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Key Points:

- Low cloud over the tropical North Atlantic is highly sensitive to SST
- Trade wind speed response to midlatitude SST anomaly is critical to understand tropical arm of AMO
- Models do not simulate the cloud and dust feedback and miss the tropical arm of AMO

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Positive low cloud and dust feedbacks amplify tropical North Atlantic Multidecadal Oscillation

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Abstract The Atlantic Multidecadal Oscillation (AMO) is characterized by a horseshoe pattern of sea surface temperature (SST) anomalies and has a wide range of climatic impacts. While the tropical arm of AMO is responsible for many of these impacts, it is either too weak or completely absent in many climate model simulations. Here we show, using both observational and model evidence, that the radiative effect of positive low cloud and dust feedbacks is strong enough to generate the tropical arm of AMO, with the low cloud feedback more dominant. The feedbacks can be understood in a consistent dynamical framework: weakened tropical trade wind speed in response to a warm middle latitude SST anomaly reduces dust loading and low cloud fraction over the tropical Atlantic, which warms the tropical North Atlantic SST. Together they contribute to the appearance of the tropical arm of AMO. Most current climate models miss both the critical wind speed response and two positive feedbacks though realistic simulations of them may be essential for many climatic studies related to the AMO.

1. Introduction

The North Atlantic sea surface temperature (SST) undergoes coherent multidecadal oscillations with a period of about 60–80 years, known as the Atlantic Multidecadal Oscillation (AMO) (also referred to as the Atlantic Multidecadal Variability) [*Kerr*, 2000]. The AMO index can be computed as the 10 year running mean of linearly detrended North Atlantic SST anomalies [*Enfield et al.*, 2001] (see SOM and Figure S1 in the supporting information) and the SST expression of a warm phase of AMO is shown in Figure 1. It is characterized by a horseshoe pattern with two arms: one south of Greenland and the other in the tropical North Atlantic. The tropical arm of North Atlantic SST anomalies is responsible for many climatic impacts on neighboring and remote regions through teleconnection including Atlantic hurricanes, North American, and North African climate [*Knight et al.*, 2006; *Zhang and Delworth*, 2006; *Sutton and Hodson*, 2007]. The low frequency nature of AMO is an important source of predictability for decadal climate predictions using coupled models [*Meehl et al.*, 2014]. Improved understanding of the AMO helps to better attribute signals of anthropogenic climate change [*Bindoff et al.*, 2013]. Understanding the processes that give rise to the AMO is thus of great scientific and societal interest.

However, the tropical arm of North Atlantic SST anomalies is often significantly underestimated or entirely absent (Figure 1b) in the models of the Fifth Coupled Model Intercomparison Project (CMIP5) despite its importance and an often correctly simulated midlatitude arm [*Delworth et al.*, 2007; *Kavvada et al.*, 2013; *Ba et al.*, 2014; *Martin et al.*, 2014]. Observational analyses suggest that the focus of North Atlantic SST anomalies is in the midlatitudes during the mature stage of AMO and SST anomalies subsequently propagate into the tropics [*Guan and Nigam*, 2009; *Kavvada et al.*, 2013; *Hodson et al.*, 2014]. Recent studies suggest that ocean-atmosphere coupled processes are needed to propagate SST anomalies from midlatitudes to south of 34°N [*Zhang and Zhang*, 2015]. Regardless of the trigger mechanism in the midlatitudes, the inability to propagate the midlatitude anomaly into the tropics is a common shortcoming in climate models, which may indicate shortcomings in the representation of crucial model processes. Anthropogenic aerosol radiative forcing has been proposed as a driver for the tropical arm of SST anomalies [*Mann and Emanuel*, 2006; *Delworth et al.*, 2007; *Booth et al.*, 2012] and analyses of model simulations seem to support the role of

