Unforced interannual to decadal variability of global radiation imbalance: Role of low clouds

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Submitted to Journal of Climate on June 7th, 2025

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ABSTRACT: The global radiation imbalance at the top of the atmosphere (TOA) is an important indicator of the climate response to anthropogenic greenhouse forcing. Natural variability perturbs this radiation imbalance on interannual and decadal timescales, confounding the externally forced signal. However, limited observations hinder the effort to understand the mechanisms for internally generated radiation imbalance. This study investigates the natural variability of global TOA radiation using a 500-year preindustrial coupled simulation with the Community Earth System Model version 2, and a corresponding atmospheric model simulation forced with daily sea surface temperature (SST) and sea ice from the coupled run. Variations in global TOA radiation lead those in tropical Pacific SST and global-mean surface temperature by 90° in phase. The analysis reveals the dominant role of low-cloud radiative effects with timescale-dependent spatial patterns. On interannual timescales, low cloud anomalies are distributed across tropical and extratropical oceans, with maxima over the equatorial northeast Pacific. In contrast, decadal variability of global TOA radiation is due to variations of eastern subtropical low cloud decks, coupled with underlying SST anomalies. These low cloud-SST co-variations are triggered by stochastic extratropical atmospheric variability. This timescale dependence likely reflects the characteristics of these drivers: the amplitude of El Niño-Southern Oscillation peaks at interannual timescales due to tropical ocean dynamics, whereas extratropical stochastic forcing becomes increasingly important on decadal and longer timescales. Recent satellite observations of TOA radiation corroborate both mechanisms. This study underscores the importance of subtropical low cloud-SST co-variations induced by extratropical atmospheric forcing in unforced variability of global energy imbalance.

1. Introduction

The Earth's energy budget is a fundamental physical property of the climate system, and governs the rise in global-mean surface temperature due to greenhouse gas forcing. The climate system responds to radiative forcing (F) at the top of the atmosphere (TOA) by modifying longwave emission and incoming solar radiation via changes in surface temperature. This global-mean energy budget may be cast as

$$N = F + \lambda T \tag{1}$$

where N is the net radiation (hereafter expressed as GMTOA; positive value for downward flux) and λ the climate feedback parameter which characterizes radiative feedback from surface temperature perturbation T (Gregory et al. 2004). Reducing uncertainty in future projections of global warming is a pressing task for the climate research community. Radiative feedback estimates from historical changes $(N - F)/T$ have been extensively investigated to improve our understanding of and constraints on future warming (Sherwood et al. 2020).

In this context, decadal changes in GMTOA garner considerable attention. Satellite observations from Clouds and the Earth's Radiant Energy System (CERES) have yielded a continuous record of GMTOA for more than two decades (Loeb et al. 2024). The data record features a striking positive trend in planetary energy uptake, exceeding that simulated by global climate models (Raghuraman et al 2021; Olonscheck and Rugenstein 2023). GMTOA and associated radiative feedbacks in the historical period are often estimated using the AMIP (Atmospheric Model Intercomparison Project) protocol, which refers to atmospheric general circulation models (AGCMs) forced with observed sea surface temperature (SST) and sea ice. Substantial decadal variations in radiative feedbacks have been identified through the model simulations (Gregory and Andrews 2016; Andrews et al. 2018).

Variations in anthropogenic forcing (e.g., CO_2, aerosols) cause evolving SST patterns, which in turn affect GMTOA (Senior and Mitchell 2000; Dong et al. 2020; Andrews et al. 2022). Natural variability plays a significant role in the temporal variations of GMTOA, confounding radiative feedback estimates (Dessler et al. 2018; Wills et al. 2021). This underscores the importance of understanding unforced energy fluctuations.

Natural variability of the global energy budget is distinct from anthropogenic changes. Unlike the forced response, the concurrent correlation between natural variations of GMTOA and GMST is nearly zero, and the peak correlation occurs with GMTOA leading by 90° in phase as required by planetary energy budget (Xie et al. 2016). This peak lagged correlation is modest on decadal timescales (Xie et al. 2016), indicating their complex relationship. The unforced GMST changes are mainly controlled by equatorial Pacific SST anomalies and high-latitude intrinsic atmospheric dynamics (Kosaka and Xie 2013; Xie et al. 2025; Deser et al. 2017).

This raises an important question: what causes natural decadal GMTOA variations? While changes in low clouds induced by tropical convection have recently been highlighted (e.g., Zhou et al. 2016), GMTOA is almost in quadrature with tropical Pacific SST and GMST (Xie et al. 2016). Subtropical-extratropical low clouds form positive feedback with underlying SST (low cloud-SST feedback; Norris and Leovy 1994; Clement et al. 2009; Yang et al. 2023). The low cloud-SST feedback can operate without the tropical Pacific effect (Larson et al. 2024; Miyamoto and Xie 2025). This suggests that the subtropical marine low clouds may play an important role in GMTOA variations.

The present study investigates unforced interannual to decadal variability of GMTOA based on a 500-year simulation under constant preindustrial radiative forcing with a state-of-the-art global climate model, Community Earth System Model version 2 (CESM2; Danabasoglu et al. 2020), and a corresponding "perfect-model" AMIP simulation forced with daily SST and sea ice from the coupled run. Unlike previous studies that examined GMTOA variations associated with GMST or selected SST modes (e.g., Xie et al. 2016; Wills et al. 2021), this study focuses on GMTOA itself without making any *a priori* assumptions about a relationship with SST modes. Our GMTOA-centric analysis reveals the regions and radiative components (cloud and clear-sky contributions) that are crucial in unforced GMTOA variability. The perfect AMIP run is utilized to disentangle stochastic atmospheric forcing and SST effects on the GMTOA variations. A comparison of interannual and decadal GMTOA variations aids the interpretation of the short CERES record.

The rest of the paper is organized as follows. Section 2 describes the data used in this study. Sections 3 and 4 document the natural GMTOA variability in CESM2 on interannual and decadal timescales, respectively. Section 5 compares with the recent CERES observations and discusses the timescale dependence. Section 6 concludes the paper with a summary of the key findings.

2. Data

a. Preindustrial simulations

We use a 500-year fully coupled CESM2 simulation with constant 1850-level radiative forcing (Danabasoglu et al. 2020; hereafter labeled CESM). Its atmosphere and ocean resolutions are nominally 1° in the horizontal with increasing meridional ocean resolution toward the equator. CESM2 simulates natural variability such as El Niño-Southern Oscillation (ENSO) and tropical Pacific decadal variability (TPDV) (Danabasoglu et al. 2020; Capotondi et al. 2020). The model also reproduces subtropical low clouds off the west coasts of continents (Fig. 1) and their positive feedback with underlying SST (Kang et al. 2023; Larson et al. 2024; Miyamoto and Xie 2025).

