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- Multidecadal cloud feedbacks amplify the tropical branch of the AMO
- Observed cloud cover covaries with the AMO on multidecadal timescales
- Cloud feedbacks account for 10%-31% of the observed tropical SST anomalies of the AMO

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- Supporting Information S1

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New observational evidence for a positive cloud feedback that amplifies the Atlantic Multidecadal Oscillation

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Abstract The Atlantic Multidecadal Oscillation (AMO) affects climate variability in the North Atlantic basin and adjacent continents with potential societal impacts. Previous studies based on model simulations and short-term satellite retrievals hypothesized an important role for cloud radiative forcing in modulating the persistence of the AMO in the tropics, but this mechanism remains to be tested with long-term observational records. Here we analyze data sets that span multiple decades and present new observational evidence for a positive feedback between total cloud amount, sea surface temperature (SST), and atmospheric circulation that can strengthen the persistence and amplitude of the tropical branch of the AMO. In addition, we estimate cloud amount feedback from observations and quantify its impact on SST with idealized modeling experiments. From these experiments we conclude that cloud feedbacks can account for 10% to 31% of the observed SST anomalies associated with the AMO over the tropics.

1. Introduction

The Atlantic Multidecadal Oscillation (AMO) is the dominant mode of climate variability in the North Atlantic [*Kerr*, 2000]. It is characterized by a horseshoe pattern of coherent sea surface temperature (SST) anomalies across the North Atlantic (Figure 1), which changes phase on multidecadal timescales (Figure 2a) [*Deser and Blackmon*, 1993; *Kushnir*, 1994]. The SST anomalies associated with the AMO have far-reaching impacts over the surrounding continents, both during winter and summer [*Enfield et al.*, 2001; *Sutton and Hodson*, 2005; *Omrani et al.*, 2014; *Peings and Magnusdottir*, 2014]. In particular, the tropical branch of the AMO is linked to changes in hurricanes and rainfall over Europe, North America, and Africa [*Goldenberg et al.*, 2001; *Knight et al.*, 2006; *Zhang and Delworth*, 2006; *Sutton and Hodson*, 2007].

Previous studies have shown that climate models in the Coupled Model Intercomparison Project phase 5 (CMIP5) archive can correctly simulate the amplitude of the AMO in the midlatitudes; however, the same models underestimate the amplitude and persistence of the AMO in the tropics [*Kavvada et al.*, 2013; *Ba et al.*, 2014]. Recent studies have suggested that feedbacks and forcing from SST, clouds, dust, and aerosols may help propagate SST anomalies from the midlatitudes to the tropics [*Brown et al.*, 2016; *Yuan et al.*, 2016] and amplify the tropical branch of the AMO [*Mann and Emanuel*, 2006; *Booth et al.*, 2012; *Evan et al.*, 2013; *Martin et al.*, 2014; *Zhang et al.*, 2010]. In particular, *Yuan et al.* [2016] examined changes in low-level cloud cover over the tropical North Atlantic in satellite data sets and provided observational evidence for a positive low-level cloud feedback on the AMO. They found positive cloud cover anomalies from the beginning of the satellite record to the mid-1990s when the AMO was in its negative phase, and negative cloud cover anomalies from the mid-1990 to 2009 when the AMO was in its positive phase. They compared observations with models in the CMIP5 archive, finding that models underestimated the strength of the observed positive low-level cloud feedback.

Here we build upon these previous studies, in particular *Yuan et al.* [2016], and quantify the role of cloud feedbacks in amplifying the tropical branch of the AMO. The cloud anomalies found by *Yuan et al.* [2016] are not detrended and so are affected by long-term trends due to greenhouse gas forcing. Instead, here we examine changes in detrended cloud cover in ship-based observations, which span a longer period of time from 1954 to 2008 and therefore cover two phase shifts of the AMO. Further, we provide an estimate of the change in SST due to the observed cloud amount feedback using an idealized modeling approach.