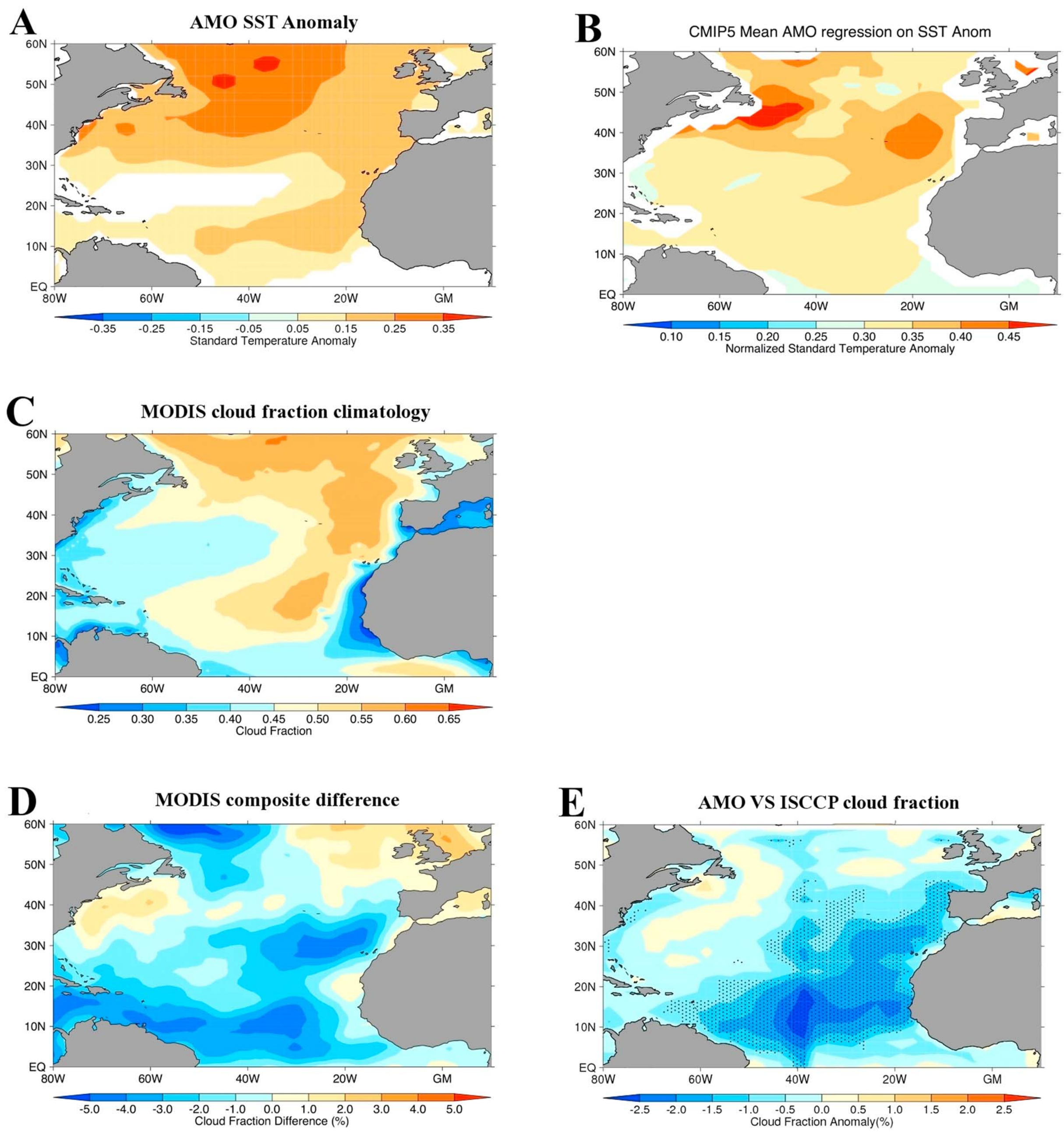

Figure 1. Spatial patterns of the Atlantic Multidecadal Oscillation from (a) observation and (b) model mean, (c) MODIS low cloud fraction climatology, (d) composite difference in MODIS low cloud fraction between high and low AMO index years, and regression of ISCCP low cloud fraction against AMO index. The AMO index is defined as the detrended areal weighted North Atlantic SST anomaly averaged over the region [0 ~ 60 N, 80 W ~ 10 E]. The spatial pattern (a) of AMO is defined as the regression of the North Atlantic SST against the AMO index scaled by one standard deviation of the AMO index. For models (b), each spatial pattern is first normalized before averaging. MODIS low cloud fraction climatology (c) is an average of Terra and Aqua with low clouds defined as those with cloud top pressure greater than 600 hPa. MODIS composite difference (d) is between the average of top (2003, 2004, and 2008) and bottom (2007, 2009, and 2011) AMO years during MODIS era. Stippled areas in Figure 1e indicate that correlation is statistically significant at 95% level; the spatial pattern is calculated in the same way as Figure 1a except the AMO index time series during the ISCCP period is normalized before regression.

