Why is there a minimum in projected warming in the tropical North Atlantic Ocean?

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Received 8 April 2009; accepted 12 June 2009; published 17 July 2009.

[1] In IPCC projections for the 21st Century, the sea surface temperature (SST) shows several regions of minimum warming in the tropical oceans. These patterns appear both in fully coupled ocean-atmosphere general circulation models (GCMs) and also in atmospheric GCMs coupled to mixed-layer ocean models, and are robust across a multi-model ensemble. The present study focuses on the minimum warming in the tropical North Atlantic, as it has implications for the influence of greenhouse gas-induced climate change on hurricane development. The surface heat budget is analyzed in order to determine the causes for this minimum warming. It is found that the primary contribution is through the influence of the climatological mean wind speed on the efficiency of latent heat flux. In regions of high wind speed, radiative heating can be balanced by latent heat flux with a smaller change in SST than in other regions of the tropics. **Citation:** Leloup, J., and A. Clement (2009), Why is there a minimum in projected warming in the tropical North Atlantic Ocean?, *Geophys. Res. Lett.*, *36*, L14802, doi:10.1029/2009GL038609.

1. Introduction

[2] In IPCC projections for the 21st Century, the Sea Surface Temperature (SST) warms everywhere in the tropical ocean, but not uniformly [*Meehl et al.*, 2007b; *Vecchi and Soden*, 2007]. In particular, the multi-model mean shows several regions of minimum warming including the South-Eastern tropical Pacific, the South tropical Atlantic, and the North tropical Atlantic.

[3] Using coupled simulations from the SRES-A1B scenario, *Vecchi and Soden* [2007] show a local minimum warming in the Northern Atlantic, covering a broad area, extending from the Caribbean Sea to the northwest coasts of Africa. This region is of great interest as it is the "main development region" of tropical cyclone in the Atlantic [*Mann and Emanuel*, 2006]. *Vecchi and Soden* [2007] find that the minimum warming in this region relative to the rest of the tropics results in an increase in the static stability of the troposphere which would (among other factors) limit development of intense tropical cyclones in that region.

[4] This feature appears in fully coupled ocean-atmosphere general circulation models (GCMs) (Figure 1b), but also exists in atmospheric GCMs coupled to mixed-layer ocean models (Slab models, Figure 1a). In both the coupled and Slab models, the warming in the northern tropical Atlantic is about 0.5 degree less than the tropical mean warming consistent with the values used by *Vecchi and Soden* [2007]. While this may appear to be a subtle difference, numerous papers have shown that this difference can have an impact on hurricane development [*Emanuel et al.*, 2008; *Santer et al.*, 2006; *Vecchi and Soden*, 2007]. In addition, the minimum warming is robust across each multi-model ensemble, as shown in Figures 1c and 1d. All of the 7 Slab models and 20 coupled models used here simulate a warming in the tropical North Atlantic that is less than the tropical mean.

[5] The presence of the minimum warming in the Slab models, where ocean heat transports are fixed (no advection nor mixing), suggests that the main processes driving this response mainly come from atmospheric mechanisms. As such, we analyze the Slab model to provide the most simple framework for interpreting the pattern of SST change. The change in net surface heat flux in the region of minimum warming in the coupled models (which in equilibrium will balance the net effect of ocean heat transports) is about 1 W m^{-2}. Hence, we conclude that ocean heat transports are not important in determining the SST change in the region of minimum warming, and that the processes are therefore similar in the Slab and coupled models. In section 2 we describe the model simulations that are used, and then in Section 3, the minimum warming in the (North) tropical Atlantic is interpreted in the context of the equilibrium surface heat budget.

2. Model Simulations

[6] All the model outputs used in this study are from the World Climate Research Programme's (WCRP's) Coupled Model Intercomparison Project phase 3 (CMIP3) multi-model data set [*Meehl et al.*, 2007a].

[7] The SST difference ΔT in Figure 1a comes from seven models available for the Slab experiments (a *control* run and a $2 \times CO_2$ perturbation run). This is the ensemble mean difference for all those seven models: $\Delta T = T_{2\times CO_2} - T_{control}$. For the analysis of the heat budget, we are restricted to those seven models which provide all of the necessary diagnostics for at least 20 years of monthly data for the Slab experiments. Daily data were used from the same experiments to compute the surface wind speed.