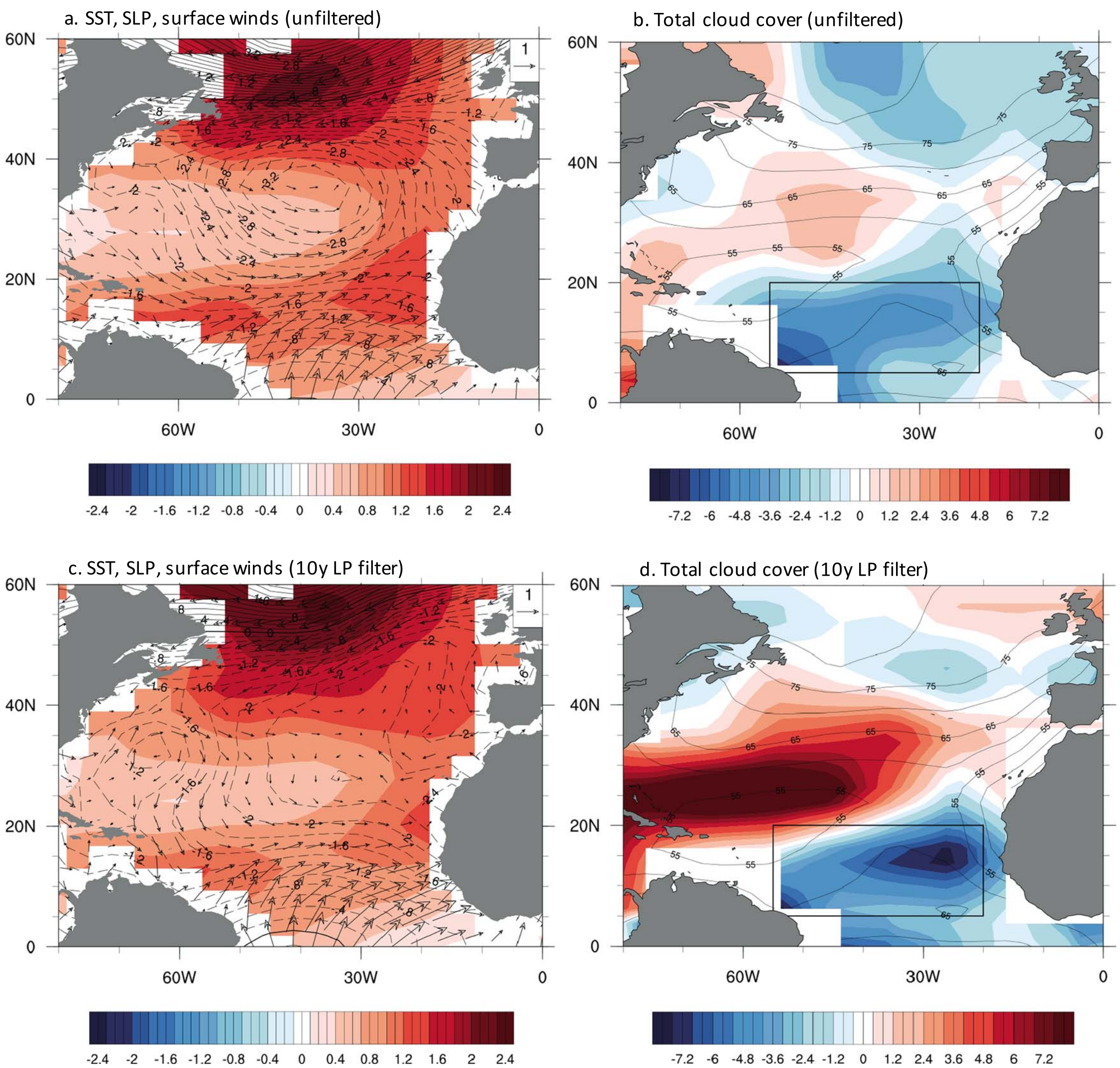

Figure 1. Regressions on the NASST index (SST averaged 0°–60°N, 80°W–0°): (a) Regressions of SST (shaded), SLP (contours), and surface winds (vectors) for the years 1950–2014. Contours range from −4 hPa to 4 hPa with intervals of 0.2 hPa. (b) Regression of total cloud cover (shaded) from EECRA and climatological mean (contours, units of %) for the years 1954–2008. (c and d) Same as Figure 1a and 1b but filtered with at 10 year low-pass Lanczos filter. Units for regressions are of K (SST), hPa (SLP), ms^{-1} (surface winds), and % (total cloud cover) per K^{-1} of the NASST index.