aerosols and clouds [*Martin et al.*, 2014], but considerable doubts remain regarding its magnitude and its consistency with observed ocean heat content, spatial pattern of SST, and surface salinity [*Zhang et al.*, 2013].

Here we propose that two positive feedback mechanisms, one involving low clouds and one involving dust, may be responsible for amplifying the tropical arm of the AMO. We assess their representation in current climate models by examining the recent AMO phase switch within the satellite era. Since early 1980s, tropical

North Atlantic SST has increased at a fast rate of nearly 0.2 C/decade as AMO switched from a negative to a positive phase. Because ocean variability is of much lower frequency than atmospheric variability, we assume that on multidecadal time scales, the ocean is the slow driver and fast atmospheric processes are responses/feedbacks. Our investigation uses satellite data as well as other observational and modeling data (see SOM for details).

Figure 1c shows the climatology of maritime low cloud fraction from two Moderate Resolution Imaging Spectrometer (MODIS) instruments (2002–2013) on both the Aqua and Terra satellites [*Platnick et al*., 2003]. The spatial pattern bears close resemblance to the pattern of the SST anomaly associated with AMO (Figure 1a): a maximum across the midlatitude North Atlantic from east of the Canadian coast to west of the Iberian Peninsula and a secondary maximum in the tropics west of the African continent. Composite analyses for MODIS and International Satellite Cloud Climatology Project (ISCCP, 1982–2009) periods suggest that in both data sets low cloud fraction generally decreases with increasing AMO index (Figures 1c, S2, and S3). The decrease is pronounced in the tropical North Atlantic downwind of West African continent, namely, the vicinity of the tropical arms of AMO and low cloud fraction. During the MODIS period, annual mean low cloud fraction in this area decreases by 2–4% on average for the 3 years with the highest AMO indices compared to the 3 years with the lowest (Figure 1d). For the longer ISCCP record, the tropical North Atlantic low cloud fraction is smaller by 3 to 8% in the 2000s (positive phase of AMO) than in the 1980s (negative phase of AMO) depending on location (Figure S2). The time series of annual low cloud fraction anomaly averaged over the tropical arm region is strongly anticorrelated with annual AMO index during the ISCCP period ($r = -0.68$; Figures 1e and 2a). When AMO switches from a negative phase to a positive phase in the mid-1990s, the low cloud fraction anomaly transitions into a negative regime around the same time (Figure 2a).

A regression of low cloud fraction onto AMO index measures the rate of low cloud fraction decrease with positive AMO index. Rates from both cloud data sets are about 2% of low cloud fraction anomaly per one standard deviation of normalized AMO index during MODIS and ISCCP periods, or about $-10\%\,K^{-1}$. Regressions of MODIS and ISCCP cloud fraction against AMO index show quite similar patterns with a pronounced peak in the tropical North Atlantic (Figures 1 and S3). The similarity is remarkable because sampling, sensor characteristics, and algorithms used for MODIS and ISCCP are all different, and the phase and variability of AMO are also quite different for the time periods covered by the two data sets. The consistency and sensitivity thus indicate a robust and strong low cloud response to underlying tropical North Atlantic SST.