[8] All data sets were interpolated onto a common grid (2.5×2.5) to allow intercomparison and ensemble mean. We consider annual mean as we did not find any seasonality in the minimum warming (not shown). And given the robustness across the ensemble of models, we will show ensemble mean for the calculation made in the following.

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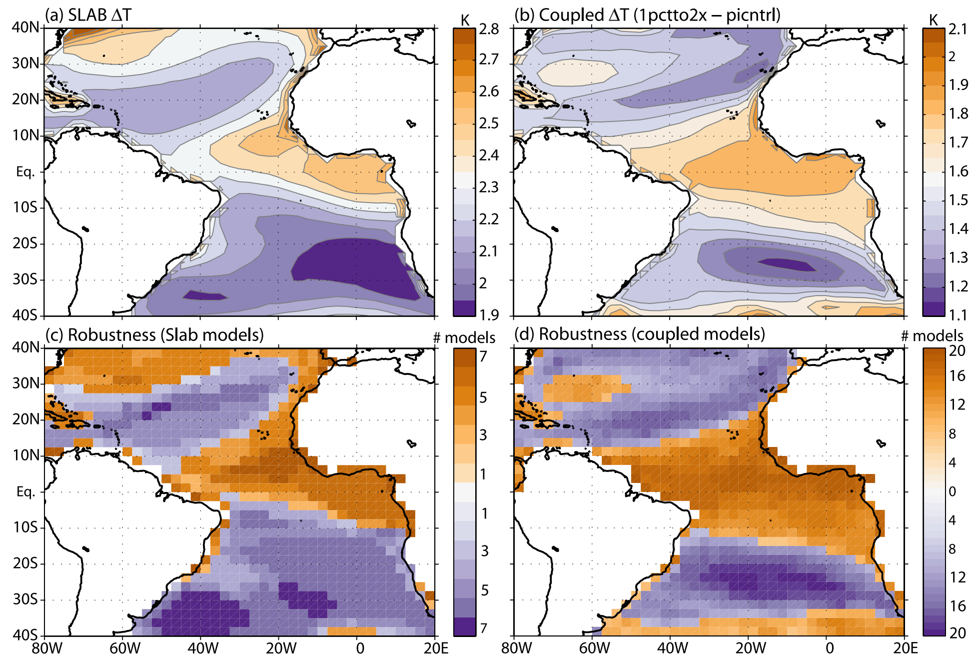

Figure 1. (a) Ensemble mean SST difference for the Slab experiment $\Delta T = T_{2\times CO_2} - T_{control}$. (b) Ensemble mean SST difference for twenty coupled models available for both the "pre-industrial" (picntrl) and the "1% to CO_2" (1pctto2x) scenarios: $\Delta T = T_{1pctto2x} - T_{picntrl}$. (c) Robustness of SST warming among the ensemble of Slab models: The sign indicates whether a region warms by more or less than the tropical mean. The color indicates the number of models that have the same sign as the ensemble mean. (d) Same as Figure 1c for the ensemble of Coupled models.

[9] The models used were: CSIRO-MK3.0, GFDL-CM2.0, INMCM3.0, MIROC3.2-HR, MIROC3.2-MR, MPI-ECHAM5, MRI-CGCM2.3.2a.

3. Analysis of the Surface Heat Budget

[10] The surface heat budget is analyzed in order to provide an explanation for the minimum warming in the Northern tropical Atlantic, defined as the region where the warming is nominally lower than 2.2K for the Slab models in Figure 1a (this choice of region is arbitrary, but will serve to guide the reader in the discussion of the results). The equilibrium difference in surface heat budget for the Slab models (in which ocean heat transport does not change) between the *control* and 2 × CO_2 experiments can be written as follows:

$$\underbrace{\Delta Q_{SW} + \Delta Q_{LW} + \Delta Q_{SH}}_{heating} = \underbrace{\Delta Q_{LH}}_{cooling}.$$

where ΔQ_{SW}, ΔQ_{LW}, ΔQ_{SH}, and ΔQ_{LH} are respectively the changes in shortwave, longwave, sensible heat, and latent heat fluxes. Both the longwave and sensible heat fluxes changes warm the ocean throughout the tropics in all models. The shortwave changes in the models are not robust, and are grouped with the "heating" terms because the ocean does not have direct control over this. The latent heat flux, on the other hand, is directly tied to the ocean surface temperature, and is the primary means by which the ocean can balance the heating.

