Distinct Patterns of Cloud Changes Associated with Decadal Variability and Their Contribution to Observed Cloud Cover Trends

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(Manuscript received 10 July 2018, in final form 18 June 2019)

ABSTRACT

With the goal of understanding the relative roles of anthropogenic and natural factors in driving observed cloud trends, this study investigates cloud changes associated with decadal variability including the Pacific decadal oscillation (PDO) and the Atlantic multidecadal oscillation (AMO). In the preindustrial simulations of CMIP5 global climate models (GCMs), the spatial patterns and the vertical structures of the PDO-related cloud cover changes in the Pacific are consistent among models. Meanwhile, the models show consistent AMO impacts on high cloud cover in the tropical Atlantic, subtropical eastern Pacific, and equatorial central Pacific, and on low cloud cover in the North Atlantic and subtropical northeast Pacific. The cloud cover changes associated with the PDO and the AMO can be understood via the relationships between large-scale meteorological parameters and clouds on interannual time scales. When compared to the satellite records during the period of 1983–2009, the patterns of total and low cloud cover trends associated with decadal variability are significantly correlated with patterns of cloud cover trends in ISCCP observations. On the other hand, the pattern of the estimated greenhouse gas (GHG)-forced trends of total cloud cover differs from that related to decadal variability, and may explain the positive trends in the subtropical southeast Pacific, negative trends in the midlatitudes, and positive trends poleward of 50°N/S. In most models, the magnitude of the estimated decadal variability contribution to the observed cloud cover trends is larger than that contributed by GHG, suggesting the observed cloud cover trends are more closely related to decadal variability than to GHG-induced warming.

1. Introduction

As the temperature rises in response to increasing greenhouse gas (GHG) concentration, a variety of feedback processes arise, including those due to changes in the lapse rate, water vapor concentration, surface

albedo, and cloud properties. Equilibrium climate sensitivity (ECS), the change of global mean surface air temperature needed to balance the radiative forcing of doubled CO₂ concentration, depends on the magnitudes of the accumulated effect of these feedbacks. Among all the feedback processes, the cloud feedback contributes the most to intermodel spread in ECS (Bony and Dufresne 2005; Caldwell et al. 2016; Dufresne and Bony 2008; Vial et al. 2013). This makes understanding how clouds will change with global warming an abiding goal of climate science.

Observations provide clues of how clouds respond to external forcing in nature. However, variations in cloud cover and radiative fluxes in observations contain mixed signals from natural variability and external forcings. On

Denotes content that is immediately available upon publication as open access.

Supplemental information related to this paper is available at the Journals Online website: <https://doi.org/10.1175/JCLI-D-18-0443.s1>.

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DOI: 10.1175/JCLI-D-18-0443.1

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interannual time scales, variations of clouds are primarily attributable to natural variability (Davis et al. 2013; Li et al. 2014; Li et al. 2017; Yang et al. 2016), while on multidecadal time scales, impacts of external forcing may become detectable for some cloud properties (Marvel et al. 2015; Norris et al. 2016). Dessler (2010) showed that the cloud feedback obtained from natural variability differs significantly from that occurring under CO₂-induced global warming in eight CMIP3 GCMs, serving as a caution call for using short-term observational records to evaluate GCMs' ECS. On the other hand, Zhou et al. (2015) reported that cloud feedbacks in response to natural variability are well correlated ($r = 0.74$) across CMIP5 models with those in response to greenhouse warming, offering the possibility of narrowing uncertainty in ECS via observationally constraining internal cloud variability. Isolating cloud changes due to natural variability and external forcings in satellite observations is an essential step for using observed data to evaluate GCMs and for assessing the relevance of observed cloud trends to cloud feedback and ECS.

Over the historical period, variations of cloud cover and cloud-mediated radiative fluxes are closely linked with variations in both the mean and pattern of SSTs (Gregory and Andrews 2016; Zhou et al. 2015, 2016). The climate feedback parameter shows substantial decadal variations in simulations performed with two atmospheric general circulation models (AGCMs) forced with prescribed observed historical SSTs (Gregory and Andrews 2016). These variations were shown to be driven by changing SST patterns. In similar experiments performed using CAM5 with prescribed historical SSTs, Zhou et al. (2016) showed that changing SST patterns can explain the cloud radiative effect (CRE) variations better than uniform warming over the historical period. They argue that changes in SST patterns lead to changes in the thermodynamic structure of the atmosphere that drive tropical marine low cloud anomalies, which are the primary contributors to CRE variations.

Several large-scale patterns of SST anomalies appear periodically due to internal dynamics of the ocean and atmosphere, and their interactions, recognized as climate modes (Deser et al. 2010). On decadal time scales, the Pacific decadal oscillation (PDO; Mantua et al. 1997) and the Atlantic multidecadal oscillation (AMO; Kerr 2000) represent the most important dynamical modes of global SST variation (Deser et al. 2010; Parker et al. 2007; Tung et al. 2019) and are reported to have important and near-global impacts on the climate system. The PDO has been shown to be associated with anomalous temperature and precipitation in North America

(Barlow et al. 2001; Kurtzman and Scanlon 2007; McAfee 2014; Shukla et al. 2015; Wei et al. 2018), low cloud cover over the North American west coast (Schwartz et al. 2014), summer rainfall in East China (Chan and Zhou 2005; Yang et al. 2017; Yu et al. 2015; Zhu et al. 2015), and the tropical belt width (Allen et al. 2014). The AMO has been shown to impact the nearby regions (Enfield et al. 2001; Goldenberg et al. 2001; Knight et al. 2006; Lyu and Yu 2017; Sutton and Hodson 2005) and can also remotely impact the circulation in the Pacific (Kucharski et al. 2016a,b; Li et al. 2016; McGregor et al. 2014; Zhang and Delworth 2007). During the past four decades, both the PDO and the AMO indices exhibit a nonzero trend. Around the year 2000, the AMO switched from a negative to a positive phase, and the PDO switched from a positive to a negative phase. These strongly influenced the SST trends in recent decades and contributed to a decrease in the warming rate of global mean temperature in the early 2000s (Chen and Tung 2018; Dai et al. 2015; England et al. 2014; Kosaka and Xie 2013; Kucharski et al. 2016a; Li et al. 2016; Maher et al. 2014; McGregor et al. 2014; Meehl et al. 2013; Trenberth et al. 2014). However, the contribution of the phase shift of the PDO and the AMO to the observed cloud cover trends is not well established.

In this study, we investigate the PDO and the AMO impacts on cloud cover, and estimate the contribution of these decadal climate modes to the observed cloud cover trends during the satellite era. The data and methodology are described in section 2. GCM-derived estimates of the PDO- and AMO-induced cloud anomalies in the absence of forcing are described in section 3. The estimated contributions of the decadal variability to the observed cloud cover trends, along with the estimated contributions of GHG forcing, are described in section 4. In section 5 we summarize our findings.

2. Data and methodology

a. Data

The SST data used in this study were obtained from the Hadley Centre Global Sea Ice and Sea Surface Temperature dataset (HadISST; Rayner et al. 2003) and the Extended Reconstructed Sea Surface Temperature version 5 (ERSST v5; Huang et al. 2017) dataset. The HadISST dataset contains monthly global SST and sea ice concentration at 1° spatial resolution from 1870 to the present. The ERSST v5 dataset provides monthly mean global SST at 2° spatial resolution from 1854 to the present. In this study, SSTs from 1920 to 2018 are used. Monthly vertical motion at 500 hPa (ω_{500}) was obtained

from the NOAA-CIRES Twentieth Century Reanalysis V2c (Compo et al. 2011), which only assimilates the observed sea level pressure and provides a long record of data (1851–2014). The period of 1920–2014 is used. We also use ω_{500} from ERA-40 (Uppala et al. 2005), which assimilates satellite, surface, radiosonde and pilot observations. The period of 1957–2002 is used.

