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Weakening AMOC connects Equatorial Atlantic and Pacific interannual variability

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Abstract Observations indicate that since the 1970s Equatorial Atlantic sea surface temperature (SST) variations in boreal summer tend to modulate El Niño in the following seasons, indicating that the Atlantic Ocean can have importance for predicting the El Niño–Southern Oscillation (ENSO). The cause of the change in the recent decades remains unknown. Here we show that in the Bergen Climate Model (BCM), a freshwater forced weakening of the Atlantic meridional overturning circulation (AMOC) results in a strengthening of the relation between the Atlantic and the Pacific similar to that observed since the 1970s. During the weakening AMOC phase, SST and precipitation increase in the central Equatorial Atlantic, while the mean state of the Pacific does not change significantly. In the Equatorial Atlantic the SST variability has also increased, with a peak in variability in boreal summer. In addition, the characteristic timescales of ENSO variability is shifted towards higher frequencies. TheBCM version used here is flux-adjusted, and hence Atlantic variability is realistic in contrast to in many other models. These results indicate that in the BCM a weakening AMOC can change the mean background state of the Tropical Atlantic surface conditions, enhancing Equatorial Atlantic variability, and resulting in a stronger relationship between the Tropical Atlantic and Pacific Oceans. This in turn alters the variability in the Pacific.

Keywords El Niño–Southern Oscillation · Atlantic Zonal Mode · Atlantic Niño · Atlantic Meridional Overturning Circulation · Bergen Climate Model · Interannual variability

1 Introduction

The El Niño–Southern Oscillation (ENSO) is the dominant mode of natural climate variability on interannual timescales and one of the most important sources of variability in the Tropical Pacific (e.g., Philander 1990; Latif and Keenlyside 2009). ENSO-like variability also exists in the Tropical Atlantic and is commonly referred to as the Atlantic Zonal Mode or Atlantic Niño (e.g., Zebiak 1993; Latif and Grötzner 2000; Keenlyside and Latif 2007; Jansen et al. 2009). However, variability in the Atlantic is weaker than in the Pacific and it peaks in boreal summer, rather than in boreal winter (Keenlyside and Latif 2007). In this study, we investigate the relationship between the Pacific and the Atlantic Niños.

ENSO can affect climate, ecology and economy in the Tropical Pacific region, as well as weather and climate globally (Brönnimann 2007). An example of such widespread effects is the warm phase of ENSO in the winter of

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1997/1998 that contributed to a high global surface temperature record in 1998 (Latif and Keenlyside 2009). There have been numerous studies on how the global climate is affected by ENSO, for instance the effect of ENSO on the extratropics (Wallace and Gutzler 1980; Hoskins and Karoly 1981), and on remote regions such as the Atlantic (e.g., Zebiak 1993; Enfield and Mayer 1997; Latif and Grötzner 2000) and Indian Oceans (e.g., Venzke et al. 2000). Predicting ENSO is therefore of considerable practical importance.

Recent studies have also shown that variability in the Atlantic and Indian Oceans can impact Pacific variability, and this knowledge could enhance ENSO prediction. In particular, sea surface temperature (SST) variations in these two regions may impact ENSO through atmospheric teleconnections (Dommenget et al. 2006; Jansen et al. 2009; Rodriguez-Fonseca et al. 2009). Thus, there is a two-way relationship between the Pacific and the Atlantic. However, the Atlantic Zonal Mode dominates the variations in the Equatorial Atlantic, and there is less direct impact by ENSO (Chang et al. 2006). Keenlyside and Latif (2007) showed that the correlation between SST anomalies in the central Equatorial Pacific and the central Equatorial Atlantic is strongest when theAtlantic is leading the Pacific with 6 months, implying a possible feedback from the Atlantic to the Pacific. Wang (2006) indicate that there is an inter-basin feedback that enhances the relation between the basins. Jansen et al. (2009) found that a cold (warm) Atlantic is associated with a warming (cooling) in the eastern Equatorial Pacific and a deepening (shallowing) of the average Equatorial Pacific thermocline about 6 months later. Frauen and Dommenget (2012) coupled an atmospheric general circulation model with a simplified ENSO recharge oscillator ocean model and found that the initial conditions of the Tropical Atlantic has a strong impact on ENSO predictability. Keenlyside et al. (2013) further show that observed ENSO predictions initiated in February can be enhanced by accurate knowledge ofEquatorial Atlantic SST. North Tropical Atlantic SST may also influence ENSO variability (Ham et al. 2013). These studies suggest that the variability in the Atlantic Ocean may be important for both increasing the understanding of ENSO variability and predicting ENSO.

Rodriguez-Fonseca et al. (2009) support that there is a connection between the Equatorial Atlantic and the Pacific, with Atlantic SST anomalies leading Pacific SST anomalies by 6 months, although they also indicate that the strength of this connection varies in time. Based on observations, they propose that warm (cold) SST anomalies in boreal summer associated with the Zonal Mode events drive a strengthening (weakening) of the equatorial Trade Winds over the Pacific through the Bjerknes positive feedback (Bjerknes 1969) leading to La Niña (El Niño) like SST anomalies. Ding et al. (2012) reproduced the relationship and the teleconnection mechanism suggested by Rodriguez-Fonseca et al. (2009) with a partially coupled general circulation model. Recent studies (e.g., Keenlyside and Latif 2007; Rodriguez-Fonseca et al. 2009) have indicated a strengthening of the relationship between the Atlantic Zonal Mode and ENSO since the 1970s, but causes for a change in this relationship are still unknown. Several possibilities have been proposed: stochastic changes (Ding et al. 2012), or changes in the background state of the Pacific and the Atlantic (Rodriguez-Fonseca et al. 2009). Here we investigate if changes induced by a weakening Atlantic meridional overturning circulation (AMOC) can impact this relationship.