In parallel with the coupled run, a single-member AMIP simulation with identical boundary conditions (hereafter labeled CAM) was conducted by the CESM2 Climate Variability and Change Working Group (CVCWG). In this setup, the atmospheric component of CESM2, the Community Atmosphere Model version 6 (CAM6), is forced with daily SST and sea ice from CESM. This perfect-model/SST framework allows for a direct comparison with CESM (in a statistical sense owing to the limited ensemble size). If CAM fails to reproduce an anomaly of cloud, temperature, and wind in CESM, the anomaly can be attributed to stochastic atmospheric variability.

b. Observational data

For observational datasets, we use the Optimum Interpolation Sea Surface Temperature (OISST) version 2 (Huang et al. 2021) for SST, the CERES Energy Balanced and Filled edition 4.2 (Loeb et al. 2018) for radiative fluxes, the Moderate Resolution Imaging Spectroradiometer (MODIS) onboard Terra collection 6.1 (Platnick et al. 2003) for cloud cover, and the ERA5 global atmospheric reanalysis (Hersbach et al. 2020) for other meteorological variables. The horizontal resolution of the datasets is 0.25° for OISST and 1° for the others. For the MODIS low cloud cover, the random overlap assumption is applied to suppress the shielding effect of high clouds, as in Miyamoto and Xie (2025).

c. AMIP simulations with observed SST

In parallel with the CERES observations, we analyze two versions of AMIP simulations prescribed with observed SST and sea ice; one conducted by CVCWG uses CAM6 and the other uses the Geophysical Fluid Dynamics Laboratory Atmospheric Model version 4 (AM4; Zhao et al. 2018). Hereafter, we refer to them as CAMobs and AMobs, respectively. The resolution of AM4 is approximately 100 km with 33 levels in the vertical. CAMobs covers the period 1880-2021, while AM4obs covers 1982-2021. Both models are radiatively forced by historical forcing up to 2014 and Shared Socioeconomic Pathway scenarios (SSP3-7.0 for CAMobs and SSP2-4.5 for AMobs) from the Coupled Model Intercomparison Project phase 6 (Eyring et al. 2016). CAMobs uses the monthly-mean SST and sea ice from Extended Reconstructed SST version 5 (Huang et al. 2017) and OISST version 2, respectively, whereas AMobs uses daily-mean OISST version 2 for both SST and sea ice. Both models have 10 ensemble members each, and their ensemble averages are analyzed.

d. Preprocessing

All data are interpolated onto a 2.5° grid, linearly detrended, and smoothed with a 12-month running mean. To decompose the anomalies into interannual and decadal components, a Lanczos filter with a cutoff period of 10 years was applied to the 500-year CESM and CAM simulations. For the decadal components, only calendar-year (January-to-December) 12-month averages are analyzed. The observational data and AMIP simulations are analyzed over the period from 2001 to 2021. The Lanczos time filtering was not applied because of the short observational record.

e. Statistical test

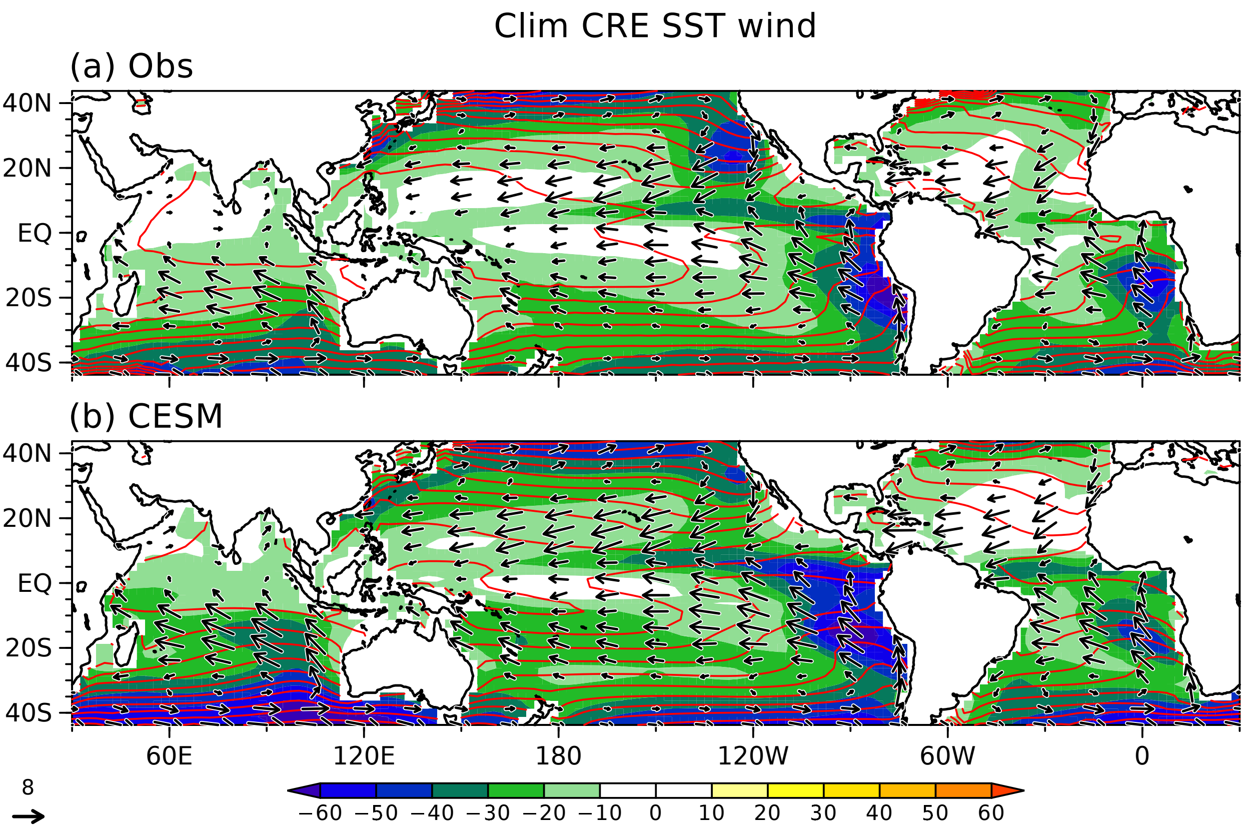
We assess the statistical significance of correlation and regression coefficients using a Student's t-test. To estimate the effective sample size, we calculate effective decorrelation time T_e following Metz (1991):

$$T_e = 1 + 2\sum_{\tau=1}^{L}\left(1 - \frac{\tau}{L}\right) R_{XX}(\tau) R_{YY}(\tau). \tag{2}$$

$R_{XX}(\tau)$ and $R_{YY}(\tau)$ denote autocorrelation functions of variables X and Y at a lag of τ months/years. L is set to 120 months for interannual anomalies and 50 years for decadal anomalies. The effective sample size N_e is then given by

$$N_e = \frac{N}{T_e} \tag{3}$$

where N is the number of samples.