Cloud cover was obtained from the International Satellite Cloud Climatology Project (ISCCP; Schiffer and Rossow 1983) and the Pathfinder Atmospheres–Extended (PATMOS-x; Heidinger et al. 2014), two independent satellite cloud cover datasets with long periods of record (1983–2009 for ISCCP and 1982–present for PATMOS-x). The period of 1983–2009 is used. We make use of the ISCCP and PATMOS-x datasets that were adjusted for artifacts arising from changing satellite view angles, changing solar zenith angles, and other sources of spurious trends in the records (Norris and Evan 2015). During the procedure of removing spurious variability, global-mean cloud variability has also been subtracted, so the resulting dataset contains spatial anomalies relative to an unknown global mean value. The monthly total cloud cover, high cloud cover (above 440 hPa), and low cloud cover are used. The low and midlevel clouds in ISCCP are combined to yield a better estimation of low-level cloud cover (Seethala et al. 2015); as a result, the cloud cover below 440 hPa is considered as low cloud cover in ISCCP, whereas in PATMOS-x the low cloud cover is defined as the cloud cover below 680 hPa. These satellite data are provided on an evenly spaced 2.5° latitude–longitude grid. When calculating the spatial correlation between the trend patterns of cloud cover in satellite datasets with those obtained from GCM outputs, we average the satellite data into a $5^\circ \times 5^\circ$ grid to make our results comparable with Norris et al. (2016). The patterns of cloud cover trends obtained from GCM are linearly interpolated into the $5^\circ \times 5^\circ$ grid.

GCM simulations from CMIP5 are used to investigate the cloud responses to decadal variability and GHG forcing (see Table 1). The first 300 years of the pre-industrial control (“piControl”) experiments are used to investigate cloud variations associated with the PDO and the AMO. In each model, the long-term trend is subtracted to remove the impacts of model drift. The removal of the long-term trend does not influence our results. The first 140 years of the “1pctCO2” (1% CO₂) experiments, in which the CO₂ concentration is increased by 1% each year, are used to investigate cloud responses to GHG forcing. A total of 29 models in CMIP5 provide cloud cover data for at least 300 years in

piControl experiments and for the first 140 years of 1% CO₂ experiments. The first ensemble member of each model is used.

The best practice for comparing modeled and observed cloud cover is to use output from satellite simulators applied to model cloud fields, which effectively translates the cloud cover produced by a GCM into fields comparable to what would be observed by a satellite (Bodas-Salcedo et al. 2011). However, this approach greatly limits the number of models for which such a comparison can be made. As an alternative, in this study we compute estimates of high cloud cover and low cloud cover in the models as follows. High cloud cover is defined as the sum of cloud fraction above 440 hPa assuming random overlapping in the vertical direction, in which the joint cloud cover of two layers with cloud fraction C_1 and C_2 is

$$C_{\text{random}} = C_1 + C_2 - C_1 \times C_2. \quad (1)$$

The low cloud cover is defined as the maximum cloud cover below 680 hPa, which is equivalent to assuming clouds are maximally overlapped. In GCMs, the cloud cover estimated by maximum overlapping assumption is correlated with the cloud cover estimated by the ISCCP simulator, but tends to underestimate its temporal variability (Figs. S3 and S4 of Zhou et al. 2015).

b. PDO and AMO signal

The PDO is defined by the leading EOF mode of the monthly anomalous SST in the North Pacific (poleward of 20°N) with global mean SST (GMSST) anomaly subtracted, following Mantua et al. (1997). The corresponding principal component (PC), after low-pass filtering and normalizing by its standard deviation, is defined as the PDO index. For the AMO, we make use of a revised AMO index proposed by Trenberth and Shea (2006), which is defined by the low-pass filtered time series of the area-weighted mean of the SST anomalies in the North Atlantic (0° – 60°N , 0° – 80°W), with near-global mean SST anomaly in 60°S – 60°N subtracted to eliminate the impacts of background warming. In observational data, this revised AMO index yields a similar pattern of AMO-related SST anomalies compared to that of a traditional AMO index, which is defined by the average of North Atlantic SST anomalies of detrended data (Enfield et al. 2001). In GCM simulations, however, the revised AMO index better captures the observed remote impacts of the AMO outside of the North Atlantic, especially in the Pacific (Lyu and Yu 2017). As discussed in Lyu and Yu (2017), this may be associated with a bias of GCMs: the decadal

TABLE 1. Models used for the diagnostics of the preindustrial simulation and 1% CO₂ simulation.

Institution	Model	Years in preindustrial simulation used for diagnostics
Commonwealth Scientific and Industrial Research Organization/Bureau of Meteorology (CSIRO-BOM)	ACCESS1.0	300–599
	ACCESS1.3	250–549
Beijing Climate Center (BCC)	BCC-CSM1.1	1–300
	BCC-CSM1.1m	1–300
Beijing Normal University (BNU)	BNU-ESM	1450–1749
Canadian Centre for Climate Modeling and Analysis (CCCma)	CanESM2	2015–2314
National Center for Atmospheric Research (NCAR)	CCSM4	800–1099
	CESM1-BGC	101–400
	CESM1-CAM5	1–300
Centro Euro-Mediterraneo sui Cambiamenti Climatici (CMCC)	CMCC-CM	1550–1849
Commonwealth Scientific and Industrial Research Organization/Queensland Climate Change Centre of Excellence (CSIRO-QCCCE)	CSIRO-Mk3.6.0	1–300
Geophysical Fluid Dynamics Laboratory (NOAA-GFDL)	GFDL CM3	1–300
	GFDL-ESM2G	1–300
	GFDL-ESM2M	1–300
Goddard Institute for Space Studies (NASA-GISS)	GISS-E2-H	3331–3630
	GISS-E2-R	2410–2709
Institute of Atmospheric Physics, Chinese Academy of Sciences (LASG-CESS)	FGOALS-g2	201–500
Institute of Atmospheric Physics, Chinese Academy of Sciences (LASG-AP)	FGOALS-s2	1850–2149
Russian Academy of Sciences, Institute of Numerical Mathematics (INM)	INM-CM4	1850–2149
Institut Pierre Simon Laplace (IPSL)	IPSL-CM5A-LR	1800–2099
	IPSL-CM5A-MR	1800–2099
	IPSL-CM5B-LR	1830–2129
Atmosphere and Ocean Research Institute (The University of Tokyo), National Institute for Environmental Studies, and Japan Agency for Marine-Earth Science and Technology (MIROC)	MIROC5	2000–2299
	MIROC-ESM	1800–2099
Max Planck Institute for Meteorology (MPI-M)	MPI-ESM-LR	1850–2149
	MPI-ESM-MR	1850–2149
	MPI-ESM-P	1850–2149
Meteorological Research Institute (MRI)	MRI-CGCM3	1851–2150
Bjerknes Centre for Climate Research, Norwegian Meteorological Institute (NCC, NMI)	NorESM1-M	700–999

variability of GMSST is more associated with the SST anomalies in the Atlantic than those in the Pacific in observation, while the reverse is true in GCMs. As a result, in GCM simulations, the traditional AMO index, which includes the signals of GMSST variation, would also include those signals of Pacific SST variation that are biasedly correlated with GMSST, leading to the biased SST responses in the Pacific. The removal of near-global mean temperature in the calculation of the revised AMO index minimizes the impacts of this bias,

making it a more feasible index to apply in GCMs. Using the revised AMO index instead of the traditional AMO index may quantitatively affect our estimation of AMO's contribution to observed cloud cover trend; however, the short observational records and the GCM biases limit us from estimating this potential uncertainty. The monthly anomalous fields of each variable, after low-pass filtering, are regressed onto the PDO index or the AMO index to obtain the PDO- or AMO-related anomalous fields. A Lanczos filter with a

cutoff period of 13 years is used in all the filtering processes.