Model studies and sediment cores have indicated that an increased freshwater flux to the Arctic Ocean associated with melting of ice sheets can affect the general circulation in the atmosphere and the ocean, such as the thermohaline circulation in the Atlantic Ocean and the associated AMOC. Records also indicate that these changes in the AMOC could have played an important role in historic climate change (Broecker et al. 1985; Ganopolski and Rahmstorf 2001). In addition, numerous climate modeling studies have shown that an increase of greenhouse gas concentrations in the atmosphere leads to an increase of freshwater fluxes from the Arctic to the Atlantic (Räisänen 2001). This may in turn affect global climate through for instance changes in the AMOC (e.g., Stouffer et al. 2006). However, there are uncertainties and large variations in the estimates of the overall effect of an increased freshwater flux to the Atlantic in past and future climate scenarios (Schmittner et al. 2005; Kuhlbrodt et al. 2007).

Several recent studies (Dong and Sutton 2002; Zhang and Delworth 2005; Dong and Sutton 2007; Sutton and Hodson 2007), have investigated how an increased freshwater flux to the Atlantic and associated changes in the thermohaline circulation and AMOC can induce changes in ENSO. These studies have explained the response in ENSO by, for instance, atmospheric bridges over Central America and westward propagating Rossby waves excited in the Atlantic. All these studies identify significant changes in the mean state of the Tropical Pacific, and indicate that these induce ENSO changes.

In the present study, we investigate how a weakening AMOC, resulting from an increase in freshwater runoff from the Arctic region into the North Atlantic Ocean, and the associated change in the Atlantic mean state, can change the relationship between the Tropical Atlantic and Pacific regions. Our hypothesis is that a weakening AMOC may strengthen the relationship between Atlantic and Pacific variability. The present study is based on previous studies (Otterå et al. 2004; Breiteig 2009) using a global coupled atmosphere–ocean-sea ice model, the Bergen climate model (BCM).

The paper is organized as follows. In Sect. 2 the model and the data are described. The results are presented in Sect. 3. In Sect. 4 the results and possible mechanisms connecting the Atlantic and the Pacific are discussed and summarized.

2 Data: model and experimental design

BCM (Furevik et al. 2003), a coupled atmosphere–ocean-sea ice model, consists of the atmospheric general circulation model ARPEGE/IFS (Deque et al. 1994) and the ocean model MICOM (Bleck et al. 1992). The atmosphere model is run with a linear T_L63 grid. It has a horizontal resolution of about 2.8 × 2.8 degrees for physical processes and has 31 vertical layers from the surface up to 10 hPa. The ocean model has a non-isopycnic surface mixed layer on top of 23 isopycnic layers, and has a horizontal resolution of 2.4 × 2.4 degrees, which increases to 0.8 degrees near the equator. The ocean model is coupled with a dynamic and thermodynamic sea ice model. The atmosphere and ocean components are coupled with the OASIS coupler version 2.2 (Terray et al. 1995, 1998), and exchange data once every day. The heat and freshwater fluxes are adjusted based on a flux-correction to avoid drifting from the climatological SST and sea surface salinity fields (Furevik et al. 2003).

Since the BCM version used in this study is flux-adjusted, it is able to model an Atlantic cold tongue, something many general circulation models (GCM) fail to reproduce (Davey et al. 2002; Richter and Xie 2008). Hence, in BCM the Tropical Atlantic mean state and features of the Atlantic Zonal Mode are fairly well simulated compared to in many other GCMs (Fig. 1), as may be expected because the feedbacks responsible for the zonal mode depends on the mean state (Keenlyside and Latif 2007; Ding et al. 2010). However, Latif et al. (2001) showed that flux-corrected GCMs do not necessarily improve the state of the Tropical Pacific, but un-corrected models generally still simulate a cold tongue. In general, BCM simulatesrealistic ENSO characteristics, in strength, frequency and spatial pattern, but as common in many GCMs the Pacific cold tongue is slightly too narrow and stretches too far west compared to observations (Furevik et al. 2003).

Otterå et al. (2004) performed a sensitivity experiment with BCM, consisting of two model simulations. One control simulation was carried out with greenhouse gas concentrations and aerosol particle concentrations kept constant at present-day values. This gave a simulated freshwater runoff to the North Atlantic Ocean of 0.1 Sv (1 Sv = 10^6 m^3 s^{-1}) equal to estimates of today's observed climatological situation. A second so-called freshwater experiment (FW) was carried out with an artificially increased freshwater runoff of 0.4 Sv (Otterå et al. 2003, 2004). This increased runoff is of a size comparable to the runoff found for earlier deglaciations (Simonsen 1996) and also to model estimates for a situation where the greenhouse gas concentration of the preindustrial atmosphere is quadrupled (Manabe and Stouffer 1994). The control run was integrated for a 300-year period. FW was initiated at year 100 of the control run and integrated for 150 years. The freshwater flux increase in FW was located at coastal regions in the Nordic seas and the Arctic Ocean, and was added continuously throughout the 150-year integration. The freshwater flux was added artificially to the internal hydrological cycle, so the experiment can only be considered as a sensitivity experiment (Otterå et al. 2004).