The sensitivity of low cloud fraction to SST in this region is consistent with our physical understanding. The climatological SST over the tropical North Atlantic, where cloud fraction sensitivity peaks, is warmer than typical subtropical stratocumulus dominated regions and cooler than tropical trade cumulus dominated regions [*Teixeira et al*., 2011]. In this transitional cloud regime, an increase in SST tends to decrease the inversion strength, increase entrainment mixing, and reduce cloud fraction [*Bretherton et al*., 2013; *Qu et al*., 2014]. Physically consistent changes in other meteorological variables such as sea level pressure and subsidence are also associated with the shift in SST and low cloud fraction [*Clement et al*., 2009]. For example, the increasing tropical North Atlantic SST in this region from the 1980s to 2000s was accompanied by lower sea level pressure and weaker trade winds, which are consistent with reduced low cloud fraction (Figure 2a).

The low cloud fraction reduction associated with a positive AMO exerts significant radiative impact on the tropical North Atlantic SST because low clouds have a high net surface cooling efficiency of roughly $1\,Wm^{-2}$ per percent change of cloud fraction [*Klein and Hartmann*, 1993]. Using this efficiency, we estimate that the low cloud fraction reduction in the 3 years with the highest AMO index during the MODIS period increases the net surface radiative heating by 2–4 Wm^{-2} compared to the 3 years of the lowest AMO index. Similarly, an increase of surface radiative flux by 3–8 Wm^{-2} in the tropical North Atlantic is expected due the reduction in low cloud fraction indicated by ISCCP since the 1980s. Such estimates are consistent with concurrent satellite-based radiative flux estimates during the MODIS era (Figures S5 and S6). Considering only the change in the surface energy budget due to low clouds [*Clement et al*., 2009], the anomalous cloud radiative effect is sufficient to alter SST by a few tenths of a degree, enough to account for a large portion of AMO related SST variation [*Sutton and Hodson*, 2005; *Booth et al*., 2012]. Taken together, these results illustrate a potentially strong positive feedback between low clouds and SST on multidecadal time scales in the tropical North Atlantic: More solar radiation is absorbed by the sea surface due to decreased cloud fraction, which helps to maintain and amplify warm SST anomalies associated with the AMO and vice versa.

