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Fast teleconnections to the tropical Atlantic sector from Atlantic thermohaline adjustment

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[1] We examine the first decade of adjustment of the North Atlantic climate following an abrupt freshening of the high North Atlantic and resulting slowdown of the Atlantic meridional overturning circulation (AMOC) in a coupled model, with reference to previously proposed teleconnection mechanisms. After an initial ocean-driven cooling at the subtropical-subpolar gyre boundary, subsequent equatorward progression is dominated by an atmosphere-surface ocean mechanism driving evaporative cooling in the tropical North Atlantic, and resulting in an anomalous southward ITCZ displacement by year 2. The tropical cooling is countered by fast ocean baroclinic adjustment to the reduced overturning that increases the volume of tropical surface waters, as well as ocean circulation changes to the increased northeasterly trades. However, it is the atmosphere-surface ocean teleconnection that explains the hemispheric asymmetric nature of the tropical Atlantic surface climate impact to AMOC slowdown, underlining its crucial importance in high-to-low latitude teleconnections. **Citation:** Chiang, J. C. H., W. Cheng, and C. M. Bitz (2008), Fast teleconnections to the tropical Atlantic sector from Atlantic thermohaline adjustment, *Geophys. Res. Lett.*, *35*, L07704, doi:10.1029/2008GL033292.

1. Introduction

[2] The pronounced southward displacement of the tropical Atlantic ITCZ and associated meridional sea surface temperature (SST) gradient change are robust climate responses to the slowdown of the Atlantic Meridional Overturning Circulation (AMOC), coming from both model simulation [*Stouffer et al.*, 2006] and paleoproxy [e.g., *Peterson et al.*, 2000; *Wang et al.*, 2004] perspectives. Previously proposed mechanisms provide a starting point to understanding this teleconnection. *Kawase* [1987] showed the basic baroclinic ocean adjustment that communicates the AMOC slowdown to the equatorial region through ocean Kelvin and Rossby waves, and variants of this mechanism have been invoked for teleconnecting high latitude North Atlantic climate change to the tropics [*Yang*, 1999; *Huang et al.*, 2000; *Cessi et al.*, 2004; *Timmermann et al.*, 2005]. A naïve application of this mechanism would predict warmer tropical sea surface temperatures symmetric about the equator after an AMOC slowdown, given a reduction to the northward flow of warm tropical waters and deepening the volume of surface water symmetrically about the equator. This is, however, at variance with coupled model simulations that clearly show an asymmetric response about the equator with cooler SST conditions over the entire North Atlantic (Figure 1).

[3] The atmosphere provides additional possibilities. *Dong and Sutton* [2002] argued for atmospheric feedbacks in the global climate adjustment to AMOC slowdown, but without explicitly discussing mechanisms. *Chiang and Bitz* [2005] showed in simulations with an atmospheric general circulation model (AGCM) coupled to a slab ocean that cooling in the high latitudes is readily communicated to the tropical oceans though atmospheric thermodynamic interaction with the ocean mixed layer; a mechanism resembling the subtropical wind-evaporation-SST (WES) originally proposed by *Xie* [1999] in the context of tropical Atlantic decadal variability. This mechanism readily provides a surface climate change in the AGCM–slab ocean configuration that resemble those seen in fully-coupled AMOC slowdown simulations; in particular the wide-spread cooling over the entire North Atlantic and southward ITCZ displacement. However, a cursory examination of the surface flux changes to AMOC slowdown (Figure 1, contours – the simulations giving these results are introduced in the next section) reveals that the widespread North Atlantic cooling belies a complex pattern of surface flux anomalies that help create it, with interdispersed regions of flux anomalies of both signs, indicating that both atmospheric and oceanic processes influence the teleconnection.

[4] In this paper, we examine the transient adjustment of the North Atlantic to the AMOC slowdown in a coupled general circulation model to understand the teleconnection dynamics to the tropical Atlantic. Climate changes in the North Atlantic to AMOC slowdown reach quasi-steady character by the first decade, and we limit our analysis to that time frame. This restriction, however, excludes the later adjustment in the equatorial and south tropical Atlantic that is the focus of P. Chang et al. (An oceanic teleconnection between abrupt changes in high-latitude North Atlantic climate and African monsoon, submitted to *Nature Geoscience*, 2008), which we mention in context.