In FW the AMOC weakens by about 6 Sv in the first third of the integration (Otterå et al. 2003, 2004). The strength of the AMOC, defined as the maximum overturning across the 24°N latitude belt in the Atlantic, for the control run (thick line) and FW (thin line) is illustrated in Fig. 2, where the grey box indicates the weakening phase. A weakening of the AMOC due to increased freshwater runoff to the Atlantic is consistent with findings from other GCM studies (e.g., Manabe and Stouffer 1997; Rind et al. 2001). However, BCM shows a weakening of the AMOC only in the first third of the integration period after which the AMOC gradually recovers to the same level as in the control run, whereas most studies show a reduction or a shutdown of the AMOC through the wholeintegration period (Otterå et al. 2004). Furthermore, the magnitude of the weakening is smaller in BCM compared to other similar studies, possibly related to a weaker freshwater forcing or different locations of the increased freshwater fluxes. Otterå et al. (2004) suggest that the weakened AMOC is caused by a reduction of the rates of deepwater formation in the Nordic Seas and the North Atlantic subpolar gyre, and a reduction of the southward flow of intermediate water masses through the Fram Strait. The strengthening of the AMOC with time is suggested to be caused by increased basin-wide upwelling in the Atlantic Ocean, northwards transport of saline waters from the western tropical North Atlantic, and a surface wind pattern maintaining the Atlantic water inflow between the Faroes and Scotland to the Nordic Seas (Otterå et al. 2004).

In this study we compare monthly mean data from the subperiod of FW where the AMOC weakens with a 150-year period of the control run that overlaps the period when FW was integrated. This 150-year subperiod in the control run is hereinafter referred to as CTRL. The subperiod in FW when the AMOC weakens is defined as the period from year 7 to 40 in FW. This 34-year period is referred to as DECLINE, illustrated by the grey box in Fig. 2. This DECLINE period is chosen because at year 7

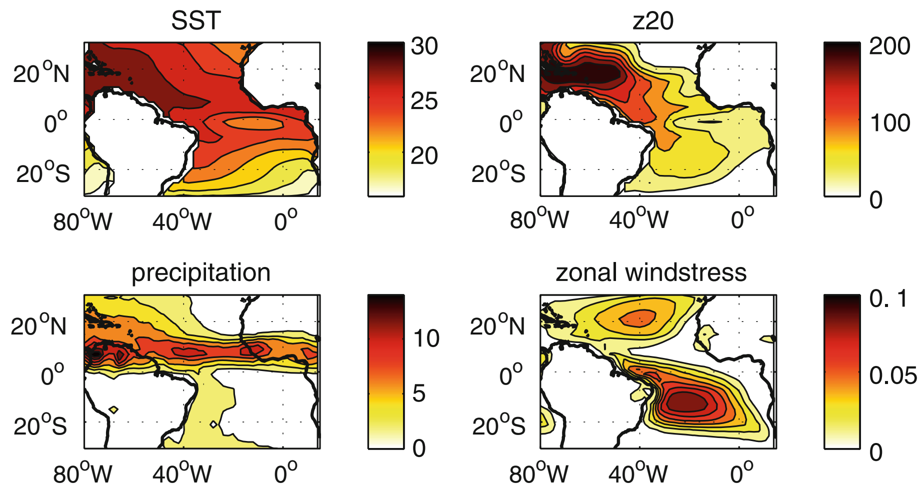

Fig. 1 Mean SST (°C), thermocline depth (m), convective precipitation (mm day^{-1}) and zonal windstress (N m^{-2}) in the Tropical Atlantic for summer (June-July-August) in CTRL. Positive zonal wind direction is towards the east

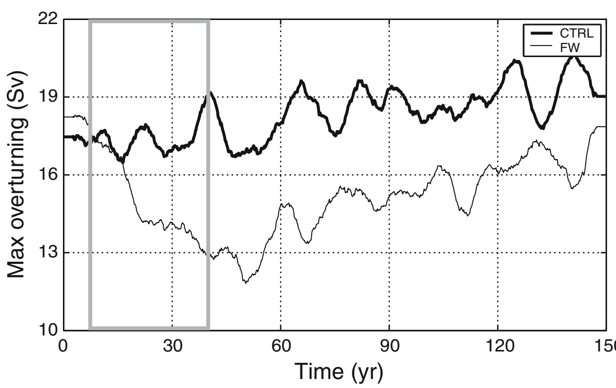

Fig. 2 5-year running mean of the time series of the maximum AMOC (Sv) in the control run (*thick line*) and FW (*thin line*). The *grey box* indicates the phase when the AMOC is weakening (DECLINE). From Breiteig (2009)

the AMOC has achieved a relatively steep weakening rate, and by year 40 the AMOC shows the first tendency to strengthen again (Breiteig 2009). The strengthening AMOC phase is defined as the period from year 50 throughout the integration in FW, since this is after the AMOC reaches its minimum value in the maximum overturning. This recovering phase in FW is referred to as RCVR.

3 Results

3.1 Strengthened relation between Equatorial Atlantic and Pacific variability

To investigate the connection between the Equatorial Atlantic and Pacific, the cross-correlation between SST anomalies in the Atl3 region (3°S–3°N; 20°W–0°E) and the Niño 3.4 region (5°S–5°N; 170°W–120 W) is calculated for CTRL and DECLINE (Fig. 3). We define anomalies by removing the seasonal cycle. For comparison the corresponding cross-correlations for SST data from HADISST (Rayner et al. 2003) from 1870 to 2011 and 1970 to 2011, and for RCVR are included. For DECLINE correlations above 0.15 are statistically significant at a 95 % confidence level using a Fisher's Z test.

For all datasets, there is a negative correlation when Atlantic SSTs lead Pacific SSTs with about six months, and a positive correlation when Pacific SSTs lead Atlantic SSTs with about six months. The negative correlation, when the Atlantic leads the Pacific, is three times as strong in DECLINE compared to in CTRL. Using a Fisher's Z-transformation, negative correlations stronger than −0.14 at a six-month lead of the Atlantic in DECLINE is significantly different from CTRL at a 95 % confidence level. To quantify the uncertainty, we calculated the cross-correlation between Equatorial Atlantic and Pacific SST anomalies for 100 randomly selected 34-year periods from CTRL. A 95 % confidence interval for the CTRL ensemble correlation for a six-month lead of the Atlantic is then [−0.15, −0.12]. We repeated the cross correlation for 1000 randomly selected 34-year periods from CTRL, and found that the 2.5-percentile (−0.22) is weaker than the correlation in DECLINE. We conclude that thecorrelation of −0.29 found in DECLINE is significantly stronger than what is found in CTRL. In contrast, the difference in correlation between DECLINE and CTRL is smaller when the Pacific is leading the Atlantic. The relation between Atlantic and Pacific SST anomalies in DECLINE is similar to that observed during the period 1970 till present. However, stronger correlations are observed when the Pacific leads. We suggest that there exists a physical explanation for the difference in the correlation between CTRL and DECLINE, rather than mainly being a statistical property.