[11] It is worth noting that although one cannot infer direct causality in equilibrium SST change, numerous previous studies have wrestled with this challenge, and have developed tool for interpreting equilibrium SST change [*Hartmann and Michelsen*, 1993; *Knutson and Manabe*, 1995; *Clement and Seager*, 1999; *Seager and Murtugudde*, 1997; *Seager et al.*, 2000]. Here we follow some of the same ideas and develop a simple framework for interpreting the pattern of SST change.

[12] As shown on Figure 2, the structure of ΔQ_{LH} (or equivalently the heating of the ocean) is not sufficient to explain the warming structure in ΔT, as the minimum warming zones do not coincide with minima in the latent heat flux (or equivalently a minimum in the heating of the ocean). The pattern correlation between these two (ΔT and ΔQ_{LH}) is 0.2, so that the spatial structure in latent heat flux, if linearly related to the SST (which it is not) would only explain 4% of the spatial variance of SST. Instead, we hypothesize that the structure in ΔT comes from the spatial structure of the dependence of ΔQ_{LH} on T [*Seager and*

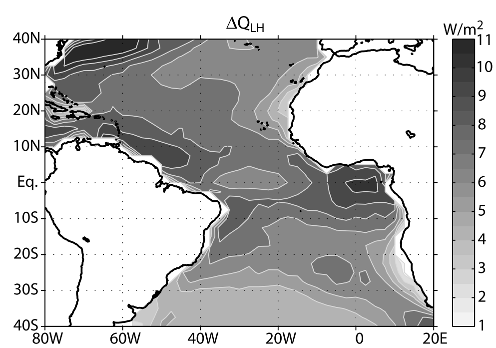

Figure 2. Latent heat flux change ΔQ_{LH} between the 2 × CO_2 and *control* runs (ensemble mean).

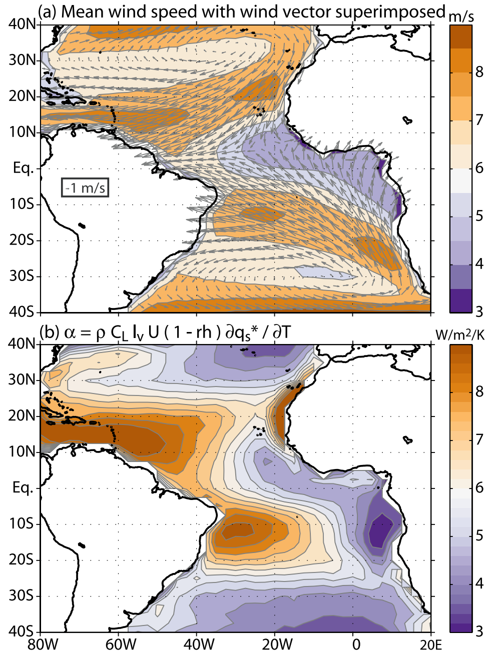

Figure 3. (a) Mean wind speed (color) with wind vectors on top (arrows) for the *control* run and (b) α as calculated from equation (3).

Murtugudde, 1997; *Hartmann and Michelsen*, 1993; *Liu et al.*, 2005]. To test this hypothesis, first let us consider the Bulk formula for latent heat flux:

$$Q_{LH} = \rho C_L l_v U(1 - rh)q_s^*, \tag{1}$$

where $\rho = 1.3\ kg/m^3$ is the density of air, $C_L = 1.35 \times 10^{-3}$ is the latent heat transfer coefficient, $l_v = 2.5 \times 10^6\ J/kg$ is the coefficient of latent heat of evaporation, U in m/s is the mean wind speed, rh is the relative humidity, and q_s^* is the saturation specific humidity.