c. Attributing PDO- and AMO-related cloud cover changes to changes in large-scale meteorological parameters

To attribute cloud changes associated with the PDO and the AMO to changes in large-scale meteorological parameters, a simple linear regression analysis is applied:

$$\frac{dC_x}{d\text{PDO/AMO}} = \frac{dC}{dX} \times \frac{dX}{d\text{PDO/AMO}},$$

$$C \in \{\text{high clouds, low clouds}\},$$

$$X \in \{\omega_{500}, \text{SST}\}. \quad (2)$$

In Eq. (2), the C can be either high cloud cover or low cloud cover, and the left-hand side represents the changes of high cloud cover or low cloud cover corresponding to the PDO- or AMO-related changes of variable X . For high clouds, the variable X is chosen to be ω_{500} , as high clouds are often associated with convection; for low clouds, the variable X is chosen to be SST and this will be discussed later in this section. The first term on the right-hand side is obtained by regressing the unfiltered monthly anomalous cloud cover onto the monthly anomalous field of X . Statistically, the term dC/dX represents the relationship between clouds and not only variable X , but also everything that is correlated with variable X . Since the monthly data we used have a time scale much longer than the adjustment time scales of clouds, the term dC/dX represents the relationships between the large-scale environment and the cloud cover at its quasi-equilibrium state. The second term on the right-hand side represents the responses of variable X to the PDO or the AMO, and can be obtained by regressing the low-pass filtered monthly anomalous field of X onto the PDO index or the AMO index (Fig. 1). The physical meaning of the left-hand side can be thought of as the cloud changes corresponding to PDO- or AMO-related changes of X in a coupled climate system. It is worth noting that constructing our statistical model based on the Eulerian perspective may miss some nonlocal effects from the upwind environment (Clemesha et al. 2017; Klein et al. 1995; Mauger and Norris 2010).

In addition to SST, previous studies find that the variation of low clouds is also related to changes in estimated inversion strength (EIS; Wood and Bretherton 2006) (Myers and Norris 2015; Qu et al. 2014, 2015b; Zhou et al. 2016). In particular, Zhou et al. (2016) reported a dominant role of EIS in driving the decadal variation of tropical marine low clouds over the historical period in CAM5 simulations. Employing multiple

regression analysis with SST and EIS as independent variables to isolate their individual impacts on low clouds results in very little improvement to our statistical model, and SST anomalies explain the overall pattern of PDO- and the AMO-related low cloud changes better than EIS anomalies (Fig. S1 in the online supplemental material). The differences between our analysis and previous studies that report an important role of EIS in driving the variation of low clouds in the historical period may arise for two reasons. First, both external forcings and natural variability contribute to the variation of low cloud cover, SST, and EIS in the historical period. Certain regions can experience large EIS variations and hence large EIS-induced variations of low cloud cover in the historical period, while having few changes of EIS and small EIS-induced low cloud cover changes in the case of the PDO or the AMO. Second, in previous studies, multiple regression analysis is performed on data averaged over larger regions, whereas in the present study multiple regression analysis is applied on grid point data. Taking a regional mean before the multiple regression minimizes the nonlocal effects on low clouds from the upwind environment; also, it reduces the variability of SST, EIS, and low clouds to smaller ranges, in which the linear assumption may hold better and the nonlinear effects may be smaller. In regions with abundant stratiform clouds in CMIP5 models, typical magnitudes of correlations between monthly anomalous SST or EIS and low cloud cover are between 0.2 and 0.3 at the grid box scale. The magnitudes of these correlations increase to 0.5–0.7 if regionally averaged time series of SST, EIS, and low clouds are used (Fig. S2a), indicating that taking the regional mean improves the performance of the regression analysis. We have also examined multiple regression analysis on larger domain-averaged quantities (Fig. S2), similar to those methods used by previous studies (Qu et al. 2014, 2015b; Zhou et al. 2016). Off the coast of California, in the equatorial eastern Pacific, and in the tropical North Atlantic, the contributions of EIS to low cloud cover changes are larger than the contributions of SST. However, in the subtropical northeast Pacific and the midlatitudes of the northwest Pacific and the North Atlantic, the EIS contributions are much smaller. We choose to use grid point data for the statistical model, as they have the benefit of explaining the overall patterns of the PDO- and AMO-related low cloud changes, which are the particular interest of the present study. SST appears to be the variable that works the best for our grid point statistical model. We acknowledge that SST is not independent of many typical environmental conditions that are often used to understand low cloud changes such as inversion stability, subsidence, horizontal temperature advection,

FIG. 1. Regression coefficient of SST and ω_{500} on the (left) PDO index and (right) AMO index: (a),(b) SST obtained from HadISST, (c),(d) ω_{500} from the Twentieth Century Reanalysis, and (e),(f) SST and (g),(h) ω_{500} obtained from CMIP5 model simulations. Dots in (e)–(h) indicate over 80% of models agree in sign.

humidity, and surface wind speed. The influences of these cloud-controlling factors on low clouds are also implicitly included in our single-variable statistical model to the extent that they are correlated with SST.

d. Estimating the impacts of decadal variability and GHG forcing on observed cloud cover trends

The contribution of decadal variability to the observed trend of cloud cover from 1983 to 2009 in satellite records is estimated by

$$\frac{dC_{DV}}{dt} = \frac{\partial C}{\partial \text{PDO}} \times \frac{d\text{PDO}}{dt} + \frac{\partial C}{\partial \text{AMO}} \times \frac{d\text{AMO}}{dt}. \quad (3)$$

The term dC_{DV}/dt represents the estimated contribution of decadal variability to cloud cover trends during the

satellite era. The term C in Eq. (3) represents the spatial-temporal fields of total cloud cover, high cloud cover, or low cloud cover, whereas the PDO and the AMO in Eq. (3) represent the time series of the PDO and the AMO indices. The terms $\partial C/\partial \text{PDO}$ and $\partial C/\partial \text{AMO}$ represent the cloud cover changes corresponding to a unit change of the PDO or the AMO indices, and are obtained by multiple linear regression analysis, with the PDO index and the AMO index being the independent variables and the monthly anomalous cloud cover being the dependent variable. This approach, however, has the potential to underestimate the AMO impacts in the Pacific region, as the two climate modes are not independent of each other and a portion of the variation in the PDO index may be driven by the AMO (Ham et al. 2013a,b; Kucharski et al. 2016a; Zhang and Delworth

2007). Physically, it may be more reasonable to attribute the impacts of this AMO-driven variation of the PDO index as part of the AMO impacts. However, the multiple linear regression analysis applied in Eq. (3) tends to attribute this AMO-driven PDO as part of the PDO impacts, since it is more closely linked with the PDO index than the AMO index. Compared to the regression of cloud cover fields on the AMO index obtained by a simple linear regression method, in which the PDO signal is not isolated, the magnitudes of the AMO impacts on cloud cover in the Pacific are reduced by about 30%–50% when the PDO signal is subtracted by applying the multiple linear regression analysis, which provides an upper bound of the underestimation of the AMO impacts in the Pacific due to the shortage of the statistical method. To better compare with the satellite record, in which the monthly mean trend of cloud cover within 60°S–60°N has been subtracted in the procedure of removing spurious signals, the 60°S–60°N mean values of $\partial C/\partial \text{PDO}$ and $\partial C/\partial \text{AMO}$ are subtracted. The terms $d\text{PDO}/dt$ and $d\text{AMO}/dt$ represent the trend of the PDO index and the AMO index, respectively. Because of the phase shift around the year 2000, over the period of 1983–2009 the trend of the PDO index is -0.98 standard deviations per decade and the trend of the AMO index is $0.16 \text{ K decade}^{-1}$ in the HadISST. The ERSST also shows a similar magnitude of trends of PDO index (-1.03 standard deviations decade^{-1}) and AMO index ($0.15 \text{ K decade}^{-1}$). The mean values of -1.0 standard deviations per decade for $d\text{PDO}/dt$ and $0.16 \text{ K decade}^{-1}$ for $d\text{AMO}/dt$ are used.