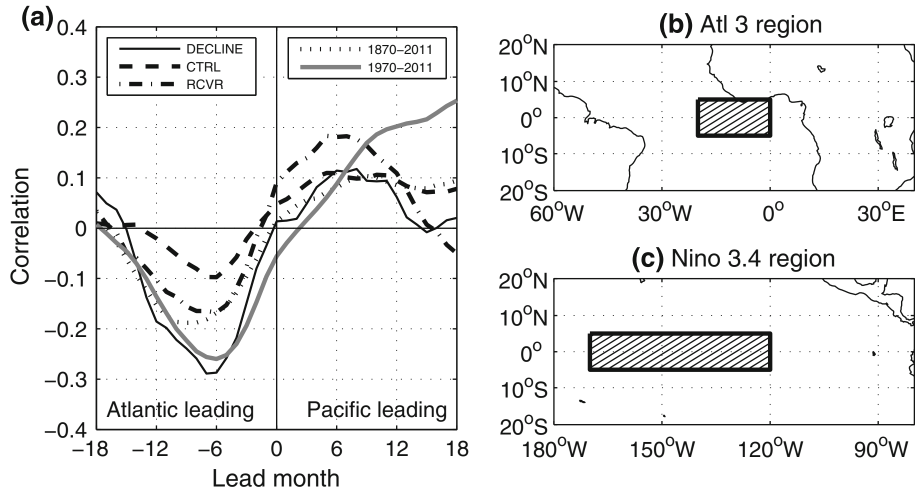

Fig. 3 **a** Cross-correlation of monthly SST anomalies in the Atl3 and Niño 3.4 regions for DECLINE (*solid black line*), CTRL (*dashed line*), RCVR (*dash-dotted line*), and from HADISST data from 1870 to 2011 (*dotted line*) and from 1970 to 2011 (*solid grey line*). Correlations for DECLINE >0.15 are significant at a 95 % confidence level. **b** The Atl3 region and **c** the Niño 3.4 region indicated by the *hatched boxes*

3.2 Pacific and Atlantic mean state

In DECLINE there is a distinct shift in the frequency of ENSO (Breiteig 2009), to a more pronounced 2-year period (Fig. 4). The power spectra of SST anomalies in the Niño 3.4 region indicate a peak in frequency at about 36 months in CTRL, while in DECLINE the peak is shifted to 25 months. To test the strength of the biennial mode of ENSO in DECLINE, a one-year lag correlation of the yearly winter (December-January–February) SST anomalies in the Niño 3.4 region is calculated. The correlation coefficient of $R = -0.55$ is statistically significant at a 95 % confidence level ($p = 0.009$). The corresponding correlation coefficient for CTRL is $R = -0.06$. Monte Carlo simulations indicate that there is a likelihood of less than 0.06 % for a correlation coefficient as large as in DECLINE to be achieved by chance. We conclude that the shift in frequency is statisticallysignificant. In addition, the ENSO frequency shifts back to the value in CTRL when the AMOC strengthens again (Breiteig 2009). However, the significance of these changes is difficult to assess given the shortness of our record (Wittenberg 2009). In the Tropical Atlantic there is seemingly no significant change in the power spectrum of the Atl3 index as a response to the weakening AMOC (not shown).

There is no significant difference in the mean state of the Equatorial Pacific between CTRL and DECLINE (Fig. 5). The correlation of the annual average SST, precipitation, zonal wind stress and thermocline depth in the Equatorial Pacific between CTRL and DECLINE is high for all variables, with a pattern correlation coefficient R close to unity, and a small root mean square error. This illustrates that the shift in ENSO frequency may not be attributed to a change in the mean state of the Equatorial Pacific.

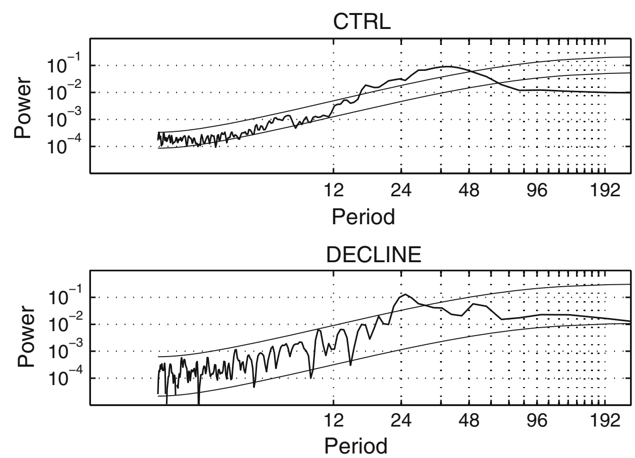

Fig. 4 Power spectrum of monthly SST anomalies in the Niño 3.4 region for CTRL (*upper*) and DECLINE (*lower*). The *smooth black lines* indicate the 95 % uncertainty range of a red-noise process. The period in months is shown along the *lower axis*

Figure 6 shows the standard deviation of monthly SST in the Atl3 region for CTRL, DECLINE and RCVR, where the long-term mean has been removed prior to the calculation. SST data from HADISST from 1870 to 2011 are also included. The seasonal cycle of Equatorial Atlantic SST variability in BCM is substantially stronger than observed. In DECLINE the variability has increased during boreal spring and summer compared to in CTRL. For all three BCM datasets there is a peak in variability in April that is not observed. However, in DECLINE and RCVR there is a second peak in boreal summer that tends to coincide with observations. The additional large June peak in DECLINE is not present in CTRL or RCVR, and the variability in June is 20 % larger in DECLINE than in CTRL. This increase in variability is statistically significant at a 90 % confidence level by a two-tailed F-test.