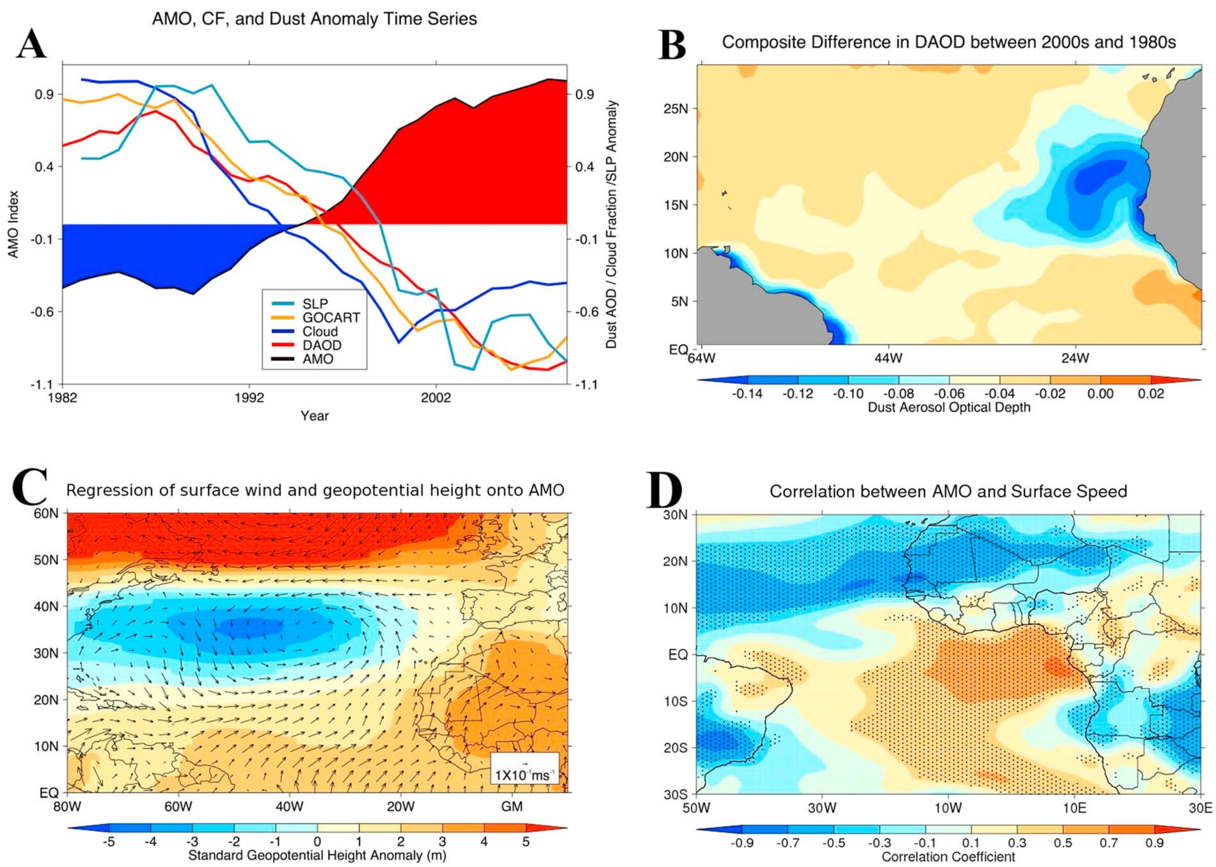

Figure 2. (a) Normalized time series of AMO index, dust aerosol optical depth (AOD), ISCCP low cloud fraction, Modern Era Retrospective analysis for Research and Applications (MERRA) sea level pressure, and GOCART modeled dust AOD from 1982 to 2009. All variables except AMO index are averaged over the tropical Atlantic (0°–20°N, 60°W ~ 20°W). Normalization is achieved by dividing each time series by its maximum absolute value. (b) Dust AOD composite difference between the 2000s and the 1980s. (c) Same as Figure 1a, but for anomalies of MERRA surface wind vector and 700 mb geopotential height (shaded). (d) Correlation coefficient between AMO index and MERRA near surface (10 m) wind speed. North of about 10°N, trade wind is weaker with a positive AMO phase and the opposite is true for the area to the south, which agrees well with the theory [*Chiang and Friedman*, 2012]. Stippled areas indicate places where correlation is statistically significant at the 95% level.

Aerosols are another factor that affects SST. Dust from North Africa is the dominant aerosol type over the tropical North Atlantic and it modulates the surface energy balance and hence SST [*Evan et al.*, 2009]. As shown in Figure 2a, dust loading over the tropical North Atlantic, both observed by satellites [*Evan and Mukhopadhyay*, 2010] and modeled by the Goddard Chemistry Aerosol Radiation Transport (GOCART) [*Chin et al.*, 2014], has decreased from 1982 to 2009. The decrease in dust loading is highly correlated with the increase in AMO index and matches the timing of regime change during the mid-1990s rather well. Both correlations are statistically significant ($p < 0.01$). Longer dust records, extending back to 1950s and 1960s, also support this relationship (Figure S4). The decrease of dust optical depth since early 1980s (Figure 2b) has been estimated to have increased the downwelling surface radiative flux by up to 1–2Wm^{-2}, thereby potentially increasing the tropical North Atlantic SST by up to a few tenths of a degree [*Foltz and McPhaden*, 2008; *Evan et al.*, 2009; *Mahowald et al.*, 2010]. Thus, the negative correlation between the tropical North Atlantic dust loading and North Atlantic SST [*Wang et al.*, 2012; *Chin et al.*, 2014; *Ridley et al.*, 2014] suggests an additional positive feedback: smaller dust loading during a positive AMO phase amplifies the tropical North Atlantic SST warming.