2. Analysis of the North Atlantic Cooling to AMOC Slowdown

[5] The simulations with the Community Climate System Model 3.0 (CCSM3) that we analyze are the same as in the work by *Bitz et al.* [2007] and *Cheng et al.* [2007]. Six freshwater pulse experiments are branched from a 1,000-year modern-day control [*Collins et al.*, 2006] by instantaneously freshening the upper 970 m of the North Atlantic

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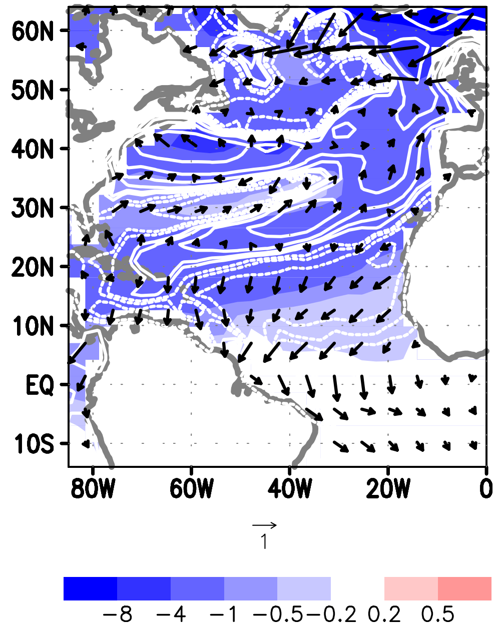

Figure 1. Winds (arrows; reference vector is 1 m/s) and SST (shaded; units are K) anomalies averaged over years 7 and 8, superposed on the surface flux pattern (contours, positive values are solid lines and defined to be *into* the ocean) averaged over years 1–7. The first positive contour is 2 W/m^2, followed by 4, 8, 16, 32, and 64; negative contours have the same magnitudes. By year 8, the north Atlantic SST anomalies have reached a quasi-steady state, and the tropical winds assume the familiar southwards cross-equatorial pattern.

and Arctic Oceans from 55–90°N, 90°W–20°E by an average of 2 psu (higher at the top and tapering with depth), similar to *Vellinga and Wood* [2002]. All ensemble members extend to at least 8 years, and three extend past 20 years. The anomalies discussed are differences between the perturbed ensemble mean, and the mean of the six 'unperturbed ensembles' in the control run years corresponding to the same years as the branched experiments.

[6] Almost immediately after the freshwater pulse, an intense surface ocean cooling – up to 6 K at the core - appears just off the east coast of North America around 40°N at the latitude of the subtropical/subpolar gyre boundary, and progresses directly eastwards to the opposite side of the basin over the course of the following two years (Figure 2a). Along this route the anomalous surface heat flux is into the ocean (Figure 1), indicating that the cooling originates from ocean circulation changes. This intense midlatitude cooling results from a weakening of both the subtropical and subpolar gyre circulations as inferred from the anomalous barotropic streamfunction (not shown). The gyre responses are thought to result from weakening of the bottom pressure torque where strong topographic features are present [*Vellinga et al.*, 2002], and is a direct response to the density changes resulting from the freshening. The cold SST anomaly progress eastward following the direction of the mean current, and turns poleward and equatorward at the eastern boundary at year 3. The weakening subpolar gyre western boundary current (i.e., the Labrador Current) also causes the warming occurring north of intense midlatitude cooling because of reduced cold advection; however, this warm SST anomaly is short-lived and reverses sign after year 2 (Figure 3a). The surface current anomalies (Figure 2c) also reflect a slowing of the northward surface flow of the AMOC along the western boundary extending from the North Atlantic into the South Atlantic, similar to what *Dong and Sutton* [2002] found.