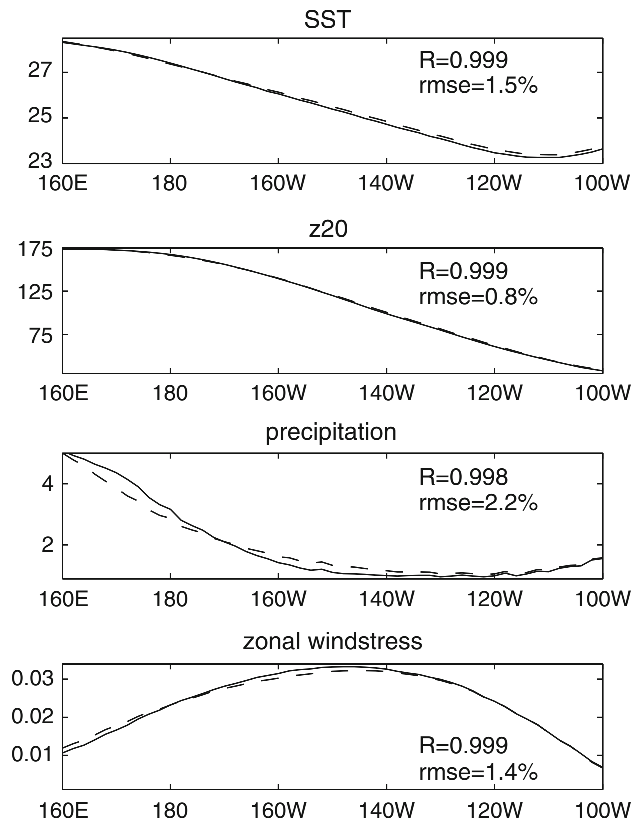

Fig. 5 Annual mean Equatorial Pacific SST (°C), thermocline depth (m), convective precipitation (mm day^{-1}) and zonal windstress (N m^{-2}), averaged between 5°S–5°N for CTRL (*dashed line*) and DECLINE (*solid line*). The correlation (R) and the normalized root mean square errors (rmse) between DECLINE and CTRL are shown for each variable

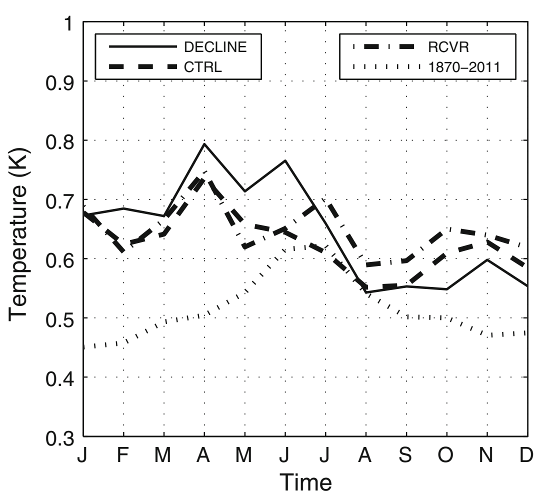

Fig. 6 Standard deviation of SST (°C) anomalies in the Atl3 region for DECLINE (*solid line*), CTRL (*dashed line*), RCVR (*dashed-dotted line*) and from HADISST data from 1870 to 2011 (*dotted line*)

Furthermore, by taking the standard deviation of SST for June for 100 randomly selected 34-year periods from CTRL, we find that the standard deviation in DECLINE is larger than the 95th-percentile. We conclude that the increased variability in DECLINE is significantly larger than in CTRL at a 90 % confidence level.

In DECLINE, the peaks in SST variability in April and June coincide with the months when the monthly mean Equatorial Atlantic SST response to the weakening AMOC, defined as DECLINE-CTRL, is at its maximum (shown for June in Fig. 7). The response shows that SSTs are higher in the central Atlantic region in DECLINE compared to in CTRL. There is also a slightly negative response in the eastern Equatorial Atlantic from April to September (not shown). Combined this indicates a stronger zonal SST gradient in the Equatorial Atlantic. Figure 7 also shows the June response for precipitation, zonal wind stress and thermocline depth. Thermocline depth (z20) is defined as the depth of the 20 °C isotherm. There is a deepening of the thermocline in the central Equatorial Atlantic indicating an increase in ocean heat content. There is also a dynamically consistent increase in precipitation and wind stress in thisregion, coinciding with the location of the positive SST response. The warming in the upper Tropical Atlantic may be due to a slow-down of the North Atlantic subpolar gyre caused by the increased freshwater flux to the Arctic Ocean, suppressing deepwater formation. This slow-down increases the travel time in the Guyana Current, increasing the temperature and salinity in the upper 600 m of the water column here (Otterå et al. 2003, 2004).

3.3 Connecting the Tropical Atlantic and Pacific regions

How does the atmospheric circulation communicate the response from the Atlantic to the Tropical Pacific, and which dynamical processes can explain the correlation between Atlantic and Pacific SST anomalies? We investigate the previously proposed hypothesis that boreal summer Atlantic SST can influence the Trade Winds over the Pacific, and that the Bjerknes positive feedback subsequently enhances these anomalies (Rodriguez-Fonseca et al. 2009; Ding et al. 2012).