138 FIG. 1. Annual-mean climatology of SST (contoured for every 2 °C), surface wind (arrows; m s^{-1}), and TOA
139 cloud radiative effect (CRE; shading; W m^{-2}; positive values for heating) from (a) observations and (b) CESM.

3. Interannual variation

a. Lead-lag relationship

The correlation analysis indicates that GMTOA and GMST are in quadrature for both interannual and decadal variability in CESM (Figs. 2, top two rows). Peak GMTOA corresponds to a positive time tendency of GMST (Figs. 2f,g), implying that TOA energy uptake is used to increase GMST as expected from energetics for unforced variability (Xie et al. 2016; Lutsko and Takahashi 2018; Proistosescu et al. 2018). This contrasts with forced climate change where radiative feedback is assumed to be proportional to GMST as in Eq. (1). Indeed, the upward clear-sky radiative flux at TOA is nearly in phase with GMST (Fig. 2i), but the GMTOA is dominated by and closely tracks cloud radiative effect (CRE; Fig. 2h). For interannual variability, approximately 75% of the GMTOA peak can be attributed to CRE, particularly its shortwave component, while longwave clear-sky flux is secondary (Table 1). The weak concurrent anomalies in GMST and clear-sky fluxes suggest that natural GMTOA variability is not primarily driven by longwave damping on temperature variability (the Planck response or lapse rate feedback). This study therefore focuses on the cloud processes that create GMTOA anomalies.

Coupled ocean-atmosphere variability and atmospheric stochastic forcing can dictate unforced GMTOA variability (Xie et al. 2016; Lutsko and Takahashi 2018; Proistosescu et al. 2018). Red lines in Figs. 2f-i indicate lag correlation of GMTOA between CESM and CAM, which assesses the SST effect on GMTOA and its lagged relationship in CESM. The autocorrelation of GMTOA in CESM is captured by CAM without significant lead-lag asymmetry (Figs. 2a,b). The correlation at lag 0 amounts to 0.71. This reflects the ability of CAM to capture the evolution of both CRE and clear-sky flux (Figs. 2h,i). This result supports the importance of the ocean effect in the GMTOA variations.

b. Spatial pattern

Figure 3 shows the time evolution of interannual radiation and surface temperature anomalies regressed onto GMTOA in CESM. At the GMTOA peak, positive net TOA anomalies are distributed across the tropics and part of the extratropical oceans (Fig. 3c). The strongest signal appears in the equatorial eastern Pacific with a secondary maximum in the Southeast Pacific. Over the Pacific, SST anomalies at lag 0 do not resemble any well-known SST modes of variability (Fig. 3d), but

Table 1. Global-mean TOA radiative flux (W m^{-2}) regressed onto GMTOA from CESM decadal, interannual, and CERES anomalies (left to right). In CESM decadal anomalies, values in parentheses indicate a +1-year lagged anomaly. SW and LW signify shortwave and longwave components, respectively. CLR denotes clear-sky flux. Boldface indicates statistical significance at the 90% confidence level.

	CESM decadal	CESM interannual	CERES
GMTOA	**0.12** (**0.11**)	**0.40**	**0.26**
Net CRE	**0.09** (**0.11**)	**0.29**	**0.19**
SW CRE	**0.09** (**0.11**)	**0.26**	**0.16**
LW CRE	0.00 (0.00)	**0.04**	0.02
Net CLR	**0.02** (0.00)	**0.11**	**0.07**
SW CLR	0.00 (0.01)	**0.03**	−0.02
LW CLR	0.02 (−0.01)	**0.08**	**0.09**

the lead-lag regression implies the effect of ENSO. In particular, the GMTOA peak is preceded by a La Niña signature (Fig. 3b) and followed by an El Niño pattern (Fig. 3f), consistent with previous studies (e.g., Lutsko and Takahashi 2018; Wills et al. 2021; Tsuchida et al. 2023). This link with the ENSO transition is confirmed by the high correlation (~ 0.7) between GMTOA and Niño3.4 SST (5°S-5°N, 170°W-120°W) at lag ±9 month (Fig. 2j). At these lags, GMST is near its peak (Fig. 2g) consistent with the ENSO pacemaker effect on GMST (Kosaka and Xie 2013), while GMTOA is almost zero (Fig. 2f) due to offsetting anomalies over the Pacific and Indian Ocean (Figs. 3a,e).

c. Mechanism

The dominance of shortwave CRE in the GMTOA variations implies that low-level clouds play a crucial role through their albedo effect (Klein and Hartmann 1993). Figure 4 shows anomalies in low cloud cover and environmental controlling factors associated with interannual GMTOA variations in CESM and CAM. Negative anomalies in low cloud cover are distributed across the tropical and extratropical oceans, leading to the increased shortwave energy uptake (Fig. 4a). This low-cloud decrease explains the TOA radiation changes well, with a pattern correlation of -0.83 between low cloud cover and net TOA radiation.

Aligned with the net radiation changes, the decrease in low clouds peaks over the equatorial eastern Pacific. As shown in its magnified figure, the peak decrease occurs not along the equator but slightly to its north (Fig. 5b). Climatologically, the northward flow toward the intertropical convergence zone crosses a sharp SST gradient on the northern flank of the equatorial cold tongue, promoting stratus and shallow cumulus clouds there (Fig. 5a; Deser and Wallace 1990; Small et al. 2005). Consistent with this climatological-mean state, the changes of equatorial low clouds are attendant with the emergence of warmer cold tongue (Fig. 5e) in the developing El Niño (Fig. 3). This causes anomalous warm-air advection to the north of the equator, which corresponds well with the decrease in low clouds (Fig. 5d). Additionally, lower-tropospheric stability measured by Estimated Inversion Strength (EIS; Wood and Bretherton 2006) also contributes to the low cloud decrease, but the EIS weakening is concentrated along the equator associated with the warming of cold tongue (Fig. 5c). Thus, we highlight that low clouds over the equatorial eastern Pacific play a crucial role in interannual GMTOA variations through the interplay of mean winds and developing El Niño.