What are the coupled processes that lead to the observed positive feedbacks associated with low clouds and dust? A credible framework emerges from our understanding of how extratropical forcing affects tropical climate [see *Chiang and Friedman*, 2012, and references therein]. A positive SST anomaly in the extratropics, e.g., the midlatitude arm of AMO, generates an anomalous local cyclonic circulation in the atmosphere [*Kushnir and Held*, 1996]. Through ocean advection, water vapor-evaporation- SST (WES) feedback [*Xie and Philander*, 1994] and weakening of ocean upwelling, warm SST anomalies expand to the southern and southeastern edge of the cyclone and extend to lower latitudes. Such anomalous circulation reduces the speed of

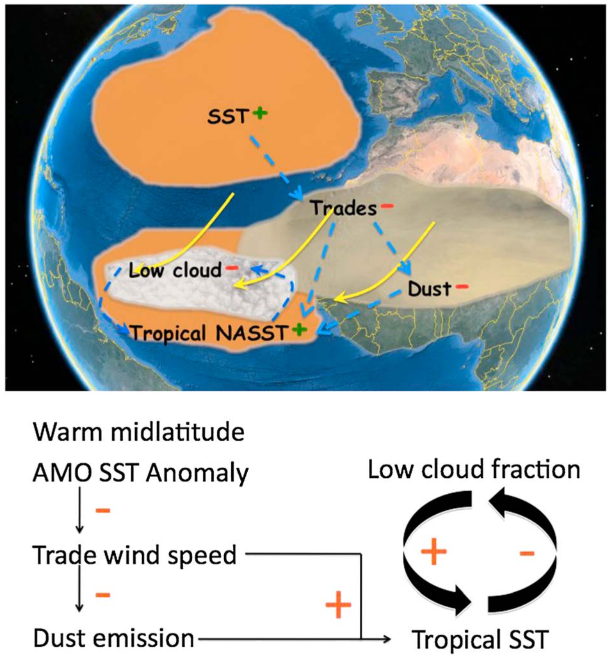

Figure 3. (top) A schematic and (bottom) flow diagram illustrating the proposed feedbacks and related processes. Warm SST anomaly in the North Atlantic introduces a cyclonic anomaly, which weakens trade winds. Weaker trade winds reduce dust emission over the Sahara and suppress dust transport toward the tropical Atlantic. They also contribute to reduce low cloud fraction over the tropical North Atlantic. Both responses yield warmer tropical North Atlantic SST.

northeasterly trade winds in the tropics, which decreases dust emission and dust loading over the tropical North Atlantic (Figure 3); at the same time, positive SST anomalies amplify through decreasing low cloud fraction over the Tropical North Atlantic. The critical link in this chain of events involves the trade wind speed and its response to the middle latitude SST anomaly because trade wind speed can affect both tropical SST and associated low clouds, and modulate dust emission and transport (Figure 3).

A regression of surface wind and 700 mb geopotential height on AMO index between 1982 and 2009 manifests this conceptual picture. It clearly shows that in association with a positive AMO phase, an anomalous low-pressure system develops in the middle of North Atlantic. The anomalous southwesterlies to the south of the low-pressure system act to weaken the tropical northeasterly trade winds both over ocean and over the dust source regions (Figures 2c and 2d). Indeed, surface wind speed is negatively correlated with AMO on multidecadal scales over these regions (Figure 2d). Decreased wind speeds suppress dust emission in source regions and reduce transport to the tropical ocean because of the anomalous onshore flow (Figure 2c) [*McGee et al*., 2010; *Chin et al*., 2014; *Ridley et al*., 2014], which then reduces overall dust radiative cooling of the sea surface, closing a positive feedback loop. Reduced trade wind speed weakens ocean upwelling and reduces evaporation, which increases tropical SST that reduces low cloud fraction and therefore closes the positive low cloud feedback loop [*Qu et al*., 2014] (Figure 3).

How well do the latest CMIP5 models reproduce these feedback processes? For cloud feedback evaluation we considered the low cloud response separately averaged over both the midlatitude arm and the tropical arm of the AMO (Figure 4a). Figure 4a shows that on average model simulations driven by observed SST fields and other observed boundary conditions (i.e., Atmospheric Model Intercomparison Project (AMIP) style experiments) show an increase in low cloud fraction in the tropical North Atlantic, opposite to observations, while correctly simulate a decrease in the midlatitudes. Indeed, only a couple of models qualitatively reproduce multidecadal low cloud temporal variability (Figures 4b and S10) in the tropics, while low cloud changes from most models are actually of opposite sign. One of the bright exceptions is CESM-CAM5, which has an updated low cloud scheme [*Bretherton and Park*, 2009], and performs the best in this regard.