[13] Because the latent heat flux depends on three different variables, U, rh, and T (through the saturation specific humidity), we can linearize the change in latent heat flux as:

$$\Delta Q_{LH} = \underbrace{\frac{\partial Q_{LH}}{\partial T}\Delta T}_{\alpha\Delta T} + \frac{\partial Q_{LH}}{\partial U}\Delta U + \frac{\partial Q_{LH}}{\partial rh}\Delta rh, \tag{2}$$

and α is:

$$\alpha = \frac{\partial Q_{LH}}{\partial T} = \rho C_L l_v U(1 - rh)\frac{\partial q_s^*}{\partial T}. \tag{3}$$

[14] This sensitivity, α, depends on the mean wind speed U, on the mean relative humidity rh and on the mean SST via $\partial q_s^*/\partial T$, however, it is strongly influenced by the spatial structure in wind speed (Figure 3). The sensitivity of latent heat flux to temperature is maximum in regions of high wind speed (i.e. in the trade winds), and minimum in regions of low wind speed (around equator for example). Thus, by setting the other factors in α to their tropical mean values (rh and $\partial q_s^*/\partial T$), we can simplify to $\alpha_{wind} = \rho C_L l_v U(1 - \langle rh\rangle) \langle\partial q_s^*/\partial T\rangle$ (where <.> denotes the tropical mean).

[15] Wind speed changes in the region of minimum warming are small (about 0.1 m s^{-1}), as are changes in rh ($\sim$1%). The coefficients $\partial Q_{LH}/\partial U$ and $\partial Q_{LH}/\partial rh$ are respectively order 10 and 100 [*Hartmann and Michelsen*, 1993], and therefore the two last terms in equation (2) are small compared to the first term in the region of minimum warming. Thus, by including only the first term on the right hand side of equation (2), we can calculate a change in temperature, ΔT^*, as follows:

$$\Delta T^* = \frac{1}{\alpha}\Delta Q_{LH}. \tag{4}$$

[16] The structure in ΔT^* depends both on the structure of α and ΔQ_{LH}. To test the influence of the mean wind speed on ΔT^*, we first consider a uniform heating (the average of ΔQ_{LH} over the tropics) and also α_{wind} (in which the spatial structure in rh and $\partial q_s^*/\partial T$ are not included). The result is shown in Figure 4a. A clear minimum in the northern tropical Atlantic emerges, and the spatial correlation, used as a measure of agreement between the predicted ΔT^* and the actual ΔT (Figure 1a), is close to 0.6 (Table 1). We note that the minimum warming in the South Atlantic can not be explained in this simple framework. Some of this is the result of the spatial structure in ΔQ_{LH} which shows a minimum in that region (Figures 2 and 4b), and there are also larger wind speed changes in that region.

[17] Other estimates of ΔT were calculated by solving equation (4) and allow for spatial structure in the heating as

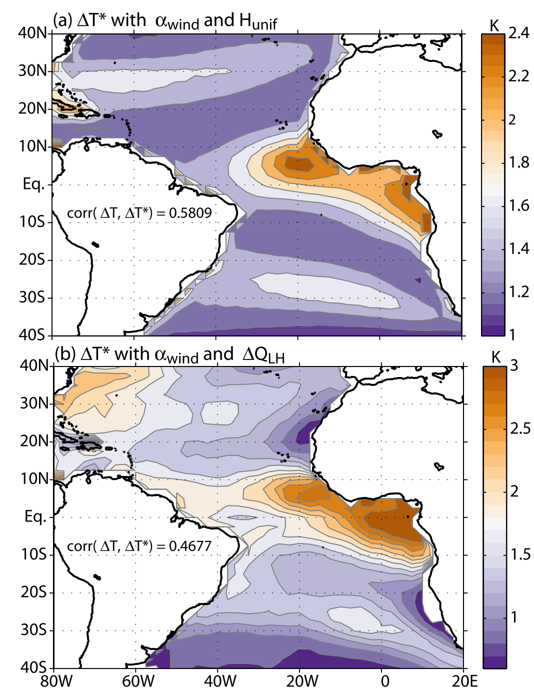

Figure 4. ΔT calculated from equation (4). (a) The case with a uniform heating H_{unif} and α_{wind} (where α is computed by only including the spatial structure in mean wind speed and rh and $\partial q_s^*/\partial T$ are set to their climatological values in equation (3)). (b) The same as Figure 4a, but using the "real" forcing ΔQ_{LH} (Figure 2).