Meanwhile, the weakening of lower tropospheric stability occurs across the tropics (Fig. 4a) due to zonally uniform tropical tropospheric cooling (Fig. 4g), a mechanism often invoked in previous literature (Zhou et al. 2016; Fueglistaler 2019; Ceppi and Fueglistaler 2021). The pan-tropical tropospheric temperature follows the moist adiabat set by SST in tropical ascent regions (Sobel et al. 2001). Consistently, negative SST anomalies are distributed in the tropical Indo-Pacific and western Atlantic (Fig. 4i), which persist after a La Niña (Figs. 3b,d,f,h; Enfield and Mayer 1997; Xie et al. 2009). Additionally, the pan-tropical cooling may induce a subsidiary positive effect on GMTOA through a reduction in clear-sky outgoing longwave emission (Fig. S1b; Andrews and Webb 2018; Ceppi and Fueglistaler 2021). We emphasize, however, that the spatial pattern of low cloud and EIS anomalies does not align with a simple picture of the free-tropospheric warming effect on the eastern subtropical low cloud decks (e.g., Zhou et al. 2016). For instance, there are weak increases in low clouds off the Californian coast and strong decreases over the western-to-central extratropical Pacific (Fig. 4a). This discrepancy is attributed to SST anomalies and resultant local stability changes (Figs. 4c,i). The North Pacific SST anomalies are likely due to the lingering effect of the Aleutian low weakening caused by the preceding La Niña (Fig. 3b; Alexander et al. 2002; Yang et al. 2023). Thus, the analysis indicates the effect of decaying La Niña as well as developing El Niño discussed in the previous paragraph. Similar SST and radiation patterns appear in the regression analysis against Niño3.4 SST (Fig. S2).

In Fig. 4, corresponding anomalies in CAM are juxtaposed with the CESM results to examine the SST effect. CAM well reproduces the CESM pattern of low cloud anomalies despite its complexity (Fig. 4b). Combined with prescribed SST and simulated tropospheric temperature (Fig. 4h), CAM accurately captures the EIS anomalies (Fig. 4d). The anomalous warm advection due to warm SST anomalies in the eastern equatorial Pacific is also reproduced (Fig. 4f). These results corroborate the effect of SST anomalies (and presumably ENSO) on low clouds and therefore GMTOA.

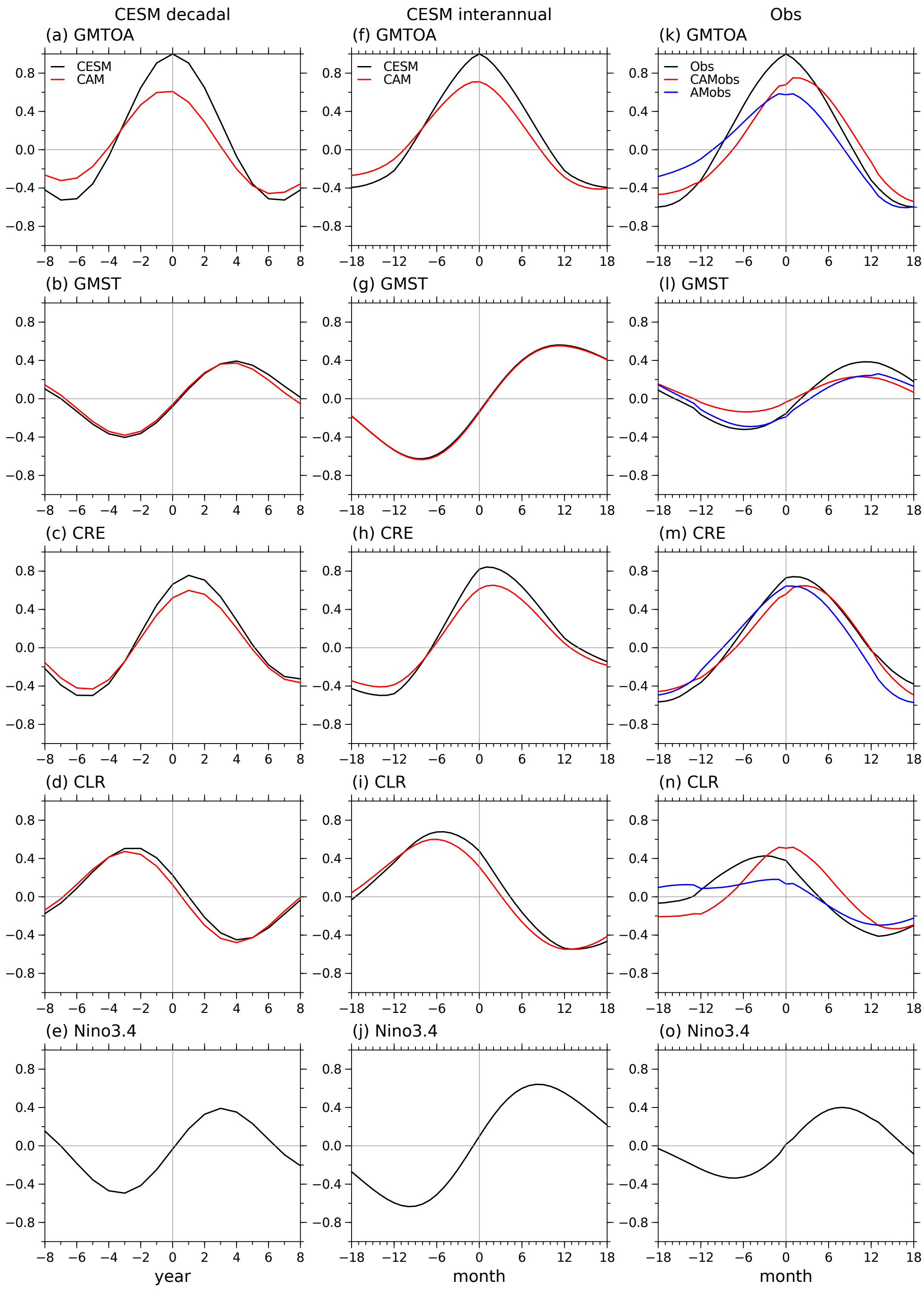


FIG. 2. Lead-lag correlation with GMTOA (positive values for heating) from (a-e) CESM decadal, (f-j) CESM interannual, and (k-o) observed anomalies (black lines). Colored lines denote the correlation of AMIP runs with CESM/CERES GMTOA. (a,f,k) GMTOA, (b,g,l) GMST, (c,h,m) global-mean CRE, (d,i,n) global-mean clear-sky flux, and (e,j,o) Niño3.4 SST. Note that a positive lag means CESM/CERES GMTOA leads.