Regarding the dust feedback, most model runs do not qualitatively reproduce the sign of dust change. Even when they do, the magnitude of absolute change is substantially underestimated, which is a common issue [*Evan et al*., 2014]. On the contrary, off-line chemical transport models [*Chin et al*., 2014; *Ridley et al*., 2014] forced by reanalysis data can accurately reproduce the dust change. This perhaps indicates that shortcomings are present in the modeled dynamical response to SST in CMIP5 models. Indeed, our examination of results from AMIP style runs shows no clear indication of improved performance either in terms of dust response or surface wind speed response (Figures 4c, S8, and S9).

It is therefore reasonable to suggest that an important cause of the models' shortcoming is rooted in the simulation of tropical trade wind speed response to SST anomalies because the wind speed response is

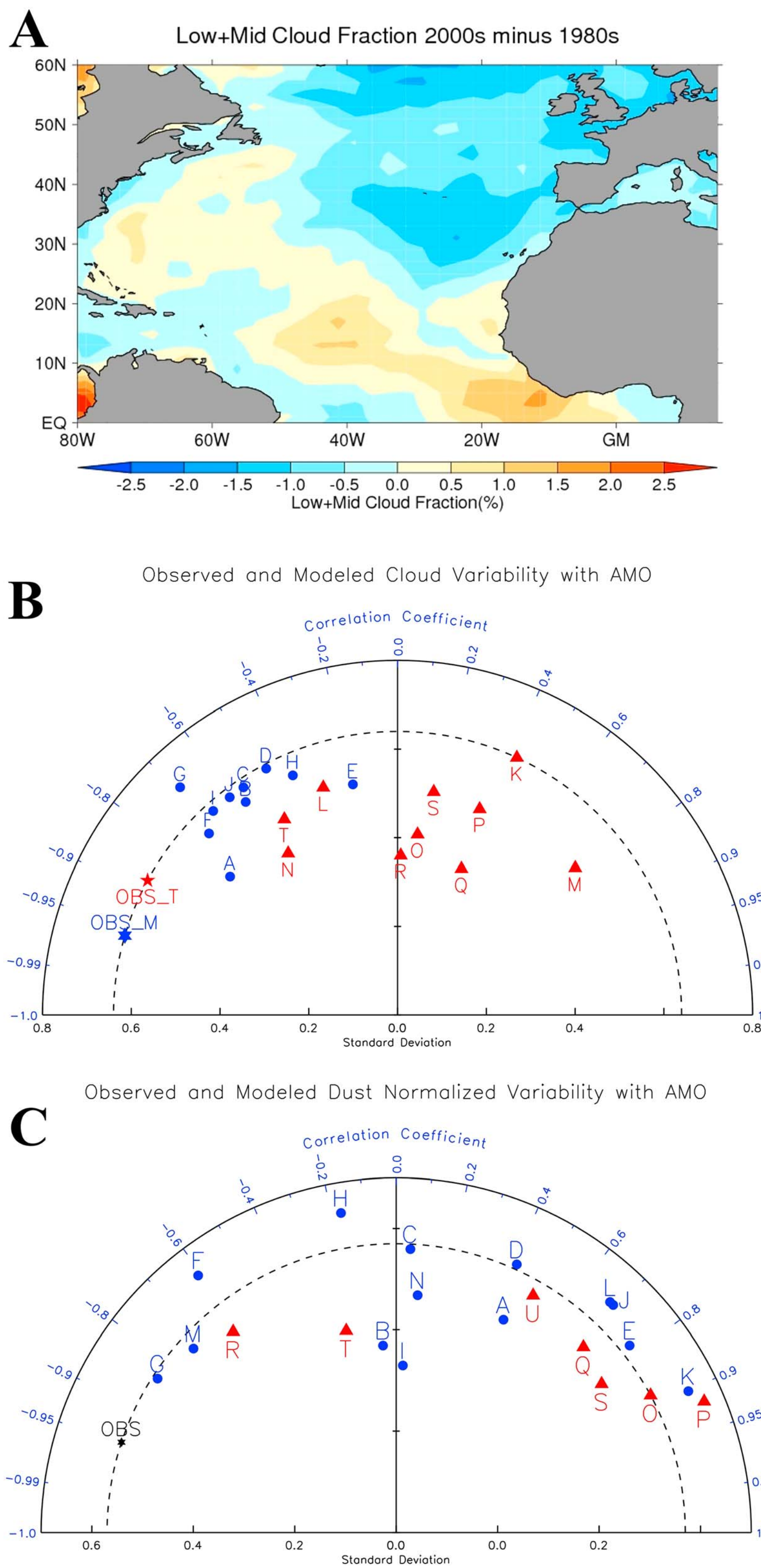

Figure 4. (a) Multimodel mean of composite difference in low cloud fraction between 2000s and 1980s. (b) A diagram showing correlation and standard deviation of normalized cloud fraction anomaly time series over midlatitude (blue) and tropical (red) arms of AMO using model results. Observed values are shown as stars with corresponding colors. (c) Similar to Figure 4b, but for normalized tropical Atlantic dust loading in models for coupled, in blue, and AMIP style, in red, runs.

critical for both feedback mechanisms. Failure to correctly capture the trade wind speed response to SST anomalies prevent two positive feedbacks to operate and the lack of radiative impact from two positive feedbacks deprives models of mechanisms to generate the observed tropical arm of AMO. This interpretation is corroborated by other analyses of modeling results [*Kang et al*., 2009; *Zhang et al*., 2010; *Hodson et al*., 2014; *Martin et al*., 2014] and the two feedback mechanisms together with the wind response offer a viable process with which the middle latitude SST anomalies can propagate into the tropics. The lack of trade wind response to AMO may be related to the fact that modeled North Atlantic Oscillation response to AMO is not correctly simulated [*Omrani et al*., 2014].