Table 1. Correlation Between the Real ΔT and Different Estimates of ΔT Depending on the Forcing and Simplifications Made to α[a]

Forcing	α	α_{wind}	α_{rh}	α_{sst}
ΔQ_{LH}	0.3641	0.4677	0.1353	0.1477
H_{unif}	0.2575	0.5809	−0.2454	−0.3128

[a]Real ΔT is from Figure 1a and estimates of ΔT are from equation (4). Forcings are real ΔQ_{LH} or uniform heating. α_{wind} is the case where the *rh* and $\partial q_s^*/\partial T$ are set to their climatological mean values and only the mean wind speed has spatial structure. Similarly, α_{rh} only includes spatial structure in *rh*, and U and $\partial q_s^*/\partial T$ are set to their tropical mean values; α_{sst} only includes spatial structure in $\partial q_s^*/\partial T$.

well as the spatial structure in *rh* and $\partial q_s^*/\partial T$ to influence α (equation (3)). The spatial correlations for these estimates of ΔT with the "real" warming are listed in Table 1. The highest correlation is for the case where only the structure in climatological wind is used. The spatial structure in *rh* and SST do not contribute in the explanation of the warming.

[18] We also note that, while the spatial structure of the change in surface temperature does appear to be explained in large part by the mean wind speed (as measured by the correlation), the absolute value and range of ΔT are not. The physics of the response to CO_2 forcing have been highly simplified here. For instance, we have not included the contributions of the changes in wind speed and relative humidity (as in equation (2)), and the simulation of latent heat flux in climate models is not as simple as the bulk formula (where we have assumed, for example, a constant drag coefficient). For this reason we do not expect the absolute temperature changes to be explained in the context of this simple framework.

4. Summary and Discussion

[19] In this short note, we have investigated the explanations for the robust minimum warming in the North Atlantic Ocean simulated by climate models. By developing a simple framework to analyze the surface heat budget, it is shown here that the primary contribution to the minimum warming in the North tropical Atlantic is through the influence of the climatological mean wind speed on the efficiency of latent heat flux: It is easy to cool off (i.e. increase latent heat flux) windy regions (like in the trade wind regions), resulting in a smaller SST change [*Seager and Murtugudde*, 1997], and it is hard to cool off (i.e. decrease latent heat flux) regions of low wind speed (like the equator), resulting in larger SST change. This simple explanation is consistent with the robustness of pattern in not only Slab models but coupled models as well. A similar argument was made by *Liu et al.* [2005] to explain the "enhanced equatorial warming." In addition a lower efficiency of surface latent heat flux in the equatorial region (because of low wind speeds), those authors also invoke ocean mixing and surface solar radiation to explain the meridional structure in warming. Our analysis suggests, however, that the geographical structure in efficiency of latent heat flux (due to climatological winds) alone is sufficient to explain much of the structure in warming in the Atlantic. Furthermore, our explanation is plausible because the simulated and the "observed" wind speed have similar spatial structure (using NCEP/NCAR reanalysis daily data, not shown).

[20] This analysis will not apply in regions where the wind speed change is significant. For example, robust increases in wind speed are simulated in the south eastern Atlantic and have also been noted in the southeastern Pacific [*Falvey and Garreaud*, 2009]. In those regions, there is a strong increase in the wind speeds that appears in all models. The changes in winds in those regions also affect the ocean circulation, so that the coupled response also contributes to the minimum warming there. However, in the tropical Atlantic, where wind speed changes are small, this robust pattern in minimum warming, which has implications for future changes in hurricane development, appears to have a simple explanation.

[21] **Acknowledgments.** We acknowledge the modeling groups, the Program for Climate Model Diagnosis and Intercomparison (PCMDI) and the WCRP's Working Group on Coupled Modelling (WGCM), for their roles in making available the WCRP CMIP3 multi-model data set. Support of this data set is provided by the Office of Science, U.S. Department of Energy. This work was funded by a NOAA Office of Global Progams grant (NA06OAR4310142).

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