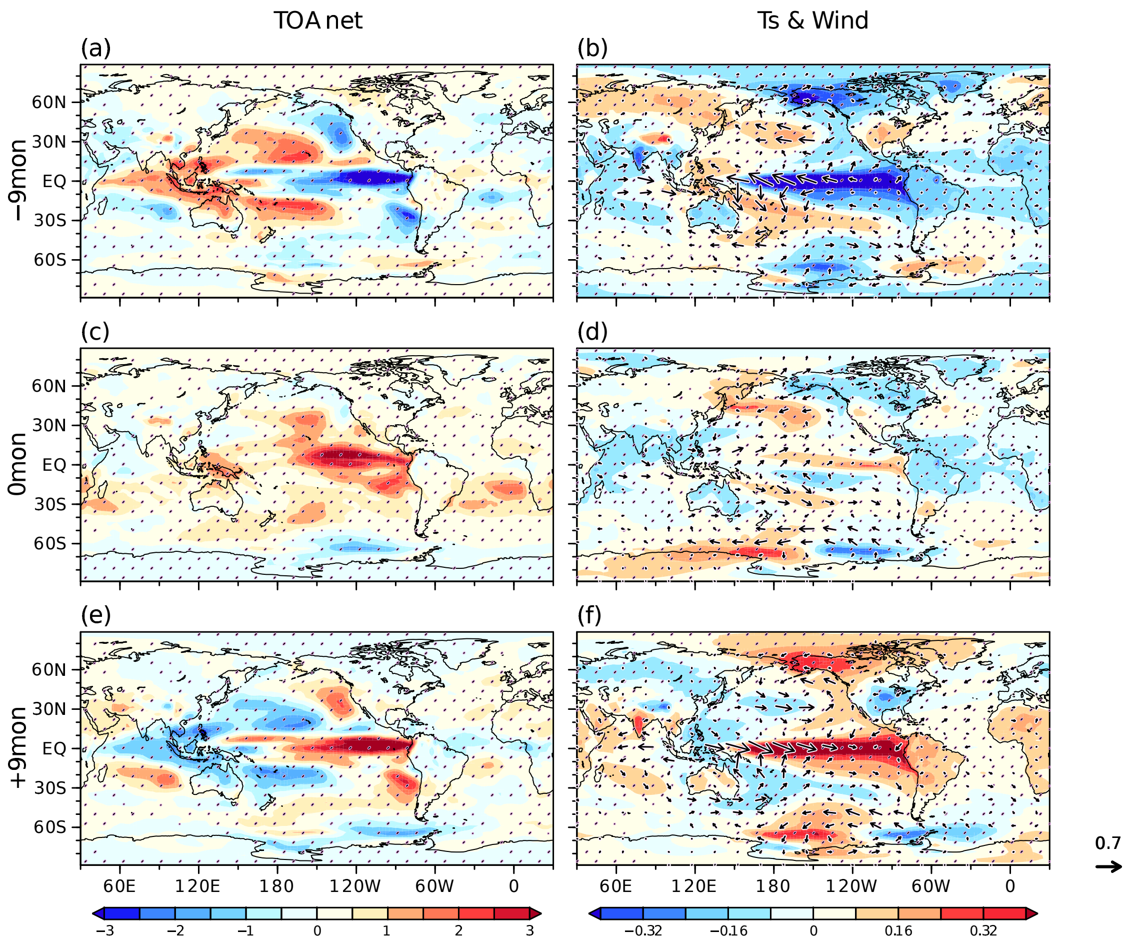


FIG. 3. Spatial pattern and time evolution of interannual GMTOA anomalies in CESM. Lagged regression maps of (a,c,e) TOA radiation (W m^{-2}), (b,d,f) surface temperature (shading; °C) and wind (arrows; m s^{-1}; only points with the 90% confidence are drawn) onto CESM interannual GMTOA anomalies at lags (a,b) −9, (c,d) 0, and (e,f) +9 months. Stippling indicates the 90% confidence.

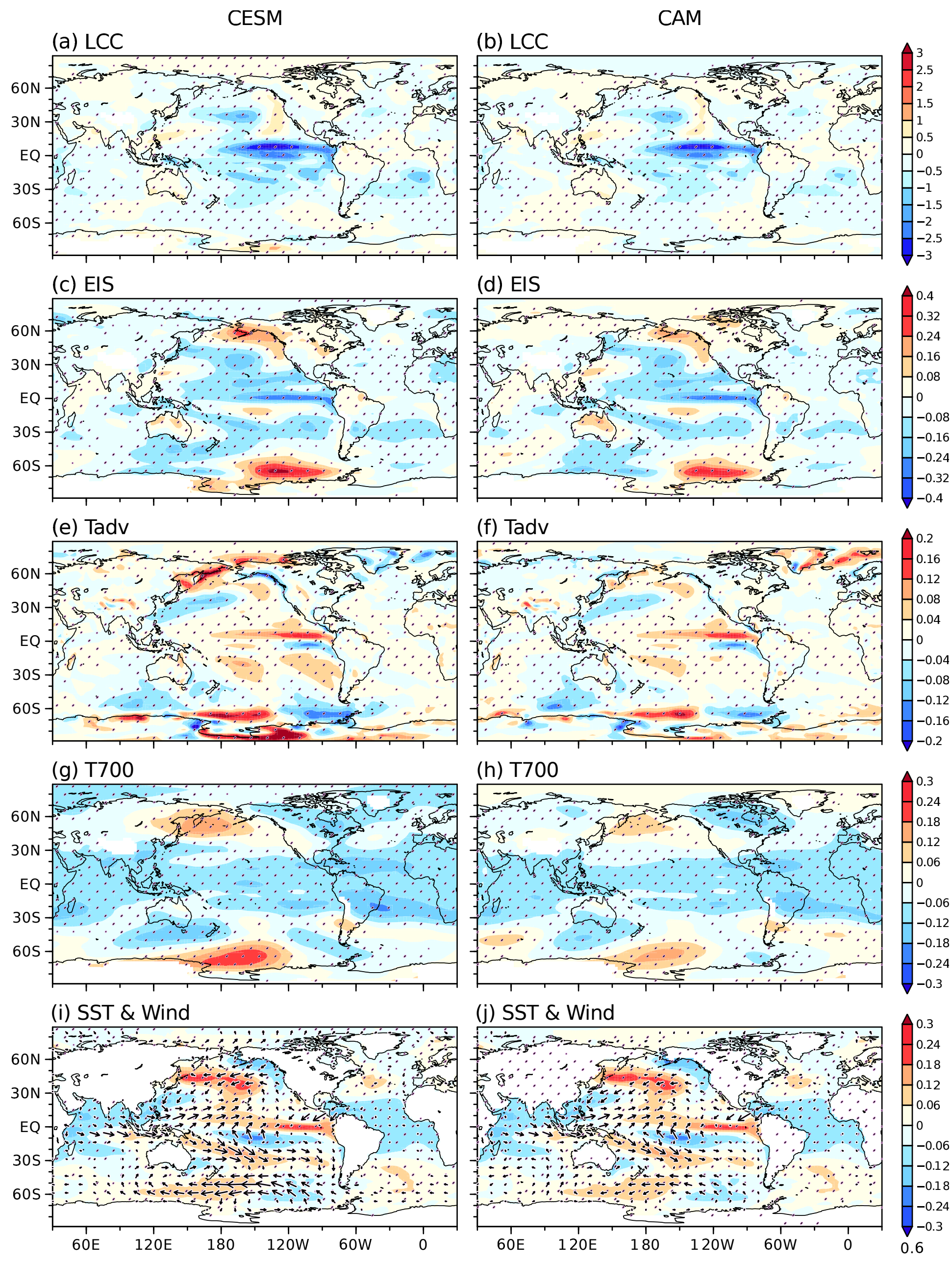


FIG. 4. Regression maps of (a,c,e,g,i) CESM and (b,d,f,h,j) CAM anomalies onto CESM interannual GMTOA at lag 0. (a,b) Low cloud cover (%). (c,d) EIS (K). (e,f) surface temperature advection (K day^{-1}). (g,h) 700-hPa temperature (K). (i,j) SST (shading; °C) and surface wind (arrows; m s^{-1}; only points with the 90% confidence are drawn). Stippling indicates the 90% confidence.