We can further substitute the terms $\partial C/\partial \text{PDO}$ and $\partial C/\partial \text{AMO}$ in Eq. (3) with the estimations from observational data by Eq. (2):

$$\frac{dC_{\text{DV-obs}}}{dt} = \frac{dC}{dX} \times \frac{\partial X}{\partial \text{PDO}} \times \frac{d\text{PDO}}{dt} + \frac{dC}{dX} \times \frac{\partial X}{\partial \text{AMO}} \times \frac{d\text{AMO}}{dt}. \quad (4)$$

The left-hand side of Eq. (4) represents the estimations of the contribution of decadal variability to the trends of cloud cover during the satellite era, estimated entirely by observational data instead of GCM outputs. As in Eq. (2), the C can be either high cloud cover or low cloud cover, and the X can be either ω_{500} or SST depending on the meaning of C . This method is not applied on high clouds and ω_{500} , however, because the patterns of anomalous ω_{500} show much larger uncertainty than the patterns of anomalous SST (Fig. 1; see also Fig. S3). As described in section 2c, the term $d\text{LCC}/d\text{SST}$ is obtained by regressing local low cloud cover anomalies onto local SST anomalies. The terms $\partial \text{SST}/\partial \text{PDO}$ and $\partial \text{SST}/$

∂AMO are obtained by multiple linear regression analysis.

For comparison, the contribution of GHG forcing on the observed trend of total, high, and low cloud cover is estimated by

$$\frac{dC_{\text{GHG}}}{dt} = \frac{dC}{d\text{GMT}} \times \frac{d\text{GMT}}{dt}. \quad (5)$$

The term dC_{GHG}/dt represents the contribution of GHG forcing. GMT represents the global mean surface temperature and is adopted as an index to quantify the impact of GHG forcing on the climate system. The term $dC/d\text{GMT}$ represents the cloud cover changes corresponding to 1-K change GMT and is obtained by regressing monthly anomalous cloud fields onto GMT in 1% CO_2 simulations. As in the case of decadal variability, the 60°S–60°N mean value of $dC/d\text{GMT}$ is subtracted. The term $d\text{GMT}/dt$ represents the rate of GMT change, and is estimated by the trend of GMT over the period of 1960–2013, in which both the HadISST and the ERSST datasets show a warming rate of $0.1 \text{ K decade}^{-1}$.

3. PDO and AMO in climate model simulations

a. Characteristics of the PDO in model simulations

To set the stage for understanding the PDO-related changes of clouds, we first examine the changes of SST and circulation associated with the PDO in observational data and GCM simulations. The positive phase of PDO is characterized by negative SST anomalies in the midlatitude central Pacific straddled by positive horseshoe-shaped SST anomalies along the west coast of North America (Figs. 1a,e). As reported by previous studies, this pattern arises from several mechanisms, including the decadal variation of the Aleutian low caused by midlatitude stochastic forcing and remote influences from the tropics, the “reemergence” effect of the ocean due to oceanic thermal inertia, and the dynamical adjustments of the ocean gyre to basinwide anomalous wind forcing (Newman et al. 2016, and the citations therein). Outside of the North Pacific, the positive phase of the PDO is accompanied by the overall warming of the tropics with a maximum in the equatorial eastern Pacific. These features are consistent with the PDO-related SST anomalies in the ERSST v5 dataset (Fig. S3a) and are robust among the models.

The PDO-related ω_{500} changes in GCMs (Fig. 1g) show an enhancement of ascending motion in the tropical central Pacific and equatorial shift of the intertropical convergence zone (ITCZ) and the South Pacific convergence zone (SPCZ). The band of anomalous ascending motion extends south and east, to the subtropical South Pacific region. In

the western North Pacific, the subtropical subsidence zone weakens and expands into midlatitudes, while in the eastern North Pacific there is anomalous ascent in the midlatitudes and anomalous descent around 15°N. These features are consistent across models, and are also roughly consistent with those in the ERA-40 and the Twentieth Century Reanalysis (Fig. 1c; see also Fig. S3c), except for the anomalous descent in the subtropical northeast Pacific that is not shown in the reanalysis datasets.

Figure 2a demonstrates the changes in total cloud cover associated with the PDO in CMIP5 models. The cloud cover increases in the central-to-western equatorial Pacific, and decreases over the Maritime Continent and the Indian Ocean. In the subtropical South Pacific, the cloud cover decreases on the western side and increases on the eastern side. In the subtropics and midlatitudes of the North Pacific, there is increasing cloud cover in the western side, and decreasing cloud cover in the regions along the west coast of North America and subtropical eastern Pacific.

The anomalous cloud cover pattern depends on altitude. The shadings in Figs. 2b and 2c show the changes of high clouds and low clouds associated with the PDO, respectively. A positive PDO index is associated with enhanced high cloud cover in the equatorial tropical Pacific and extending south and east, along with decreased high cloud cover over the Maritime Continent and the Indian Ocean. In the North Pacific, the changes of high clouds are marked by a quadrupole structure. High clouds increase in the western side and decrease in the eastern side of the subtropical North Pacific, and vice versa at midlatitudes. Associated with a positive PDO, low cloud cover increases in the north central Pacific and decreases along the west coast of North America and in the tropical eastern Pacific.

The relative contribution of anomalous low clouds and high clouds to the changes of total cloud cover depends on the climatology of the vertical distribution of clouds. Over the Maritime Continent and the tropical and subtropical western Pacific, where convective clouds are prevalent, the changes of total cloud cover are mostly due to changes in high clouds. Conversely, in the midlatitude North Pacific and the stratocumulus zone of the northeast Pacific, where the stratiform clouds are prevalent, the anomalous total cloud cover is dominated by low cloud changes. It is worth noting that the signs of total cloud changes and high cloud changes are robust across models in nearly the whole Pacific and the tropical Indian Ocean, and the sign of low cloud changes is also robust in the regions where the changes of low clouds dominate (Fig. 2).

b. Characteristics of the AMO in model simulations

The AMO represents the coherent decadal variation of the SST anomalies in the North Atlantic, which is

FIG. 2. Regression coefficient of cloud cover on the PDO index (shading) in CMIP5 models: (a) total cloud cover, (b) high cloud cover, and (c) low cloud cover; dots indicate over 80% of models agree in sign. The contours in (b) and (c) represent the omega-related high cloud changes and the SST-related low cloud changes, respectively, obtained by GCM simulations [see Eq. (2) for detailed description]. Black contours indicate positive values, and magenta contours indicate negative values, with a 0.15% per standard deviation contour interval and the zero contour omitted.

suggested to be driven by the variation in the Atlantic meridional overturning circulation (AMOC) and the associated heat transport (Ba et al. 2014; Delworth and Mann 2000; Knight et al. 2005; Latif et al. 2004; Wang and Zhang 2013; Zhang et al. 2016). The positive phase of the AMO is marked by anomalously warm SSTs over the North Atlantic Ocean, which are also in horseshoe-shaped bands (Figs. 1b,f), and anomalously cool SSTs over the South Atlantic. In the Pacific, the patterns are similar to those of PDO albeit with opposite signs. These features are shared in both observation and GCM simulations.