2. Data and Methodology

2.1. Observations

We use the Extended Reconstructed SST, version 3b (ERSSTv3b) reanalysis [*Smith et al.*, 2008] for SST, and the National Centers for Environmental Prediction-National Center for Atmospheric Research reanalysis [*Kalnay et al.*, 1996] for sea level pressure (SLP) and surface winds for the years 1950–2014, while radiative fluxes are from the Clouds and Earth's Radiation Energy System (CERES) Energy and Balance Filled, edition 2.8 (EBAF_Ed2.8) data set for the years 2001–2014 [*Loeb et al.*, 2009].

We analyze total cloud cover from ship-based observations in the Extended Edited Cloud Reports Archive (EECRA) for the years 1954–2008. This archive is on a 10°×10° grid with units of percent [*Hahn and Warren*, 2009; *Eastman et al.*, 2011]. We mask out data where there are less than 25 observations in each grid box per season, as recommended by *Eastman et al.* [2011]. The EECRA data set is affected by a spurious positive trend of unknown origin, which is coherent across all latitudes. We correct this bias by subtracting the global mean cloud cover for each year. Previous studies have shown that once this bias is taken into account, cloud observations are consistent with other satellite products [*Bellomo et al.*, 2014] and meteorological

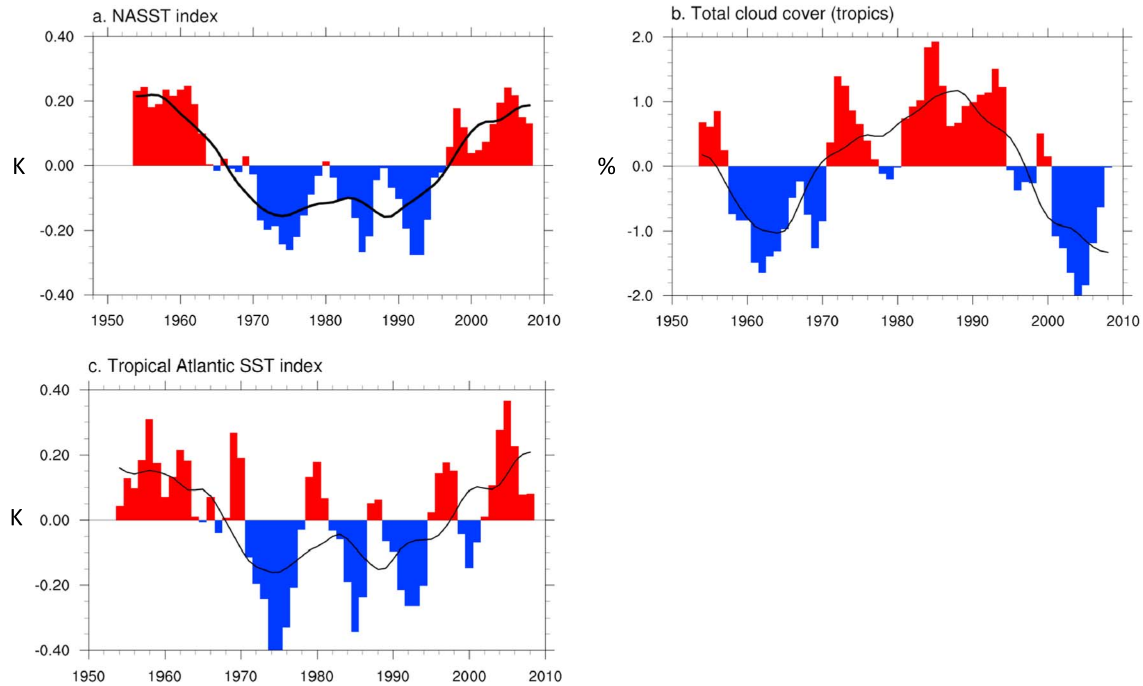

Figure 2. Time series of spatially averaged annual mean anomalies (a) NASST index, SST averaged over 0°–60°N, 80°W–0°, (b) Total cloud cover in the tropics averaged over 5°N–20°N, 55°W–20W° (same as box in Figure 1). (c) Tropical Atlantic SST (TASST) index averaged as in Figure 2b. Superimposed in solid black lines in all panels are the 10 year running averages. A 1-2-1 weighted running average is performed on all time series. Units are of kelvin for SST and percent for total cloud cover.

