negative value). The grey contour interval of latent heat flux anomaly is 10 W/m². Black curves in (a)-(d) represent the seasonal cycle of ITCZ location (the maximal precipitation location) in CTRL. Only the significant changes based on T test at a 95% significance level is drawn.
FIG. 10. The evolution of changes (ADA20N-CTRL) in latent heat flux (positive value represents downward anomaly) and SST in the northern tropical Atlantic (averaged between 5°N–15°N and 60°W–0°) in ADA20N in initial 36 months. (a) shows the influence of surface wind changes on the changes of latent heat flux. All the lines in (a) shows the changes of latent heat flux in ADA20N compared with CTRL, the difference is the way to calculate the latent heat flux of ADA20N. The latent heat flux of ADA20N in back solid line is from model output, in black dashed line is recalculated offline, in blue solid line is recalculated offline using all variables from ADA20N except for surface wind from CTRL and in blue dashed line is recalculated offline using all variables from CTRL except for surface wind from ADA20N. (b) shows the relationship between the changes of latent heat flux (blue line corresponding to the left blue axis) and the changes of SST (red line corresponding to the right red axis). The months in which the SST anomaly (ADA20N-CTRL) is warming and the latent heat flux anomaly is positive is shaded in light red.

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FIG. 11. Schematic of the extratropical influence on the equatorial Atlantic zonal SST gradient bias.
Appendix A

FIG. A1. The changes (minus CTRL) in the climatologic seasonal cycle of surface wind (vectors), SST (shading) and thermocline depth (purple contours, solid and dashed lines represent positive and negative values, respectively, contour interval is 2 m) in (a) ADA30, (c) ADA30N and (d) ADA30S along the equatorial Atlantic (5°S–5°N). (b) is the linear superposition of (c) ADA30N-CTRL and (d) ADA30S-CTRL. Only the significant changes based on T test at a 95% significance level is drawn.
FIG. A2. The biases (minus OBS) in the climatologic seasonal cycle of the surface wind (vectors), SST (shading) and thermocline depth (purple contours, solid and dashed lines represent positive and negative values, respectively, contour interval is 2 m) in (a) ADA20, (b) ADA20N and (c) ADA20S along the equatorial Atlantic (5°S–5°N).