The AMO-related ω_{500} changes in GCMs (Fig. 1h) featured a dipole structure in the tropical Atlantic, with anomalous ascent in the North Atlantic and anomalous descent in the South Atlantic. The dipole structure is accompanied by anomalous descent in the subtropical eastern Pacific of both hemispheres and the equatorial western Pacific. These features of ω_{500} changes in the Pacific are consistent with previous studies that showed how warming in the tropical Atlantic can remotely influence the eastern Pacific through Rossby waves and force anomalous descent in the central-to-western equatorial Pacific through equatorial Kelvin waves (Kucharski et al. 2016b; Li et al. 2016; McGregor et al. 2014). ERA-40 and the Twentieth Century Reanalysis also exhibit consistent large-scale features of the AMO-related ω_{500} anomalies, but the exact locations of anomalous descent in the equatorial western Pacific are different.

The AMO-related changes in cloud cover are demonstrated in Fig. 3. Similar to the changes of ω_{500} , the high cloud cover shows a dipole structure in the tropical Atlantic with increasing high clouds in the North Atlantic and decreasing high clouds in the South Atlantic, coinciding with decreasing high cloud cover in the subtropical eastern Pacific of both hemispheres and tropical central Pacific (Fig. 3b). The anomalously negative low cloud cover in the North Atlantic has a horseshoe shape that is similar to the SST pattern, and the anomalously positive low cloud cover is most significant in subtropical northeast Pacific (Fig. 3c). These characteristics are robust across models (Fig. 3c). However, the changes in high and low cloud cover compensate in most regions, leaving the extratropical North Atlantic and equatorial central Pacific as the only two regions that exhibit consistent total cloud cover changes among models (Fig. 3a).

c. Linking the PDO- and AMO-related cloud changes to meteorological factors

To understand the physics behind the robust PDO- and AMO-related cloud changes in GCM simulations and examine the reliability of GCMs, we estimate the cloud cover changes that are related to changes of SST or circulation associated with the PDO and AMO via regression analysis [Eq. (2)]. The contours in Figs. 2b and 3b show the high cloud changes related to ω_{500} in CMIP5 models. The pattern and magnitudes are similar to the anomalous high cloud cover (shading in Figs. 2b and 3b), suggesting that the PDO- and AMO-related high cloud changes are mostly due to changes in circulation. The area-weighted spatial correlations between the PDO-related high cloud changes and high cloud changes related to vertical motion are above 0.5 across

FIG. 3. Regression coefficient of cloud cover on AMO index (shading) in CMIP5 models: (a) total cloud cover, (b) high cloud cover, and (c) low cloud cover; dots indicate over 80% of models agree in sign. The contours in (b) and (c) represent the omega-related high cloud changes and the SST-related low cloud changes, respectively, obtained by GCM simulations [see Eq. (2) for detailed description]. Black contours indicate positive values, and magenta contours indicate negative values, with a $1\% \text{ K}^{-1}$ contour interval and the zero contour omitted.

all models. In the case of the AMO, the spatial correlations are higher than 0.5 in 26 out of 29 models (Table 2), indicating that the relation between changes of high clouds and circulation holds in most models.

The contours in Figs. 2c and 3c show the SST-related low cloud changes. The pattern and magnitudes are also similar to the low cloud changes associated with the PDO and the AMO (shading in Figs. 2c and 3c), suggesting that the low cloud changes associated with the PDO and the AMO are related to the SST changes. The area-weighted spatial correlations between the PDO-related low cloud changes and low cloud changes related

TABLE 2. The area-weighted spatial correlation between high cloud changes associated with PDO/AMO (shading of Figs. 2b and 3b) and vertical motion–related high cloud changes (contours of Figs. 2b and 3b), and the area-weighted spatial correlation between low cloud changes associated with PDO/AMO (shading of Figs. 2c and 3c) and SST-related low cloud changes (contours of Figs. 2c and 3c). Values higher than 0.5 are shown in bold.

Models	High cloud changes and vertical motion–related changes		Low cloud changes and SST-related changes	
	PDO	AMO	PDO	AMO
ACCESS1.0	0.84	0.77	0.73	0.68
ACCESS1.3	0.75	0.67	0.71	0.73
BCC-CSM1.1	0.67	0.63	0.69	0.62
BCC-CSM1.1m	0.66	0.54	0.78	0.72
BNU-ESM	0.78	0.70	0.60	0.42
CanESM2	0.76	0.72	0.68	0.53
CCSM4	0.74	0.53	0.74	0.49
CESM1-BGC	0.76	0.70	0.78	0.53
CESM1-CAM5	0.70	0.65	0.63	0.54
CMCC-CM	0.70	0.61	0.58	0.51
CSIRO-Mk3.6.0	0.67	0.61	0.73	0.70
FGOALS-g2	0.73	0.65	0.63	0.50
FGOALS-S2	0.78	0.74	0.69	0.75
GFDL CM3	0.63	0.55	0.56	0.68
GFDL-ESM2G	0.72	0.72	0.69	0.70
GFDL-ESM2M	0.58	0.46	0.34	0.31
GISS-E2-H	0.56	0.48	0.50	0.54
GISS-E2-R	0.55	0.59	0.53	0.52
INM-CM4	0.75	0.68	0.58	0.56
IPSL-CM5A-LR	0.59	0.51	0.70	0.72
IPSL-CM5A-MR	0.59	0.54	0.64	0.71
IPSL-CM5B-LR	0.64	0.61	0.74	0.80
MIROC5	0.79	0.45	0.71	0.64
MIROC-ESM	0.70	0.65	0.77	0.81
MPI-ESM-LR	0.71	0.70	0.44	0.44
MPI-ESM-MR	0.61	0.64	0.50	0.57
MPI-ESM-P	0.70	0.69	0.56	0.47
MRI-CGCM3	0.67	0.65	0.58	0.59
NorESM1-M	0.75	0.74	0.66	0.48

to SST are above 0.5 in 27 out of 29 models, and 23 out of 29 models in the case of the AMO (Table 2), showing that the linkage of the PDO- and AMO-related low cloud changes and SST changes hold in most of the models. This result is consistent with previous studies that found significant correlation between low clouds and SST in observational data (Clement et al. 2009; Eastman et al. 2011; McCoy et al. 2017; Myers and Norris 2015) and GCMs (Myers and Norris 2015; Qu et al. 2014, 2015b). The mechanisms of how anomalous SSTs alter the profiles of temperature and moisture and thereby influence low clouds have been investigated in previous studies. For the impacts on temperature profile, a warmer local SST leads to a decrease of EIS, which favors smaller low cloud coverage (Myers and

Norris 2015; Qu et al. 2014, 2015b; Wood and Bretherton 2006). It is worth noting that EIS changes are also influenced by temperature changes in the free atmosphere, which can be caused by nonlocal SST changes (Ceppi and Gregory 2017; Qu et al. 2015a; Zhou et al. 2016, 2017). For the impacts on moisture profile, previous studies suggest that a warmer SST increases the moisture flux from the surface to the boundary layer, and hence increases the moisture gradient between the boundary layer and the free troposphere (Bretherton 2015; Klein et al. 2017; Qu et al. 2015b). The increasing moisture gradient leads to more efficient entrainment drying, favoring reduced low cloud coverage (Bretherton and Blossey 2014; Qu et al. 2015b). As a result, warmer SSTs correspond to fewer low clouds and vice versa.