parameters [*Clement et al*., 2009; *Deser et al*., 2010] over the overlapping years of coverage. Moreover, the spurious bias affects mostly the estimates of long-term trends rather than internal variability, which is the focus of this study. In addition to ship-based observations, we analyze total and low (below 680 hPa) cloud cover from the International Satellite Cloud Climatology Project (ISCCP) satellite data set for the years 1984–2007. ISCCP is on a 2.5° × 2.5° grid with units of percent [*Rossow and Schiffer*, 1999]. Cloud cover in the ISCCP database is corrected for errors introduced by instrumentation failures and orbital drifts. Remaining errors are corrected by regressing out global mean long-term trends as described in *Norris and Evan* [2015].

Because we focus on multidecadal variability, we detrend all data sets, except ISCCP. ISCCP cannot be detrended due to its short coverage, which encompasses only one phase shift of the AMO [cf. *Yuan et al*., 2016]. We form monthly mean anomalies by removing the climatological monthly mean from each month except for the EECRA data set, for which we form seasonal mean anomalies by removing the climatological seasonal mean from each season. All months (and seasons for EECRA) are then averaged to produce annual mean anomalies. All data sets are regridded to a common 2.5° × 2.5° grid before plotting.

2.2. Model

We perform model experiments using an atmospheric general circulation model (AGCM) (European Centre/Hamburg (ECHAM) 6, version 6.1.04) coupled to a slab-ocean for the open ocean and a thermodynamic sea ice model [*Stevens et al*., 2013]. The model is run with a resolution of T31 for the horizontal grid (3.75° × 3.75°) and 31 vertical levels. The mixed layer depth of the slab-ocean model is fixed to 50 m everywhere and does not vary seasonally. In the slab-ocean configuration, interactive ocean dynamics are absent and internal climate variability is driven by radiative and turbulent fluxes at the ocean surface (i.e., shortwave and longwave radiation plus latent and sensible heat fluxes). The monthly mean ocean heat transport (commonly referred as *q-flux*) is prescribed to maintain the observed SST climatological mean but does not change from year to year. All simulations are integrated for 100 years, but the first 20 years are discarded from the analysis to avoid the possible influence of the initial conditions.

3. Results

The AMO pattern is typically shown as the regression onto the low-pass-filtered NASST (North Atlantic SST) index [*Enfield et al.*, 2001], which in this study is computed as the SST averaged over 0°–60°N, 80°W–0°. Figure 1 shows the regression of SST anomalies, surface winds, and SLP on the unfiltered NASST index (Figure 1a) and on the 10 year low-pass-filtered NASST index (Figure 1c). Low-pass filtering the SST anomalies before computing the regression does not change the observed SST horseshoe pattern (Figure 1). Largest anomalies occur at 50°N and north in both cases. Relatively larger SST anomalies occur in the midlatitudes (40°N–60°N) and in the tropics (0°–20°N) with and without the low-pass filter, although the amplitude of the anomalies below 50°N becomes smaller after applying the low-pass filter.

Over the last 60 years, the AMO changed from positive to negative in the mid-1960s and from negative to positive in the mid-1990s (Figure 2a). The AMO is accompanied by a weakening of the subtropical Azores high-pressure system (Figures 1a and 1c), and an equatorward displacement of the midlatitude jet [*Woollings et al.*, 2015]. Along with a weakening of the subtropical high, westerly anomalies in the tropics and easterly anomalies in the midlatitudes oppose the climatological mean wind circulation providing a positive latent heat flux feedback that reinforces warm SST anomalies [*Clement et al.*, 2015; *Seager et al.*, 2000]. Changes of opposite sign occur during the negative phase of the AMO reinforcing cold SST anomalies.