A regression analysis is performed for SST and precipitation anomalies (Fig. 8), and surface wind and thermocline depth anomalies (Fig. 9) regressed on to the summer (June-July–August) Atl3 index. This is done for boreal summer (June-July–August) and the following winter season (December–January–February) for DECLINE and CTRL. Only the regions where the correlation is statistically significant at the 90 % confidence level under a *t* test are shown. The patterns in the Equatorial Atlantic in boreal summer are in good agreement with the observed Zonal Mode structure in both experiments (e.g., Ding et al. 2012).

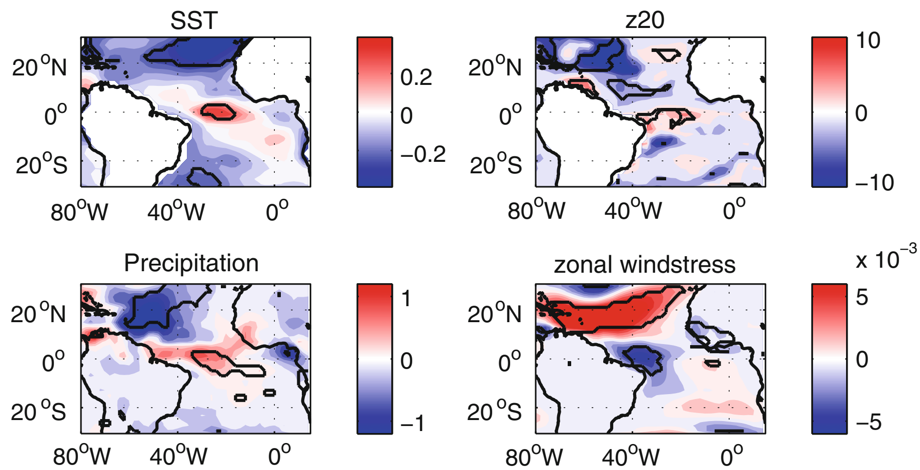

Fig. 7 Mean June response (DECLINE-CTRL) in SST (°C), thermocline depth (m), convective precipitation (mm day^{-1}) and zonal windstress (N m^{-2}) in the Tropical Atlantic. The *dashed contours* indicate the 90 % confidence level. Positive zonal wind direction is toward the east

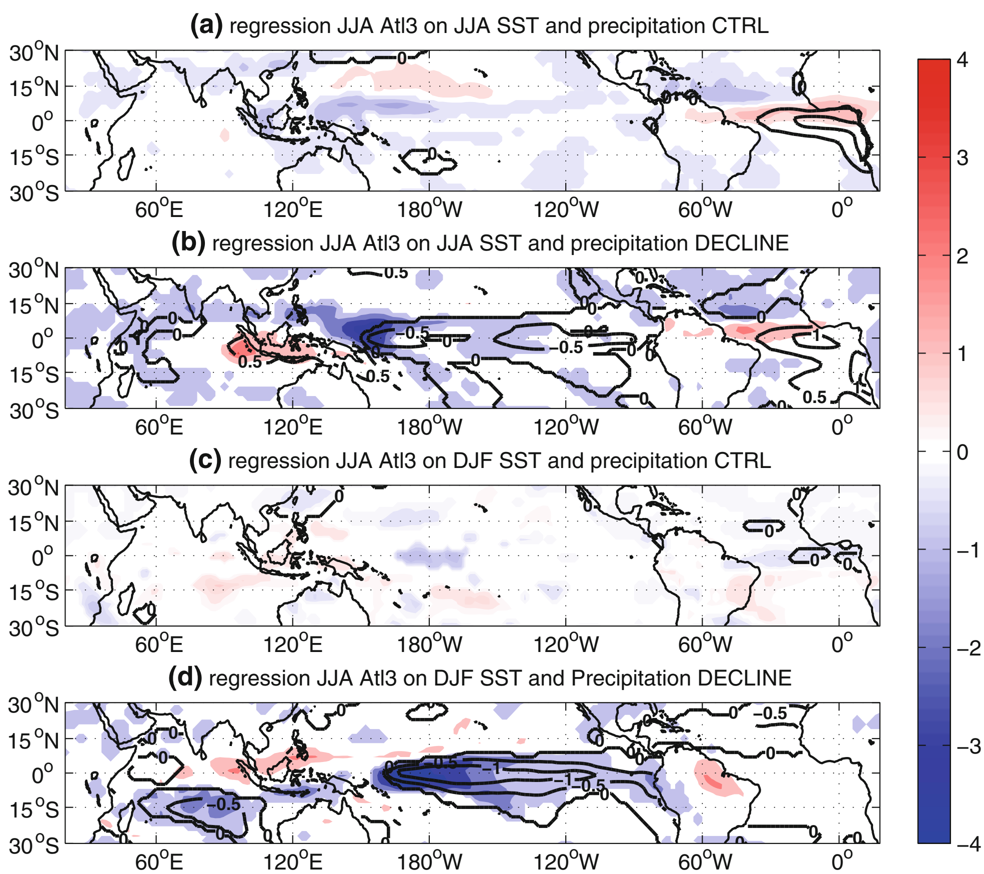

Fig. 8 Regression of SST (°C) (*contours*) and precipitation (mm day^{-1}) (*shaded*) for (**a** and **b**) boreal summer (June-July–August) and (**c** and **d**) winter (December-January–February) on the summer (June-July-August) Atl3 index for CTRL (**a** and **c**) and DECLINE (**b** and **d**). Only the regions that are statistically significant at a 90 % confidence level under a *t* test for the effective degrees of freedom are shown