[7] Once the midlatitude surface cooling is established, cooling in the subtropics and tropics arises through the atmosphere-surface ocean interactions resembling that found by *Chiang and Bitz* [2005]. Increased northeasterly subtropical trades occur to the south of the 35–40°N ocean cooling region starting a little beyond a year after freshening, resulting in increased surface fluxes, primarily latent

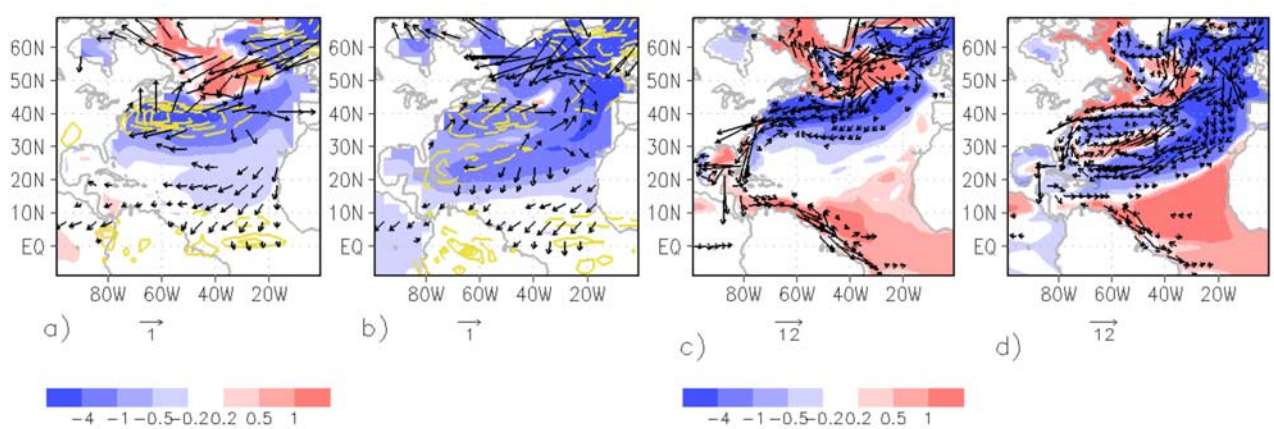

Figure 2. Climate anomalies averaged over (a, c) year 2 and (b, d) year 5. Figures 2a and 2b show anomalous SST (K), surface winds (m/s), and precipitation (mm/d). The SST and wind scales are similar to Figure 1, and precipitation is contoured in yellow at 0.5mm/d. Wind vectors are only plotted if the difference is significant at the 5% level, using a 2-sided t-test. Figures 2c and 2d show anomalous surface currents (cm/s) averaged between 0–400 m, and potential temperature (K) averaged between 100 m–400 m of the surface ocean. Current vectors are only plotted if larger than 2 cm/s, and the reference vector is 12 cm/s.

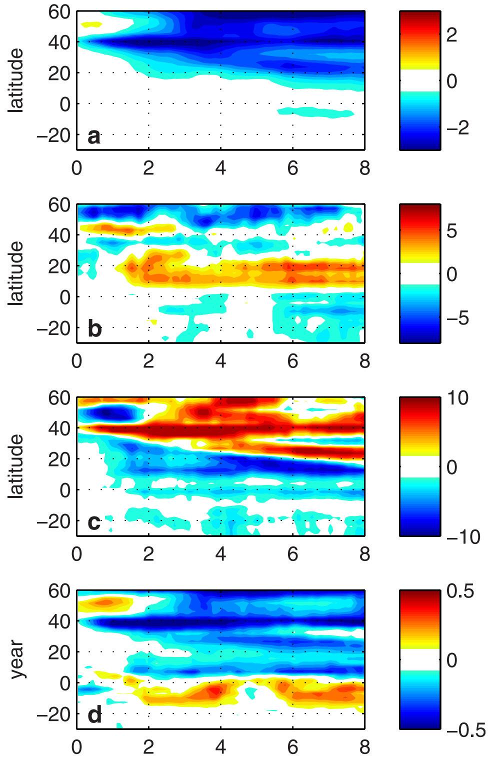

Figure 3. Temporal evolution of surface climate anomalies in the north and tropical Atlantic. Hovmoller plots of anomalies zonally averaged over the Atlantic basin of (a) SST (K); (b) surface kinetic energy (m^2/s^2) as a measure of wind speed; (c) net surface heat flux (W/m^2; positive is into the ocean); and (d) precipitation (mm/d). Quantities have been deseasonalized with a 12-month running mean. The time axis is years after freshening.

heat flux, out of the ocean, and cooling of SST just south of the original cooling latitudes (Figures 2a, 3b, and 3c). By the end of year 2, this ocean thermodynamic response to increased evaporation leads to a rapid but weak surface cooling in the entire north tropical Atlantic. An anomalous meridional SST gradient forms at the mean ITCZ latitudes that drive an anomalous cross-equatorial flow, shifting the ITCZ precipitation southwards by the beginning of year 3 (Figure 3d).