A recent study by *Yuan et al*. [2016] shows that satellite observations over the last 30 years are consistent with a positive low-level cloud feedback that can strengthen the SST anomalies associated with the AMO in the tropics. In particular, they show that low-level cloud cover anomalies have changed sign concurrently with the AMO around the mid-1990s. Low-level cloud cover increased during the cold phase of the AMO and decreased during the warm phase. However, satellite data sets cover only one phase change of the AMO; hence, signals of multidecadal variability are masked by long-term trends due to greenhouse gas forcing. The data analyzed by *Yuan et al*. [2016] are, in fact, not detrended.

Figure 1 shows the regression of detrended ship-based total cloud cover (EECRA) covering the years 1954–2008 onto the unfiltered (Figure 1b) and 10 year low-pass-filtered (Figure 1d) NASST index. Unlike SST (Figures 1a and 1c), total cloud cover anomalies display different patterns associated with unfiltered and low-pass-filtered NASST. In the unfiltered case (Figure 1b), total cloud cover decreases in both midlatitudes and tropics, while in the low-pass-filtered case the midlatitude cloud cover anomalies are very small compared to the tropics. We choose a box over the tropics (Figures 1b and 1d) and show in Figure 2 that cloud cover is anticorrelated with both tropical Atlantic SST (TASST) anomalies (Figure 2c; correlation coefficient is -0.63) and the NASST index (Figure 2a; correlation coefficient is -0.68). A reduction in cloud cover is seen during the two positive phases of the AMO, while an increase is seen during the negative phase (Figure 2b).

To compare these results with the ISCCP data set, we plot the not detrended unfiltered and low-pass-filtered regressions of total cloud cover for the years 1984–2007 (Figure S1 in the supporting information). ISCCP and EECRA display the same pattern of total cloud cover anomalies in the unfiltered case. Instead, in the low-pass filter case, ISCCP displays more negative anomalies in the midlatitudes than EECRA. Figures 2b and S2 further show that the time series of cloud cover anomalies averaged over the tropics in EECRA are consistent with the anomalies retrieved by ISCCP for the last 30 years. However, we note that anomalies are larger for ISCCP data than for EECRA.

In this study we use total cloud cover in the EECRA data set, but we note that over the tropical box in Figure 1 b total cloud cover reflects mostly low-level cloud coverage. In fact, Figure S2 shows the time series of nondetrended total cloud cover and low-level cloud cover from ISCCP averaged over the tropical box as in Figure 2b. Low-level cloud cover, which is the same as analyzed by *Yuan et al*. [2016], is consistent with the ship-based total cloud cover, while total cloud cover exhibits some discrepancies.

To quantify the radiative impact of the observed cloud cover changes, we estimate cloud amount feedback from observations. Cloud amount feedback refers to the feedback due to changes in cloud cover and is different from total cloud feedback, which also takes into account vertical variations in cloud distributions and changes in cloud optical properties [*Bellomo et al.*, 2014; *Zelinka et al.*, 2012]. To estimate cloud amount feedback, we follow the methods outlined in *Bellomo et al*. [2014]: we first compute the cloud amount