In summary, the cloud changes associated with the PDO and the AMO can be separated into changes of high clouds and low clouds, with the former being strongly tied to changes in atmospheric circulations and the latter being strongly tied to changes in local SST. In the case of a positive PDO, the warming in the eastern and central equatorial Pacific leads to an eastward shift of the tropical Pacific convective zone and an equatorial shift of the SPCZ, and hence influences the high clouds in the tropical Pacific and the Indian Ocean. On the other hand, the warm SST anomalies in the northeast Pacific and tropics lead to the decreasing low cloud cover in these regions, and the cool SST anomalies in the midlatitudes of the northwest Pacific lead to the increasing low cloud cover. In the case of a positive AMO, the hemispherically asymmetric SST anomalies in the tropical Atlantic lead to the asymmetric changes of vertical motion in the northern and southern Atlantic, which influences high cloud cover. Anomalous ascent in the tropical North Atlantic induces remote descent in the tropical central Pacific and the subtropics and midlatitudes of the eastern Pacific of both hemispheres, suppressing high cloud cover in these regions. Meanwhile, warming in the North Atlantic leads to decreasing low cloud cover. In the subtropical northeast Pacific, the anomalous cool SST and anomalous descent both contribute to the increase of EIS, resulting in increasing low cloud cover. We find that although the magnitude can vary, the patterns of cloud changes associated with the PDO are consistent and robust across models in nearly the whole Pacific and the tropical Indian Ocean. When separated into changes of high clouds and low clouds, the AMO impacts are also robust in the North Atlantic and the eastern Pacific, although the compensation between anomalous low cloud cover and high cloud cover may lead to diverse patterns of anomalous total cloud cover.

4. Estimating the contribution of decadal variability and GHG forcing to the observed cloud trends

In this section, we investigate the role of decadal variability in driving the long-term trend of the observed cloud cover and further compare with the estimated contribution of GHG forcing. Figure 4a shows the trend of total cloud cover in ISCCP observations. During the satellite era, regions with the more significant increase of total cloud cover include the tropical western Pacific, poleward and westward of the SPCZ, and the subtropical eastern Pacific of both hemispheres. Regions with decreasing total cloud cover include the equatorial central Pacific and midlatitudes of both hemispheres. In the Atlantic, the total cloud cover shows a hemispherically asymmetric structure, with decreasing cloud cover in the Northern Hemisphere and increasing cloud cover in the Southern Hemisphere. The trends of cloud cover in the ISCCP and the PATMOS-x datasets have also been discussed in Norris et al. (2016), who have shown that the increasing cloud cover in the tropical western Pacific, the poleward shift of the SPCZ, and the decreasing cloud cover in the midlatitudes of both hemispheres are consistent with trends in albedo and liquid water path in satellite observations.

The trends of high clouds and low clouds are very different, being opposite to each other in many regions. Figures 4b and 4c show the trends of high cloud cover and low cloud cover, respectively. In the tropical western Pacific, there are reducing high clouds in the equatorial region and a poleward shift of high clouds in the region around the SPCZ. The changes of high clouds are offset by changes in low clouds over the equatorial western Pacific. In the subtropical Pacific of both hemispheres, the low clouds increase in the eastern side and decrease in the western side, while the reverse is true for high clouds. In the midlatitudes of the South Pacific, high clouds increase and low clouds decrease. In the tropical Atlantic, high clouds and low clouds both exhibit a dipole structure. The positive high cloud cover trends in the Northern Hemisphere and the negative high cloud cover trends in the Southern Hemisphere are compensated by the opposite low cloud changes. It should be noted that to some extent, the compensation between high and low cloud cover may be attributable to the changes in obscuration, that increasing high clouds mask low clouds from the satellite and decreasing high clouds reveal more previously masked low clouds.

Figure 5 shows the estimated contribution of decadal variability to the observed cloud cover trends [Eq. (3), with the terms $\partial C/\partial \text{PDO}$ and $\partial C/\partial \text{AMO}$ obtained from GCM outputs, and the terms $d\text{PDO}/dt$ and $d\text{AMO}/dt$

FIG. 4. Trend of cloud cover (percentage amount) in ISCCP observations (1983–2009). The plus signs indicate the sign of the trend of PATMOS-x observations in the same period is consistent with ISCCP.

estimated from observational data]. During the satellite era, the PDO transitions from positive to negative phases, whereas the phase of the AMO switches from negative to positive. These gradual phase switches of the PDO and the AMO are accompanied by unique signatures of cloud cover changes that exhibit zonal variations in the Pacific, hemispheric asymmetry in the Atlantic, and a relatively more uniform increase in the tropical Indian Ocean (Figs. 5a,c,e). Many of these cloud pattern changes are consistent with the linear cloud cover trends observed over the satellite data (Figs. 5b,d,f). The estimated contributions of decadal variability to observed high cloud changes are characterized by a poleward and westward shift of the SPCZ, increasing high clouds in the Indian Ocean and the western North Pacific, and decreasing high clouds in the eastern North Pacific. In the tropical Atlantic, high

FIG. 5. (a),(c),(e) The estimated contribution of decadal variability (PDO + AMO) to observed cloud cover trends [Eq. (3), with the terms $\partial C/\partial \text{PDO}$ and $\partial C/\partial \text{AMO}$ obtained from GCM outputs, and the terms $d\text{PDO}/dt$ and $d\text{AMO}/dt$ estimated from observational data]. (b),(d),(f) The pattern consistency between the contribution of decadal variability in (a), (c), and (e) and the observed cloud cover trend in ISCCP (Figs. 4a–c, respectively). Orange indicates the signs are both positive, and blue indicates the signs are both negative. Dots indicate over 80% of models agree in sign. The 60°S–60°N mean trend has been removed.

cloud cover increases in the Northern Hemisphere and decreases in the Southern Hemisphere. The estimated contributions of decadal variability to observed low cloud changes are characterized by increasing low clouds in the northeast Pacific and decreasing low clouds in the northwest Pacific and the North Atlantic. In the Pacific, the estimated trends of total, high, and low cloud cover associated with decadal variability are dominated by the PDO component (Fig. 6). Exceptions are in the subtropics of the eastern Pacific and the tropical central Pacific, where the cloud cover trends associated with the AMO have magnitudes similar to those associated with the PDO (Fig. 7). In the Atlantic, the PDO component vanishes and the AMO component dominates the trends of total, high, and low cloud cover.

To account for potential biases in GCMs, we applied the method described in section 2d [Eq. (4)], in which the estimated contribution of decadal variability to the

observed low cloud cover trends is estimated entirely using observational data (Fig. 8). In the Pacific and the Atlantic, the estimated contribution of decadal variability to low cloud cover captures the features of the spatial pattern of the observed low cloud cover trends reasonably well (spatial correlation $R = 0.48$), although the magnitudes are underestimated. It is worth noting that, although the AMO-related low cloud changes in the South Atlantic are very weak and not robust in CMIP5 models (Fig. 7e), observational data suggest that the hemispherically asymmetric structure of the AMO impacts low clouds in the Atlantic, consistent with the observed low cloud cover trends (Figs. 8e,f). The AMO component dominates the estimated low cloud cover trends in the Atlantic, and also contributes to about half of the estimated trends of low clouds in the subtropical northeast Pacific. In other regions of the Pacific, the estimated low cloud cover trends are dominated by the

FIG. 6. (a),(c),(e) The estimated contribution of the PDO to observed cloud cover trends [Eq. (3)], with the term $\partial C/\partial \text{PDO}$ obtained from GCM outputs, and the term $d\text{PDO}/dt$ estimated from observational data]. (b),(d),(f) The pattern consistency between the PDO contribution in (a), (c), and (e) and the observed cloud cover trend in ISCCP (Figs. 4a–c, respectively). Orange indicates the signs are both positive, and blue indicates the signs are both negative. Dots indicate over 80% of models agree in sign. The 60°S–60°N mean trend has been removed.