[8] A slower subtropical ocean dynamical adjustment occurs in years 4–8 that subsequently advects anomalously colder and fresher midlatitude waters into the subtropical ocean (Figure 2d). The adjustment starts as a result of the meridional surface height gradient introduced by the ocean water density change, and it is augmented by the anomalous wind changes. The anomalous ocean flow is oriented towards the southwest approximately parallel to the anomalous trades, and the anomalous trades intensify just to the south of this ocean-cooling region (Figure 2b), consistent with the onset of the WES mechanism. The combination of the anomalous ocean geostrophic flow and increased trades evolve together to bring a more intense cooling in the subtropical region that slowly migrates equatorwards (Figure 3a), eventually settling into a quasi-steady state around year 8. Increased low clouds form over the anomalously colder subtropical region (not shown), amplifying the cooling through albedo effects. This evolution ultimately yields the SST and flux patterns shown in Figure 1: a northeast-to-southwest band of relatively strong cold SST anomalies in the subtropics; a band of anomalous surface flux into the ocean co-incident with the north and central portion of the cold SST anomalies; and the anomalously stronger trades in the subtropical and tropical north Atlantic.

[9] In the northern tropics, the most notable effect of the ocean dynamics is its subsurface *warming* tendency, starting around year 2 (Figure 2c) and establishing itself by year 5. The subsurface warming is pronounced in northern tropics with typical values of 1–2 K extending to 500 m depth, while there is only a slight warming tendency in the southern tropics (not shown here; but see *Cheng et al.* [2007, Figure 3]). Two factors cause this subsurface heat accumulation: first, ocean wave adjustment deepens the tropical thermocline approximately symmetrically about the equator [*Huang et al.*, 2000] within a year of the freshwater pulse, and then over the next five years, the AMOC slowdown reduces the import of colder waters into the tropical Atlantic (20°S–20°N) from the south and reduces the export of warmer waters to the north, allowing anomalous convergence of ocean heat into this region. Our results are similar in nature to those of *Goodman* [2001] who examined ocean baroclinic adjustment to AMOC change in an ocean-only model. However, our more pronounced subsurface warming in the northern tropics requires other processes to be at work there. One is an anomalous downward Ekman pumping would be induced by the strengthened northeasterly trades; another is an interaction between the weakened AMOC and the northern subtropical cell (STC) circulation that allows more of the tropical water to be kept within the tropics (Chang et al., submitted manuscript, 2008).

[10] From a surface climate viewpoint, however, the most interesting feature in the northern tropics is that the North Tropical Atlantic SST *cools*, despite the subsurface warming. This anticorrelation between the surface and subsurface has been noted in 20th century multidecadal variations by *Zhang* [2007], who argued it to be a signature of AMOC variation. The cooling is confined to a thin (~30 m) layer at the surface, overlying a much thicker (~500 m) subsurface warming. We analyze this thin layer thermodynamics with an energy budget analysis averaged 7°–15°N over the Atlantic basin. The cooling is driven by the surface fluxes, though countered by the combined horizontal and vertical ocean heat transports (Figure 4a). The surface flux cooling is dominated by latent heat flux (Figure 4b). There are contributions also from longwave and sensible heat flux, but they are significantly smaller, and mostly cancelled out by the increased shortwave flux into the ocean. We estimated the contributions of the latent heat flux from wind-speed and air-sea humidity difference changes (by linearizing the bulk formula [*Saravanan and Chang*, 2000]), and found that wind speed can explain most of the latent flux increase. The damping effect by the ocean is primarily achieved by the *vertical* heat transport changes (inferred from the residual of the heat budget after subtracting out lateral