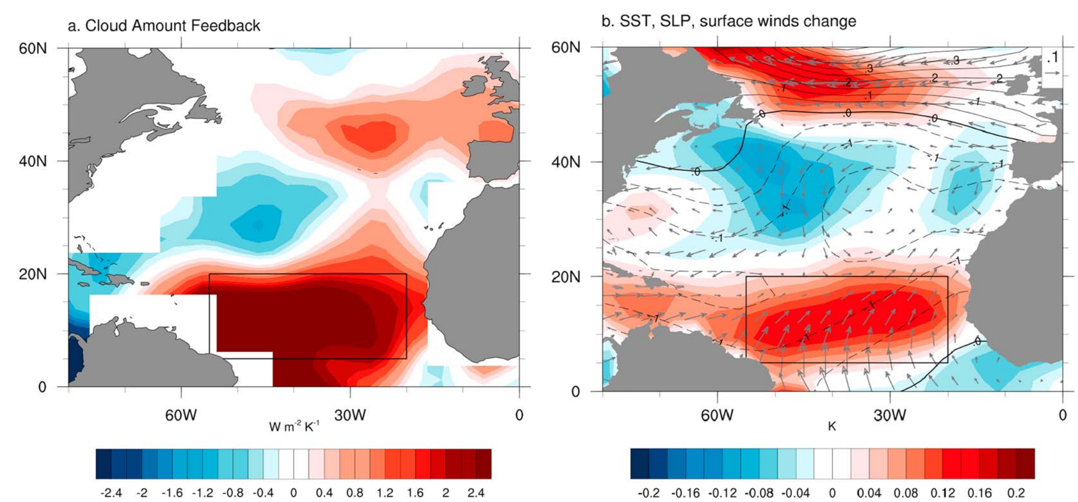

Figure 3. Cloud amount feedback. (a) Observational estimate of cloud amount feedback computed as cloud amount radiative kernel (Figure S3) multiplied by the regression of cloud amount on the NASST index (Figure 1b). Units are W m^{-2}K^{-1}. (b) Change in mean SST (shaded), SLP (contours), and surface winds (vectors) due to the imposed cloud amount feedback in the slab-ocean model experiment. Units are of kelvin (SST), hPa (SLP), and ms^{-1} (surface winds). Contours range from −0.2 hPa to 0.2 hPa with intervals of 0.02 hPa.

radiative kernel as the mean cloud radiative effect (shortwave plus longwave, all-sky minus clear-sky radiation flux at the surface) from CERES divided by mean total cloud cover from EECRA. The cloud amount radiative kernel (Figure S3) quantifies the radiative impact associated with 1% change in cloud cover and has units of W m^{-2} %$^{-1}$ (radiative impact is defined positive down). We then compute cloud amount feedback associated with NASST variability by multiplying the cloud amount radiative kernel (Figure S3) by the regression of cloud cover on the NASST index (Figure 1b). The observational estimate of cloud amount feedback associated with the unfiltered NASST index is plotted in Figure 3a. The average value over the tropical box is 2.93 W m^{-2} K^{-1}, while the average cloud amount radiative kernel is −0.77 W m^{-2} %$^{-1}$. If we repeat the same calculations for low-pass-filtered fields, we find that the average cloud amount feedback over the tropics is 3.07 W m^{-2} K^{-1}, not much different from the unfiltered case.

Next, we quantify the impact of the estimated cloud amount feedback on SST by performing idealized model experiments using an AGCM (ECHAM6) coupled to a slab-ocean model. We first run a control simulation with greenhouse gases and aerosols fixed at the preindustrial levels, and a *q-flux* computed from a control experiment with fixed SST that maintains the monthly mean SST climatology approximately equal to observations. We compare this control simulation with an experiment in which we impose the observed cloud amount feedback (Figure 3a) as an additional positive term in the *q-flux*. The difference between the cloud amount feedback experiment and the control run provides an estimate of the impact of the observed cloud amount feedback on the NASST. We note here that the added *q-flux* values of estimated cloud amount feedback refer to a change in 1 K of the NASST index. Moreover, because the concentration of greenhouse gases and aerosols does not change, the impact of cloud amount feedback on SST is just due to natural variability in these simulations, while in the real world the interplay of clouds, aerosols, and greenhouse gases also affects the magnitude of the radiative impact of clouds.

Figure 3b shows the difference in SST, SLP, and surface winds between the two simulations. Imposing the observed cloud amount feedback results in a SST change that resembles the AMO pattern in the midlatitudes and tropics. The change in SST shown in Figure 3b (ΔSST) corresponds to an imposed cloud amount feedback per 1 K change in NASST; therefore, we need to scale this value. Because we focus on the tropics, we evaluate the effect of cloud amount feedback over the tropical box: first, we compute $\overline{\mathrm{TASST}^{+}}$ as the average of positive TASST index anomalies (Figure 2c). Over the period 1954–2008, $\overline{\mathrm{TASST}^{+}}$ is 0.08 K, while the largest

Table 1. Values Averaged Over the Tropical Box 5°N–20°N, 55°W–20°W for the Two Cloud Data Sets and Model Experiments[a]

Data Set	Cloud Amount Radiative Kernel($W\,m^{-2}\,\%^{-1}$)	Cloud Amount Feedback ($W\,m^{-2}\,K^{-1}$) Unfiltered	Cloud Amount Feedback ($W\,m^{-2}\,K^{-1}$) 10 year Low-Pass Filter	$\%SST_{cloud}$	$\%SST_{cloud\text{-}max}$
EECRA	−0.77	2.93	3.07	10%	16%
ISCCP	−0.76	4.25	5.21	17%	31%