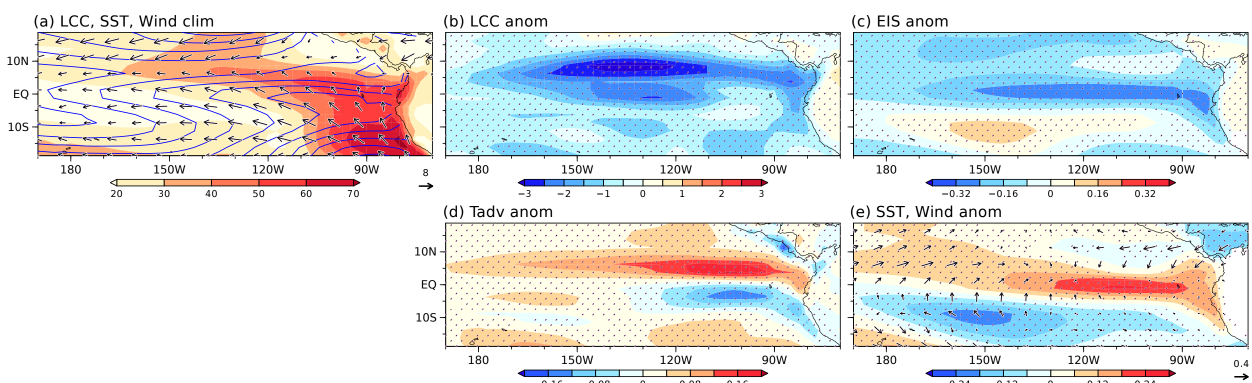


FIG. 5. (a) CESM climatology of low cloud cover (shading; %), SST (contoured for every 1 °C), and surface winds (arrows, m s^{-1}). (b-d) magnified maps of Figs. 4a,c,e,j, respectively.

4. Decadal variation

This section investigates the decadal GMTOA variability in CESM and contrasts it with the interannual counterpart. The lead-lag relationships on decadal timescales shown in Fig. 2a-d resemble those on interannual timescales. There is an approximately 90° phase-lagged relationship such that a peak in GMTOA corresponds to a positive time tendency in GMST (black lines in Figs. 2a,b; Xie et al. 2016). Compared to the interannual correlation, the weaker lag correlation with GMST on decadal timescales implies the importance of processes not tightly linked to GMST. CRE accounts for 75% of the GMTOA peak through shortwave heating, compounded by a secondary clear-sky effect (Table 1). The peak in CRE emerges one year later (Fig. 2c), explaining nearly all the GMTOA anomaly (Table 1). As shown by the red lines in Figs. 2a-d, CAM reasonably captures these features, with a GMTOA correlation coefficient of 0.61 at lag 0 between CESM and CAM. While stochastic atmospheric variability is nonnegligible, this result underpins the quantitative importance of the ocean effect.

a. Spatial pattern

Figure 6 shows the time evolution of radiation and surface temperature anomalies associated with the decadal GMTOA variations. Despite the similar temporal relationships on interannual and decadal timescales, there are marked differences in the spatial patterns of radiation. Remarkably, peak GMTOA on decadal timescales is associated with marked TOA radiation anomalies in the subtropical low-cloud regions (Figs. 6e,f). Strong positive signals occur over the Northeast Pacific and Southeast Indian Ocean, with a weaker signal over the South Atlantic. A positive signal in the eastern equatorial Pacific is much less dominant than on interannual timescales. The Southeast Pacific signal is not prominent at lag 0 but intensifies rapidly at lag +1, reaching a magnitude comparable to that in the North Pacific and Indian Ocean (Fig. 6g). These strong positive anomalies in the subtropical low cloud regions align with the peak of global-mean CRE (Fig. 2c). The corresponding SST anomalies at lag 0 and +1 years are also localized to the eastern subtropical oceans (Figs. 6f,h), in contrast to the interannual anomalies. Such timescale dependence implies marked differences in the physical processes.

It is noteworthy that the TPDV-like equatorial Pacific SST pattern transitions from a negative to a positive phase across lag 0 (Figs. 6b,j), although the maximum correlation between GMTOA and Niño3.4 SST drops to 0.4 (Fig. 2e). At lag ±3 year when the TPDV-like SST signals are maximized, strong radiation changes over the Pacific tend to cancel out (Figs. 6a,i), as in the peak phase of ENSO (Figs. 6a,e). This spatial compensation in TOA radiation associated with decadal fluctuations seems somewhat at odds with previous work claiming the effect of the equatorial Pacific warming pattern on global radiation (Zhou et al. 2016; Andrews and Webb 2018). This difference may reflect distinct global energy budgets governing natural variability and forced response. For example, unforced SST anomalies in the warm pool region, which efficiently induce GMTOA anomalies (Zhou et al. 2017), tend to be small.

b. Mechanism

Figure 7 shows anomalies in low cloud cover and environmental factors associated with decadal GMTOA variations in CESM and CAM. Here, lag +1-year fields are discussed to capture the rapid emergence of the Southeast Pacific signal and associated peak in global-mean CRE. Otherwise, the signal is qualitatively the same as the lag-0 fields (Fig. S3). With a high spatial correlation of 0.88, the net radiation anomalies are well explained by anomalous low cloud decrease maximized over over the Northeast Pacific, Southeast Pacific, and South Indian Ocean (Fig. 7a). The low cloud decrease is collocated with weakening of EIS (Fig. 7c). Because zonally uniform free-tropospheric temperature changes are not obvious (Fig. 7g), local SST warming strongly controls the decadal decrease of EIS and thus low clouds as positive low cloud-SST feedback (Fig. 7i). This low cloud-SST feedback through EIS corroborates CAM's reproducibility of the low cloud changes in CESM (Figs. 7b,d) and consequently GMTOA changes. The weakly opposing signal in the tropical Northeast Atlantic signal may be due to the lack of a pronounced low cloud deck there (Fig. 1). Rather, the North Atlantic cooling and associated Atlantic meridional mode could decrease low clouds over the South Atlantic (Tanimoto and Xie 2002) and Northeast Pacific (Miyamoto and Xie 2025). Similar to yet less dominant than the interannual variations, the decadal decrease in equatorial Pacific low clouds corresponds to the emergence of warm-phase TPDV (Figs. 7a,i).