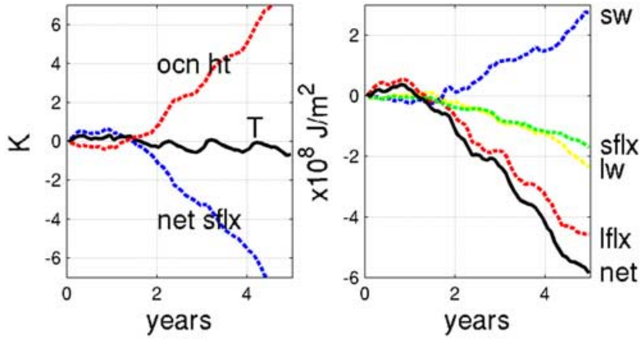

Figure 4. (a) Contributions to the ocean surface heat budget in the tropical North Atlantic 7°–15°N. Plotted are contributions, accumulated over time, of the net surface heat anomaly (blue) and inferred net ocean heat transport (red) to the 0–30 m mixed layer temperature anomaly (black). Units are K. (b) Components of the cumulative net surface flux anomaly shown in Figure 4a: shortwave (blue), longwave (yellow), sensible (green), and latent (red) heat flux. The cumulative net surface flux anomaly is shown for reference (black line). Units are in $\times 10^8$ J/m^2, and positive values are into of the ocean.

ocean heat transport), consistent with vertical advection and mixing from the anomalously warmer subsurface layer.

3. Discussion

[11] Our result argues for a significant role for the atmosphere - surface ocean mediated teleconnection in the extratropical influence to the tropical marine climate. It is nontrivial to cleanly identify teleconnection mechanisms because of the complexity of the climate adjustment; nevertheless, the influences of both baroclinic ocean adjustment and the atmospheric-surface ocean mechanisms can be discerned. It is in the north tropical Atlantic – the crucial region for perturbing ITCZ climate – where the influences of each are the clearest, and where they also come in opposition. The fact that the climate response to AMOC slowdown in the tropical Atlantic is a hemispheric asymmetric one testifies to the dominance of the atmospheric-surface ocean mechanism of *Chiang and Bitz* [2005] in setting the *surface* climate response. However, the presence of strong oceanic-driven damping might suggest that the teleconnection to the tropical Atlantic and ITCZ may well be tenuous, and perhaps differ from model to model.

[12] While we have not conclusively proven the atmospheric-surface ocean mechanism to be driven by WES, the climate changes we have thus identified in the simulations are all consistent with this mechanism; namely, the generation of stronger northeasterly trades equatorward of the cold SST (Figures 2a and 2b), and the strengthened trades leading to increased evaporation (Figures 3b and 3c) that dominates the cooling tendency in the north tropical Atlantic. There are other possible interpretations: in particular, an increase in atmospheric transient baroclinic wave activity from AMOC slowdown (as noted by *Cheng et al.* [2007]) would increase the northward export of angular momentum in the northern subtropics that in turn necessitates a stronger Hadley circulation (and hence stronger northeasterly trades) to balance the angular momentum budget there. *Bush* [2007] proposed a mechanism like this in the context of changes in the tropical Pacific to climate forcings during the Holocene and Last Glacial Maximum.

[13] With regards to the role of the ocean, it is interesting to note that ocean baroclinic adjustment has a more pronounced effect at the subtropical-subpolar gyre boundary, and less in the equatorial region as hypothesized in other studies (at least, in our simulations). However, Chang et al. (submitted manuscript, 2008) report a more pronounced warming to AMOC slowdown occurring in the equatorial and southern tropical Atlantic starting from the second decade after AMOC slowdown, a mechanism that involve interactions between the AMOC and STC. An abrupt transition in the ocean circulation occurs in the north tropical Atlantic provided the AMOC is sufficiently weakened, with the subsurface North Brazil Current reversing direction, carrying warm subtropical gyre water into the equatorial south Atlantic. This dramatically amplifies the oceanic contribution to warming over the equatorial and tropical south Atlantic. Taken together with our results, they give a more complete picture of the tropical Atlantic adjustment to AMOC slowdown.

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