[a]The percent change in SST in the tropical box (change in SST due to cloud amount feedback) is computed as $\%SST_{cloud} = \left(\Delta SST \times \overline{TASST}^{+}\right)/\overline{TASST}^{+}$ (see text for additional details).

positive anomaly, $TASST^{+}_{max}$, is 0.37 K. The scaled change $\%SST_{cloud}$ is the percent contribution of cloud amount feedback to $\overline{TASST}^{+}$ and can be obtained as

$$\%SST_{cloud} = \left(\Delta SST \times \overline{TASST}^{+}\right)/\overline{TASST}^{+} \quad (1)$$

The change in SST (ΔSST) averaged over the tropical box is +0.1 K (Figure 3b); hence, $\%SST_{cloud}$ calculated using (1) is 10%. The maximum value of ΔSST in the tropical box is +0.16 K; hence, the maximum contribution of cloud amount feedback $\%SST_{cloud\text{-}max}$ is 16%. Because cloud amount feedback over the tropics is approximately the same when low-pass filtered, the impact on tropical SST would be similar in an experiment run with low-pass-filtered cloud amount feedback, while in the midlatitudes it would be smaller.

To corroborate these results, we run a similar model experiment using ISCCP cloud cover data. The difference between the two estimates are that we use ISCCP cloud cover in the estimation of the cloud amount feedback and we run the model simulation imposing this cloud amount feedback as an additional term in the *q-flux*. Moreover, the ISCCP data cover only the years 1984–2007 and are not detrended. The cloud amount feedback estimates from ISCCP, as well as the change in SST, SLP, and surface winds are displayed in Figure S5, while the relative cloud amount radiative kernel is plotted in Figure S4. The average cloud amount feedback in the tropical box estimated with ISCCP is 4.25 $W\,m^{-2}\,K^{-1}$ (5.21 $W\,m^{-2}\,K^{-1}$ when low-pass filtered), and the cloud amount radiave kernel is $-0.76\,W\,m^{-2}\,\%^{-1}$. While the value of the cloud amount radiative kernel is very similar to the one computed from EECRA, the cloud amount feedback is larger because cloud cover anomalies are larger in ISCCP than EECRA (cf. Figures S1 and S2 with Figures 1 and 2). $\%SST_{cloud}$ for ISCCP is 17%, and $\%SST_{cloud\text{-}max}$ is 31%. These values are larger than the values obtained from EECRA. All the averages in the tropical box for the observational estimates of cloud amount feedback and model experiments are summarized in Table 1. To test the significance of these results, we computed the *t* test for SST mean change in the tropical box and found it to be significant at the 95% level for both experiments.

Finally, we note that the effect of the imposed cloud amount feedback is not limited to SST. In fact, we see a reduction in the mean wind speed and SLP (Figure 3b) especially in the trade wind region, which together suggest that there is a positive feedback between cloud cover, SST, and atmospheric circulation that can strengthen the AMO in the tropics. The wind response to the imposed cloud amount feedback in the tropics is consistent with the Wind-Evaporation-SST (WES) feedback proposed by *Xie and Philander* [1994]: an initial SST warm anomaly reduces the latent heat flux downwind of the anomaly, which favors the propagation of warm SST along the mean trade winds track. Low-level clouds in the tropics decrease in response to both warmer SST and weaker wind speed, thereby increasing the radiative flux at the surface and reinforcing the positive SST anomaly. The response in the midlatitude circulation is intriguing. It is currently an area of debate as to whether the midlatitude atmospheric circulation and cloud cover respond to SST changes, and on which timescales. Both EECRA and ISCCP data seem to suggest that there might be a small positive cloud amount feedback [cf. *Yuan et al.*, 2016, Figure 1]. However, cloud amount feedback is much smaller in the midlatitudes than in the tropics when a 10 year low-pass filter is applied (cf. Figure 1d).