PDO component with negligible contribution from the AMO component.

We also compare the estimated contribution of decadal variability to the observed cloud cover trends with the estimated contribution of GHG forcing (Fig. 9). The GHG-induced total cloud cover trends, in contrast to those associated with the PDO and the AMO, are more zonally uniform in the Pacific and are more hemispherically symmetric in the Atlantic. For high cloud cover, there are positive trends in the entire equatorial tropical Pacific region, negative trends in the tropical Atlantic in both hemispheres, and zonally uniform positive trends in poleward of 50° in both hemispheres. For low cloud cover, a westward and poleward expansion of the stratocumulus zones occurs in the subtropical southeast Pacific and the subtropics of the Atlantic in both hemispheres, while in the midlatitude North Pacific the low cloud cover decreases. These GHG-induced cloud cover trends may partially explain the observed

cloud cover trends in specific regions, including positive low cloud cover trends in the subtropical southeast Pacific, negative total cloud cover trends in the mid-latitudes, and positive high cloud cover trends poleward of 50°N/S.

Comparing Fig. 5 and Fig. 9, the multimodel mean of the estimated decadal variability contributions to the observed cloud cover trends are larger than those due to GHG-induced warming. This is also the case in most of the models when assessing them individually. We examine the absolute value of the ratio of the estimated decadal variability contribution [left-hand side of Eq. (4)] to the GHG contribution [left-hand side of Eq. (5)] at each grid point in each model. A ratio larger than 1 indicates that at the grid point, the joint impacts of the PDO and the AMO variation on cloud cover trend are larger than the impacts of GHG-induced warming during 1983–2009; on the other hand, a ratio smaller than 1 indicates that the GHG forcing is more

FIG. 7. (a),(c),(e) The estimated contribution of the AMO to observed cloud cover trends [Eq. (3), with the term $\partial C/\partial \text{AMO}$ obtained from GCM outputs, and the term $d\text{AMO}/dt$ estimated from observational data]. (b),(d),(f) The pattern consistency between the PDO contribution in (a), (c), and (e) and the observed cloud cover trend in ISCCP (Figs. 4a–c, respectively). Orange indicates the signs are both positive, and blue indicates the signs are both negative. Dots indicate over 80% of models agree in sign. The 60°S–60°N mean trend has been removed.

important. In nearly all the regions over the globe, in over 50% of the models the ratios fall within the range of 1–10 (Fig. 10a), indicating that the contribution of decadal variability to observed total cloud cover trend is larger than GHG contribution, while the two are still of the same order of magnitude and the GHG contribution is not negligible. About 30% of the models show a ratio larger than 10 in the regions of the North Pacific, the southwest Pacific, the North Atlantic, and the Indian Ocean (Fig. 10b), suggesting a dominating decadal variability contribution and a negligible GHG contribution in these regions. In the subtropical southeast Pacific, about 20% of the models show a ratio smaller than 1, suggesting that the impacts of GHG forcing may be important in that region. Overall, in the majority of the models, the contribution of decadal variability to observed cloud cover trends is more important, while the GHG contribution may be smaller but not negligible. As the observational records become longer, the

contribution of decadal variability to the observed linear trend is expected to decrease while the relative importance of GHG forcing is expected to increase.

The area-weighted spatial correlations between the observed cloud cover trends and contributions from the PDO and the AMO, the joint impacts of the PDO and the AMO, and GHG forcing are summarized in Table 3. The pattern of decadal variability-related total cloud cover trends is significantly correlated with those observed by the ISCCP ($R = 0.37$) and the combined ISCCP/PATMOS-x dataset ($R = 0.32$). However, the correlations between the patterns of decadal variability-related cloud cover trends and those observed by the PATMOS-x are weak and not significant, reflecting the large uncertainty in observed clouds trends. For the high cloud cover trends, the spatial correlations between the pattern of decadal variability-related high cloud cover trends and those in the ISCCP and the PATMOS-x observation are not significant, which may be caused by

FIG. 8. (a),(c),(e) The estimated contribution of decadal variability (PDO + AMO), the PDO, and the AMO to observed low cloud cover trends, respectively [Eq. (4), with all the terms estimated from observational data]. (b),(d),(f) The pattern consistency between the contribution of decadal variability in (a), (c), and (e) and the observed cloud cover trend in ISCCP (Figs. 4a–c, respectively). Orange indicates the signs are both positive, and blue indicates the signs are both negative. The 60°S–60°N mean trend has been removed.

the small number of spatial degrees of freedom of the high cloud data (see Table S1 in the online supplemental material). The pattern of low cloud cover trends associated with decadal variability is significantly correlated with those observed by the ISCCP ($R = 0.38$). Both the PDO and the AMO components show similar significance as in the case of the joint impacts of decadal variability, suggesting the correlations are contributed by both climate modes rather than dominated by either mode. The estimated GHG contribution shows no significant spatial correlation in any case, suggesting a relatively minor role for GHG-induced cloud changes on observed cloud trends.

Although we highlight in this study the importance of decadal variability to the cloud cover trends in the satellite era, the impacts of external forcings may not be ignored. Norris et al. (2016) found a spatial correlation of 0.39 between the cloud cover trends in the period of 1983–2009 in the combined ISCCP/PATMOS-x dataset

and that in the coincident CMIP5 historical simulations that include GHG, aerosol, ozone, and volcano forcings. The correlation of 0.39 in the case of historical simulations is higher than the correlation in the case of GHG single-forcing simulations ($R = 0.21$), suggesting that the external forcings other than GHG, which are not discussed in the present study, may also contribute to the observed cloud cover trend. Norris et al. (2016) also note that a correlation as high as 0.39 is extremely rare when considering all possible 27-yr-long chunks of 15 000 years of preindustrial simulations, suggesting that the observed trends are at least partly forced and were extremely unlikely to have occurred due solely to internal climate variability. Consistently, in our analysis, the spatial correlation between the cloud cover trends associated with decadal variability and the observed cloud cover trends in the combined ISCCP/PATMOS-x dataset ($R = 0.32$) is slightly lower than the spatial correlation between cloud cover trends in CMIP5

FIG. 9. (a),(c),(e) The estimated contribution of GHG forcing to observed cloud cover trends [Eq. (5), with the term $dC/dGMT$ obtained from GCM outputs, and the term $dGMT/dt$ estimated from observational data]. (b),(d),(f) The pattern consistency between GHG contribution in (a), (c), and (e) and the observed cloud cover trend in ISCCP (Figs. 4a–c, respectively). Orange indicates the signs are both positive, and blue indicates the signs are both negative. Dots indicate over 80% of models agree in sign. The 60°S–60°N mean trend has been removed.

historical simulations and observation found in Norris et al. (2016) ($R = 0.39$), suggesting that the joint impacts of external forcings may also be important to the pattern of observed cloud cover trends. Nevertheless, the spatial correlation between the observed cloud cover trends and decadal variability contribution ($R = 0.32$) is larger than the correlation between the observed cloud cover trends and the contribution of any single external forcing [comparing with Table 1 in Norris et al. (2016)]. By considering the specific changing rates of the PDO and the AMO indices, the present study suggests that the observed cloud cover trends are more closely linked with decadal variability than GHG forcing. Some previous studies suggest that external forcings may contribute to part of the patterns and strength of the PDO and the AMO through their impacts on regional SST (Booth et al. 2012; McGregor et al. 2014; Takahashi and Watanabe 2016; Zhang et al. 2013). While we cannot exclude the possibility of external forcings contributing

to the PDO- and AMO-like patterns of SST anomalies and hence influencing the trend of the PDO and the AMO indices in observation, our results highlight the contributions of the internally generated decadal variability to the observed cloud cover trends in the majority of the Pacific and North Atlantic via investigating unforced preindustrial simulations.