For DECLINE, there is a negative relationship between the Atl3 summer index and SST, precipitation, zonal surface wind and thermocline depth in the Equatorial Pacific in boreal summer. Thus, in DECLINE when the Equatorial Atlantic is warm in summer, there is a tendency for enhanced Trade Winds over the western Pacific, lifting the thermocline to the east and decreasing SSTs. In DECLINE, in the following winter, the negative relationship between Equatorial Atlantic SST anomalies in summer and Tropical Pacific SST, precipitation, thermocline depth and surface wind strengthens. This is consistent with ENSO dynamics and the Bjerknes feedback.There is a negative relationship between the summer Atl3 index and central Tropical Pacific SST anomalies, west-to-central Tropical Pacific convective precipitation anomalies, east Pacific thermocline depth anomalies and surface wind in winter. In CTRL the relationship between the summer Atl3 index and the Equatorial Pacific is weak and shows nosignificant changes

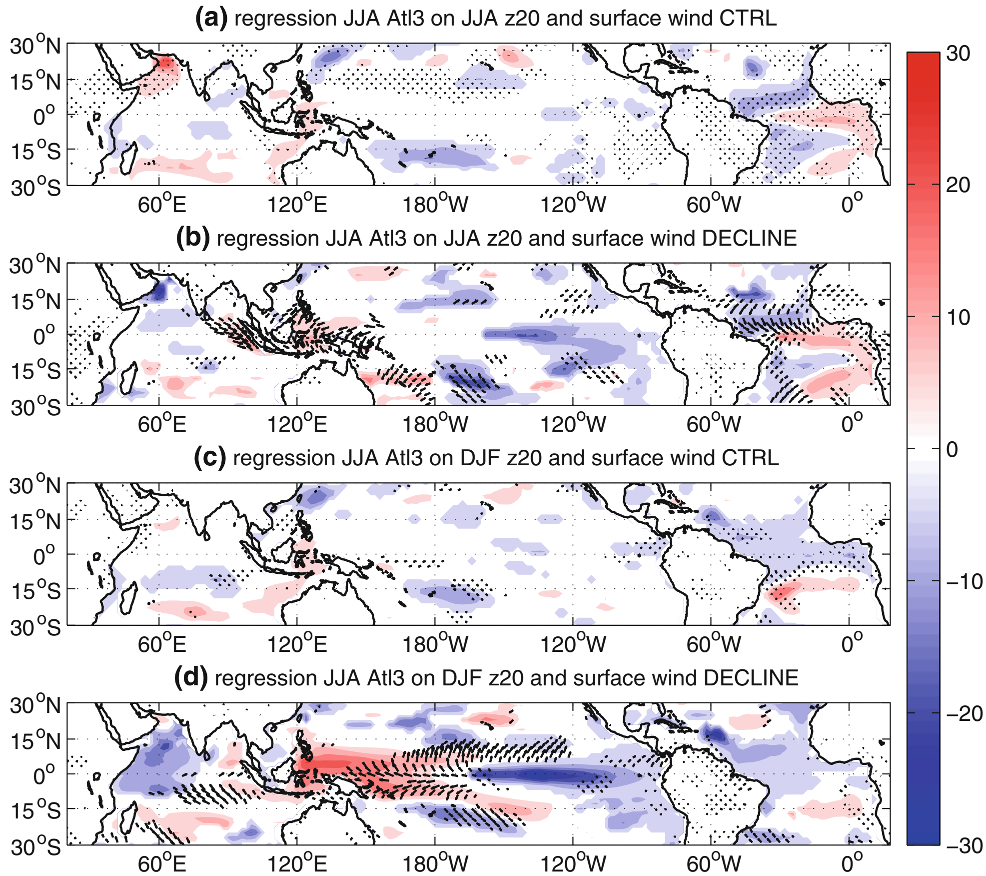

Fig. 9 Regression of thermocline depth (m) (*shaded*) and surface wind (ms^{-1}) (*vectors*) for (**a** and **b**) boreal summer (June–July–August) and (**c** and **d**) winter (December–January–February) on the summer (June-July-August) Atl3 index for CTRL (**a** and **c**) and DECLINE (**b** and **d**). Only the regions that are statistically significant at a 90 % confidence level under a *t* test for the effective degrees of freedom are shown

from summer to winter. For comparison, a similar regression analysis done on observational data is shown in Rodriguez-Fonseca et al. (2009).

4 Summary and discussion

Our results indicate that the change in the Atlantic due to a weakening AMOC can induce a stronger connection between the Atlantic and Pacific SSTs, with the Atlantic leading the Pacific by about half a year. In the Atlantic there is a peak in SST variability in boreal summer, while in the Pacific this peak is in boreal winter. In the model, when there is a cold (warm) anomaly in summer in the Atlantic, there is a tendency for a warm (cold) anomaly in the Pacific the following winter. The correlation when the Atlantic leads the Pacific is stronger in DECLINE than in CTRL. Thus, the increased variability in the Equatorial Atlantic during summer strengthens the Atlantic–Pacific connection.

Previous studies suggest warm Atlantic Zonal Mode events in summer strengthen the zonal atmospheric circulation, with rising air in the eastern Equatorial Atlantic and sinking air in the central Equatorial Pacific. The divergence shallows the equatorial thermocline and cools the SST in the eastern Pacific. These anomalies are amplified by a coupled Bjerknes feedback process, causing weak La Niña-like anomalies in the winter. Similarly, a cold-event in the Atlantic in summer induces warm anomalies in the Pacific the following winter (Rodriguez-Fonseca et al. 2009; Ding et al. 2012). In our case, the zonal SST gradient in the Equatorial Atlantic has increased in DECLINE. This response can be associated with a westward shift of the Walker circulation and a strengthening of the ascending branch over the western Equatorial Atlantic. Our results are consistent with the explanation for the connection between equatorial Atlantic and Pacific SST anomalies discussed in Rodriguez-Fonseca et al. (2009).