The positive SST anomalies over the eastern subtropical oceans are concurrently associated with poleward wind anomalies extending westward and equatorward (Fig. 7i). These weakened trade wind anomalies induce warm-air advection (Fig. 7e), suppressing turbulent heat loss from the ocean that, together with radiative heating due to the low cloud decrease, raises the SST (Table

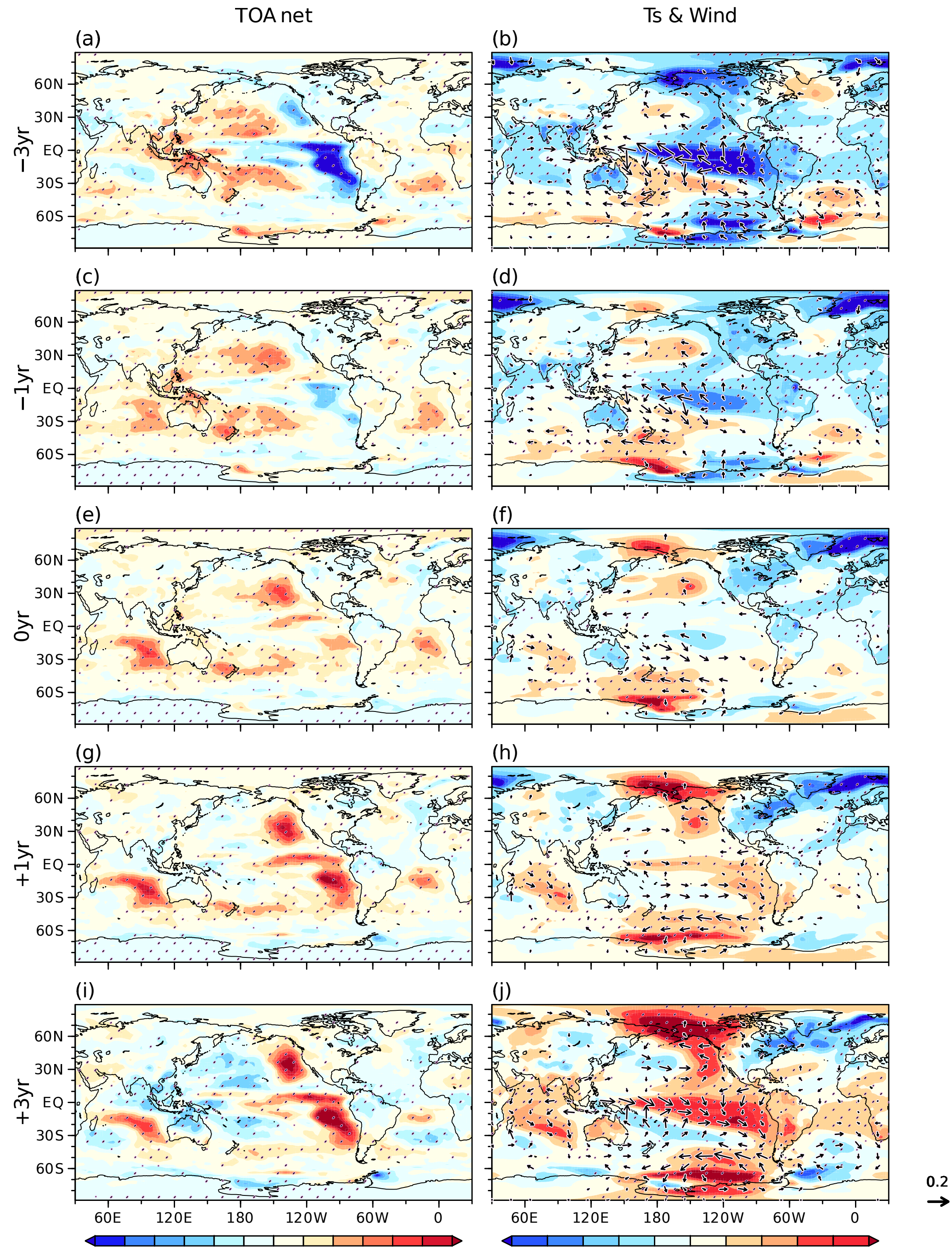


FIG. 6. As in Fig. 3, but for decadal anomalies. Lags (a,b) −3, (c,d) −1, (e,f) 0, (g,h) +1, (i,j) +3 years.

2). The equatorward extension of these SST anomalies accompanied by anomalous trade winds resembles the meridional modes (Chiang and Vimont 2004; Zhang et al. 2014) generated through wind-evaporation-SST (WES) feedback (Xie and Philander 1994). Modeling studies demonstrated that subtropical low cloud-SST feedback energizes the meridional mode-like variability as joint low cloud-WES feedback (Bellomo et al. 2014; Miyamoto et al. 2021; Miyamoto et al. 2023).

The comparison of the wind anomalies between CESM and CAM reveals the origin of the low cloud and meridional mode-like variability. In CESM, the poleward wind anomalies responsible for the elevated SSTs are embedded with extratropical circulation anomalies (Fig. 7i). While CAM partly captures the weakening of trade winds in WES feedback, it underestimates or even fails to simulate these extratropical circulations, except for the modest anomalies over the South Atlantic (Fig. 7j). Additionally, the underestimation of warm-air advection, which acts to decrease low cloud cover directly (Klein et al. 1995; Miyamoto et al. 2018), may contribute to the slightly underestimated low cloud anomalies in CAM (Figs. 7b,f). This indicates that extratropical stochastic atmospheric forcing triggers the subtropical low-cloud and meridional mode-like variability, thereby driving the GMTOA anomalies.

In summary, decadal TOA radiation changes are dominated by subtropical low cloud decks, without prominent concurrent SST anomalies in the deep tropics. Extratropical stochastic variability makes a pronounced contribution to generating subtropical SST and low cloud anomalies. These SST anomalies, in turn, enable the AMIP to capture changes in GMTOA through low cloud-SST feedback. These processes contrast sharply with the interannual GMTOA variations.

TABLE 2. Lag +1-year area-average surface heat flux (W m^{-2}; positive values for downward flux) regressed onto decadal GMTOA variability in CESM. The averaging domains are shown in Fig. 7a. See Appendix for the flux decomposition.