In summary, the estimated contribution of decadal variability to the cloud cover trends shows a similar pattern and order of magnitude with the observed cloud cover trends. The spatial patterns of the contribution of decadal variability to total cloud cover trends are significantly correlated with the cloud cover trends in the ISCCP and the combined ISCCP/PATMOS-x datasets. In the Pacific, the PDO component dominates the zonal variation cloud cover trends associated with decadal variability, except in the subtropical eastern Pacific and in the tropical central Pacific where the AMO can also contribute to the high cloud cover trends. In the

FIG. 10. Percentage of models in which absolute values of the ratio of decadal variability contribution (PDO + AMO) to GHG contribution are (a) within the range of 1–10, (b) larger than 10, and (c) smaller than 1.

Atlantic, the AMO component dominates the hemispherically asymmetric cloud cover trends associated with decadal variability. The estimated GHG contribution to total cloud cover trends shows a more zonally uniform pattern in the Pacific as well as a more

hemispheric symmetric pattern in the Atlantic. This pattern differs markedly from the cloud cover trends associated with the PDO and the AMO, and shows no significant correlation with the pattern of total cloud cover trends in observation. When compared with the contribution of GHG forcing, the contribution of decadal variability shows larger magnitudes in nearly all the regions over the globe. It has been discussed in Seethala et al. (2015) that the trends of low cloud cover in the five main stratocumulus zones during the period of 1983–2009 may be associated with natural variability rather than GHG-induced warming, as the low cloud cover trends and their relative contributions of meteorological factors are more diverse across regions compared to those in the future warming experiments simulated by climate models (Qu et al. 2014). Consistent with their results, the present study further suggests that the phase shift of the PDO and the AMO during the period of 1983–2009 has significantly contributed to the long-term trends of the observed cloud cover.

5. Summary

The relationships between clouds and the PDO and the AMO are investigated in GCM simulations. The patterns of PDO-related total cloud cover changes feature an eastward and equatorial shift of clouds in the tropical western Pacific and the Indian Ocean, along with zonally asymmetric total cloud cover changes in the North Pacific. On the other hand, patterns of AMO-related cloud cover changes are characterized by hemispherically asymmetric anomalies of high and low cloud cover in the Atlantic, as well as decreasing high cloud cover in the subtropical eastern Pacific and the tropical central Pacific. These PDO- and AMO-related changes of cloud cover can be linked to the anomalous SST and circulation. Regions with anomalous ascent experience anomalously greater high cloud fraction and vice versa; regions with anomalously cool SST tend to experience increases in low cloud fraction and vice versa. Although

TABLE 3. The area-weighted spatial correlation between the estimated decadal variability/GHG contribution (Figs. 5–7 and 9) and the observed cloud trend (Fig. 4). Parentheses show the p values obtained from a one-tailed Student's t test; values with p value < 0.05 are highlighted in bold. The effective numbers of spatial degrees of freedom used in the statistical test are listed in Table S1.

		PDO	AMO	PDO + AMO	GHG
ISCCP	Total clouds	0.33 (0.009)	0.24 (0.04)	0.37 (0.004)	0.13 (0.19)
	High clouds	0.24 (0.13)	0.28 (0.09)	0.35 (0.05)	−0.04 (0.58)
	Low clouds	0.25 (0.02)	0.30 (0.04)	0.38 (0.00)	0.00 (0.50)
PATMOS-x	Total clouds	0.03 (0.42)	0.17 (0.10)	0.12 (0.18)	0.16 (0.11)
	High clouds	0.16 (0.18)	0.26 (0.06)	0.27 (0.05)	−0.04 (0.60)
	Low clouds	0.13 (0.11)	0.07 (0.25)	0.13 (0.10)	0.08 (0.22)
Combined ISCCP/PATMOS-x		0.24 (0.04)	0.25 (0.04)	0.32 (0.01)	0.15 (0.15)

the method of linearly estimating the SST- or vertical motion-related cloud changes [Eq. (2)] is highly simplified, and may miss some nonlocal impacts, it still captures the pattern and magnitude of anomalous high and low cloud coverage associated with the PDO and the AMO.

When compared with the observed cloud cover trends during the satellite era, the estimated contribution of decadal variability shows a similar pattern and order of magnitude to the observed cloud cover trends. The spatial pattern of the decadal variability-related total and low cloud cover trends are significantly correlated with the cloud cover trends in ISCCP, while for the PATMOS-x dataset the correlation is weak and insignificant. Nevertheless, both datasets share some features that are consistent with the signature of decadal variability impacts, including the poleward shift of the SPCZ, decreasing clouds in the equatorial central Pacific, the zonally asymmetric structure in the North Pacific, and the hemispherically asymmetric anomalous low cloud cover in the Atlantic. In the majority of the models, the estimated contribution of decadal variability to cloud cover trends is larger than that from GHG forcing in nearly all regions over the globe. These results suggest that the phase shifts of the PDO and the AMO around the year 2000 have significantly contributed to the cloud cover trends in the satellite record.

Our results show that the impacts of the PDO, the AMO, and GHG forcing on clouds are regionally dependent and have different signatures. The distinct patterns may provide hints for attributing the observed cloud cover trend. To the extent that the observed cloud cover trends have a large component that does not arise from GHG-induced global warming, our results also suggest that the cloud feedback obtained from natural variability may differ from that obtained from global warming scenarios, since the associated cloud cover patterns are very different. Caution should be taken when comparing the cloud feedback obtained from observational data, in which forced and internally generated responses are comparable, with the cloud feedback obtained from warming experiments in GCM, in which the forced signal dominates over the noise of natural variability by design (Hawkins and Sutton 2009, 2011). Further attribution work that isolates the forced cloud responses from the impacts of internal variability may be needed to yield an observational constraint for climate models.

Acknowledgments. Y.-J. Chen and Y.-T. Hwang were supported by the Ministry of Science and Technology of Taiwan (105-2628-M-002-009-MY4, 107-2636-M-002-001-, and 108-2636-M-002-007-). MDZ's work was

supported by DOE's Regional and Global Model Analysis Program under the project "Identifying Robust Cloud Feedbacks in Observations and Models" and was performed under the auspices of the United States Department of Energy by Lawrence Livermore National Laboratory under contract DE-AC52-07NA27344. CZ's work was supported by NSFC (41875095). We acknowledge the World Climate Research Programme's Working Group on Coupled Modelling, which is responsible for CMIP, and we thank the climate modeling groups (listed in Table 1) for producing and making available their model output. For CMIP, the U.S. DOE's Program for Climate Model Diagnosis and Intercomparison provided coordinating support and led development of software infrastructure in partnership with the Global Organization for Earth System Science Portals. We thank Joel Norris for providing artifact-corrected ISCCP and PATMOS-x data. NOAA_ER SST_V5 data and 20th Century Reanalysis V2c data provided by the NOAA/OAR/ESRL PSD, Boulder, Colorado, USA, from their Web site at <https://www.esrl.noaa.gov/psd/>. We thank three anonymous reviewers for their constructive comments.

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