Rodriguez-Fonseca et al. (2009) investigated the connection between the Tropical Atlantic and Pacific regions in observations for the period before and after 1979, and found that the connection has increased in the later period compared to the former. In observations from before 1979 the Pacific ENSO seems to have no apparent relationship with the Atlantic Zonal Mode, similar to what is observed in CTRL. However, after 1979, the Atlantic and the Pacific seem to have opposite phases with the Atlantic leading the Pacific, as in DECLINE. Tokinaga and Xie (2011) showed that since the 1950s there has been a reduction in the zonal SST gradient in the Equatorial Atlantic and a decrease in the interannual variability of the longitudinal SST gradient (i.e. a weakening of the Atlantic Zonal Mode and a weakening of the equatorial cold tongue). In this same period SSTs have increased over the wholeEquatorial Atlantic. The weakening of the Zonal Mode can be explained by the associated deepening thermocline in the east and thus, reducing the thermocline feedback (Tokinaga and Xie 2011). This is opposite to what was simulated in DECLINE, where the SST gradient strengthened, the thermocline shoaled in the east, and the Zonal Mode variability increased. In DECLINE both the connection between the Equatorial Atlantic and Pacific has strengthened and the Equatorial Atlantic variability has increased, while in observations after 1979 the connection has also strengthened but the Atlantic variability has reduced. Thus, the mechanism found here likely does not explain the stronger Atlantic-Pacific relation observed since 1979.

The effect of changes in AMOC on the Tropical Atlantic is still debated (Wen et al. 2010). Even though our results are not consistent with observations, the mechanism seen in this study may be of importance in other time periods. Whether AMOC induced changes could contribute to the strength of the Equatorial Atlantic-Pacific relation at other times remains unclear due to the lack of observations, and more work is required to understand the impact of future AMOC changes. Our results still suggest that the Atlantic-Pacific relationship depends on the state of the Tropical Atlantic, and that seasonal ENSO forecasts and the understanding of ENSO variability can be improved by including Tropical Atlantic surface conditions (Keenlyside et al. 2013).

In DECLINE, when the relation between Atlantic and Pacific SSTs is strengthened, the power of the two-year period of ENSO is also increased (Fig. 4). Thus, the strengthened variability in the Equatorial Atlantic during summer could potentially enhance an intrinsic dominant ENSO mode of the Pacific, strengthening the two-year ENSO period in DECLINE. Similar freshwater hosing experiments to the one presented here using other GCMs, have indicated an effect of Atlantic SSTs on ENSO (Dong and Sutton 2002; Timmermann et al. 2005; Zhang and Delworth 2005; Dong and Sutton 2007; Sutton and Hodson 2007), but in contrast to these studies, our results show a response in ENSO that may not be attributed to a response in the TropicalPacific mean state. The fact that we do not find significant responses in the Tropical Pacific mean state may be related to the weaker freshwater forcing and the resulting weaker response in the AMOC in our study compared to other similar freshwater hosing experiments. In addition, several of these previous studies have not commented directly on the connection between Equatorial Atlantic and Pacific SST anomalies, and the connection may not be present in these studies. The reason for this may be due to the poor simulation of Equatorial Atlantic variability in many present GCMs. As previously mentioned, the BCM version used in this study is flux-adjusted, and the Tropical Atlantic is therefore fairly well simulated compared to in many other GCMs (Davey et al. 2002; Richter and Xie 2008). However, it should be noted that flux-adjustments might change model sensitivity to freshwater perturbations,although Schiller et al. (1997) states that this may not be of significance.

Although we have not explored changes in the structure of ENSO events in this paper, there are indications that ENSO shifts to a more central-Pacific (CP) type ENSO in DECLINE compared to a more eastern-Pacific (EP) type ENSO in CTRL. The CP-ENSO has been associated with shorter oscillation periods (Kao and Yu 2009) and SST anomalies displaced to the west in the Tropical Pacific compared to the EP-ENSO (Kao and Yu 2009; Yeh et al. 2009). A shift from an EP-ENSO to a more CP-ENSO could explain the difference in ENSO frequency between DECLINE and CTRL. We also find SST anomalies shifted farther west in DECLINE compared to in CTRL consistent with the EP-CP definitions (not shown). Previous studies (e.g., Ham et al. 2013) have argued that North Tropical Atlantic SST anomalies in spring can influence thefrequency of CP events. Our model results suggest that Equatorial Atlantic SST in summer might also be important in this respect. The mechanism for how the Equatorial Atlantic variability can change the frequency of CP events is however unclear. We hypothesize that external forcing from the Atlantic may enhance an innate ENSO-mode with a two-year period.

To summarize, in BCM the relationship between the Equatorial Atlantic and Pacific depends on the mean state of the Tropical Atlantic. This study has shown that when the AMOC weakens (1) the variability increases in the Atlantic in boreal summer, possibly due to a stronger zonal mean SST gradient, (2) Atlantic Zonal Mode events tend to lead opposite signed ENSO-like anomalies by about six months, and (3) the two-year oscillation period in the Equatorial Pacific is more pronounced, while the mean state of the Pacific does not show significant changes. Regression analysis showed that Equatorial Atlantic and Tropical Pacific SST, surface wind and thermocline depth anomalies are related in boreal summer, and these anomalies are subsequently amplified in the Pacific by the Bjerknes positive feedback, accounting for the six-month lead of the Atlantic. Under control conditions the relation between Equatorial Atlantic and Pacific boreal summer anomalies is insignificant. Thus, theweakening AMOC triggers a stronger coupling between the Tropical Atlantic and Pacific regions.

We acknowledge that the length of DECLINE is quite limited, and that interannual variability in the Equatorial

Pacific may not have statistical robustness in short time series. The existence of a physical mechanism provides some confidence in our results. However, an ensemble of AMOC weakening experiments is required to assess the robustness of our findings. Nevertheless, we hope that our results motivate further work in this direction with more models.

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