The values of cloud amount feedback and its estimated impact on the AMO can be compared with alternative estimates in the literature. *Yuan et al.* [2016] estimate that the radiative impact of low-level cloud change is about 2–4 $W\,m^{-2}\,K^{-1}$. *Yuan et al.* [2016] made this estimate by multiplying the cooling efficiency of low-level clouds per percent change of cloud fraction (1 $W\,m^{-2}\,\%^{-1}$) estimated by *Klein and Hartmann* [1993] by the difference in cloud cover anomalies between the 3 years of highest and lowest AMO index during the Moderate Resolution Imaging Spectometer (MODIS) satellite period (2002–2013) [*Platnick et al.*, 2003]. *Yuan et al.* [2016] find that their approximate range compares well with satellite radiative fluxes retrieved by CERES during the same years as MODIS (see their Figures S5 and S6). Finally, they assert that a cloud radiative forcing of 2–4 W m^{-2} K^{-1} could explain a few tenths of a degree kelvin, which would account for a large portion of the AMO. Despite that *Yuan et al*. [2016] use a higher cooling efficiency than here (1.0 versus 0.77 W m^{-2} %$^{-1}$), they find similar values of cloud radiative impact (2–4 W m^{-2} K^{-1} versus 2.93 W m^{-2} K^{-1}). However, it should be noted that their estimates refer to a difference (and so they should be compared to double our values) and to only anomalies occurred during the positive phase of the AMO index due to the limitations of satellite data coverage. Although our estimates for cloud amount feedback are larger than those estimated by *Yuan et al*. [2016], our estimated impact on tropical Atlantic SST (10%–31%) is much smaller than what it is stated in their study.

Brown et al. [2016] compare the AMO between two model simulations: one with and one without cloud feedback and conclude that cloud feedbacks are a necessary teleconnection mechanism to spread the SST anomalies from the midlatitudes (which would be driven by the Atlantic Meridional Overturning Circulation) to the tropics. In particular, in the run without cloud feedback, the midlatitude SST index loses power at multidecadal timescales, but the power spectrum of tropical Atlantic SST index remains unchanged at those timescales. However, it seems that compared to observations, cloud cover anomalies in the model they used (GFDL-CM2.1) are overestimated in the midlatitudes and underestimated in the tropics, which would explain why the midlatitude NASST loses power at low-frequency variability. Therefore, it remains to be tested whether cloud feedback could provide a feedback mechanism to spread SST anomalies from the midlatitudes to the tropics, or simply strengthen tropical SST anomalies as suggested here.

4. Conclusions

Current climate models underestimate the persistence of tropical SST anomalies associated with the AMO [*Ba et al*., 2014; *Martin et al*., 2014]. Previous studies showed that models underestimate the strength of feedbacks involving cloud, aerosols, and dust processes, which are believed to favor the persistence of multidecadal SST anomalies in the tropics [e.g., *Evan et al*., 2013; *Yuan et al*., 2016]. However, evidence for these mechanisms, in particular involving cloud feedbacks, was limited to not detrended satellite-based observations covering only the last 30 years. The AMO has undergone one shift from negative to positive during this period of time; therefore, satellite-based cloud observations alone cannot prove the existence of a cloud feedback on SST on multidecadal timescales.

In this study, we have addressed this observational gap and provided new evidence for a positive cloud amount feedback on the AMO from ship-based cloud observations. We quantified the impact of the observed cloud variability onto SST with idealized model experiments. From these experiments we concluded that the impact of the observed cloud amount feedback can explain between 10% and 31% of the observed SST anomalies associated with the AMO in the tropics.

The results presented herein are affected by observational uncertainties in both cloud observations and radiative fluxes. The estimated cloud amount feedback and change in SST were calculated on the basis of observed total cloud cover from ship-based observations. To corroborate these results, we calculated cloud amount feedback from ISCCP, which appears to be larger than in ship-based observations in the tropics.

In summary, our results are supportive of the hypothesis that cloud feedbacks favor the persistence of SST anomalies in the tropics via the WES feedback. By detrending the cloud observations, we roughly removed the influence of greenhouse gases. However, we have not examined the possible role of aerosol-cloud interactions on driving phase shifts of the AMO [*Booth et al*., 2012], which remains an open question.

4.1. Data Availability

All observational data sets are available online and documented in climatedataguide.ucar.edu. Output from model experiments is available upon request by contacting the corresponding author.

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