While evidence and mechanisms presented here are physically consistent, we recognize that they do not constitute causal proof. Unfortunately, because current models lack the ingredients to capture the wind response and two positive feedback mechanisms, we could not use numerical experiments to model the full causal chain. Improved model physics (e.g., cloud physics in CESM-CAM5) and coupled dynamics may make such experiments possible in the near future. We also want to note that while the importance of dust and cloud feedbacks are highlighted here, other factors may also play a role, such as the impact of wind speed change on WES, the influence of Sahel precipitation and vegetation on dust emission, model biases in mean states such as SST [*Wang et al*., 2014] and the strength of aerosol indirect effects [*Yuan et al*., 2011; *Booth et al*., 2012; *Doherty and Evan*, 2014], which, in our opinion, are not the primary cause of cloud fraction change (see discussion in *Zhang et al*. [2013] and the SOM).

Still, the framework proposed here links variability of North Atlantic SST, dust loading and marine low clouds together through large-scale coupled ocean-atmospheric dynamics (Figure 3), and is useful for a host of research topics. For example, the positive dust feedback could explain the drastic increase of Saharan dust emission when the North Atlantic was abnormally cold [*McGee et al*., 2013; *Murphy et al*., 2014]. It also may offer clues to how changed dust emission may fertilize the ocean, alter biological productivity, and thus perturb CO_2 sequestration by the ocean as observed in paleoclimate records [*Martinez-Garcia et al*., 2011]. The radiative impact by both positive feedbacks on the top-of-atmosphere energy balance is important to understand the response of tropical precipitation and circulation to both AMO and other extratropical forcings (e.g., ice cover change) [*Kang et al*., 2008; *Zhang et al*., 2010; *Frierson et al*., 2013]. In addition, to fully realize decadal predictability provided by nature, our results suggest that efforts are needed for improving modeled atmospheric response to North Atlantic SST anomalies and associated cloud and dust feedbacks. Finally, the robust and fast low cloud feedback, if consistently operating across ocean basins [*Clement et al*., 2009], is a powerful amplifier for SST anomalies regardless of its source, be it anthropogenic global warming, aerosol cooling, or natural variability. The lack of strong positive cloud feedback in many models might partly explain why, despite the ability to simulate the modern secular warming trend, modeled multidecadal fluctuations around this trend are too weak compared to observations [*Ting et al*., 2009].

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Erratum

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