	LH_a	SH	SW+LW	D_o
North Pacific	0.17	0.03	0.41	-0.32
South Indian Ocean	0.38	0.04	0.49	-0.53
South Pacific	0.39	0.06	0.43	-0.5

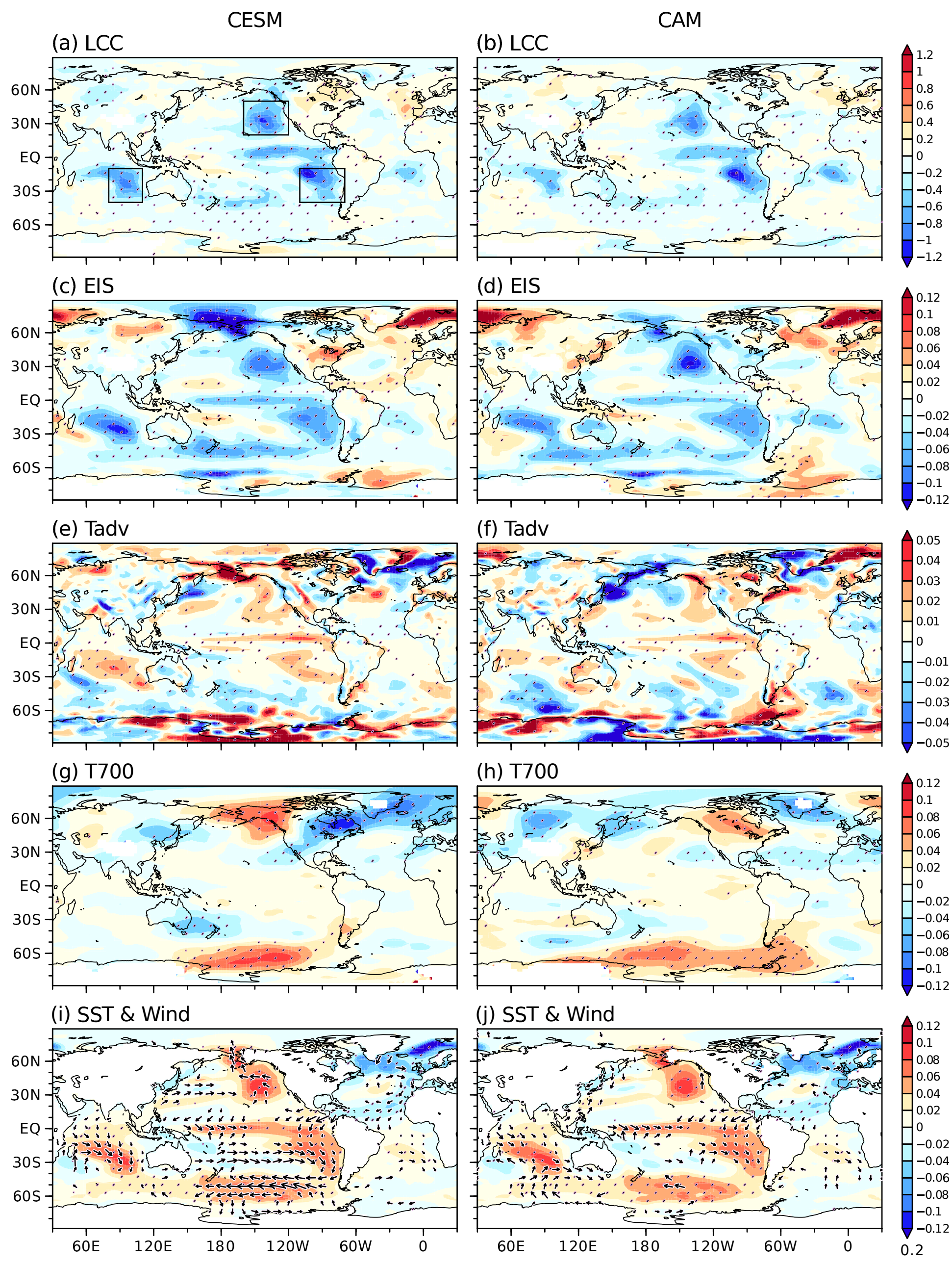


Fig. 7. As in Fig. 4, but for CESM decadal anomalies at lag +1 year.

5. Discussion

a. Comparison with CERES observations

Although short, the CERES observations serve as a valuable testbed to corroborate the findings from CESM. Here, we analyze 21-year detrended anomalies of GMTOA in the CERES observations and two AMIP simulations (CAMobs and AMobs). We perform ensemble averaging of the AMIPs prior to analysis.

Despite the weaker lagged correlation with GMST, the observed lead-lag relationship of GMTOA and GMST is out-of-phase (Figs. 2k,l), aligned with the CESM result. Around the GMTOA peak, CRE dominates the GMTOA anomalies while the clear-sky effect is secondary (Figs. 2m,n and Table 1). Although the reproducibility of AMIPs can be degraded by not only stochastic noise but also model biases, the two AMIPs reproduce the observed GMTOA and CRE reasonably well (Figs. 2k,m), indicative of the SST effect.

The corresponding patterns of net radiation in the observations and AMIP simulations are shown in Figs. 8a-c. Both the observations and AMIPs feature increased incoming radiation over the equatorial eastern Pacific and subtropical Southeast Pacific accompanied by decrease in low clouds (Figs. 8d-f). Consistent with the emergence of ENSO discussed previously, positive SST anomalies appear along the equatorial Pacific (Fig. 8g) in the phase transition from La Niña to El Niño (Fig. 2o) accompanied by anomalous warm advection (Figs. S4d-f) and decreased EIS (Figs. S4a-c). Cooling in tropical tropospheric temperature is somehow inconsistent between observations and AMIPs (Figs. S4g-i), and its effect on GMTOA is unclear. The weaker relationship with ENSO in observations may reflect the short observational record, inclusion of decadal and forced changes despite detrending, and overestimated ENSO response in CESM. We note that extending the CERES observations through 2024 to include the 2023-24 strong El Niño event (Xie et al. 2025; Minobe et al. 2025; Peng et al. 2025) leads to a marginal increase in the maximum correlation between GMTOA and Niño3.4 SST, from 0.4 to 0.5.

Meanwhile, positive TOA radiation anomalies over the subtropical Southeast Pacific correspond to a local rise in SST (Fig. 8g), suggestive of low cloud-SST feedback that is responsible for the AMIP reproducibility. These low cloud-SST co-variations are likely to be triggered by anomalous northwesterlies associated with extratropical cyclonic circulations (Fig. 8g). The AMIP simulations fail to simulate the circulation pattern (Figs. 8h,i), indicating the predominance of stochastic atmospheric variability. While we emphasize the effect of stochastically forced low cloud-SST feedback on decadal timescales, these processes are also hinted at by the interannual variations in CESM. Local maxima in net radiation and low cloud decrease over the Southeast Pacific (Figs. 3c and 4a) are accompanied by anomalous northwesterlies that are underestimated in CAM (Figs. 4i,j).

In summary, the CERES observations provide support for the contribution of ENSO and subtropical low cloud-SST feedback to GMTOA variations identified in CESM. A longer record is necessary to enhance the signal-to-noise ratio and isolate the decadal variability.

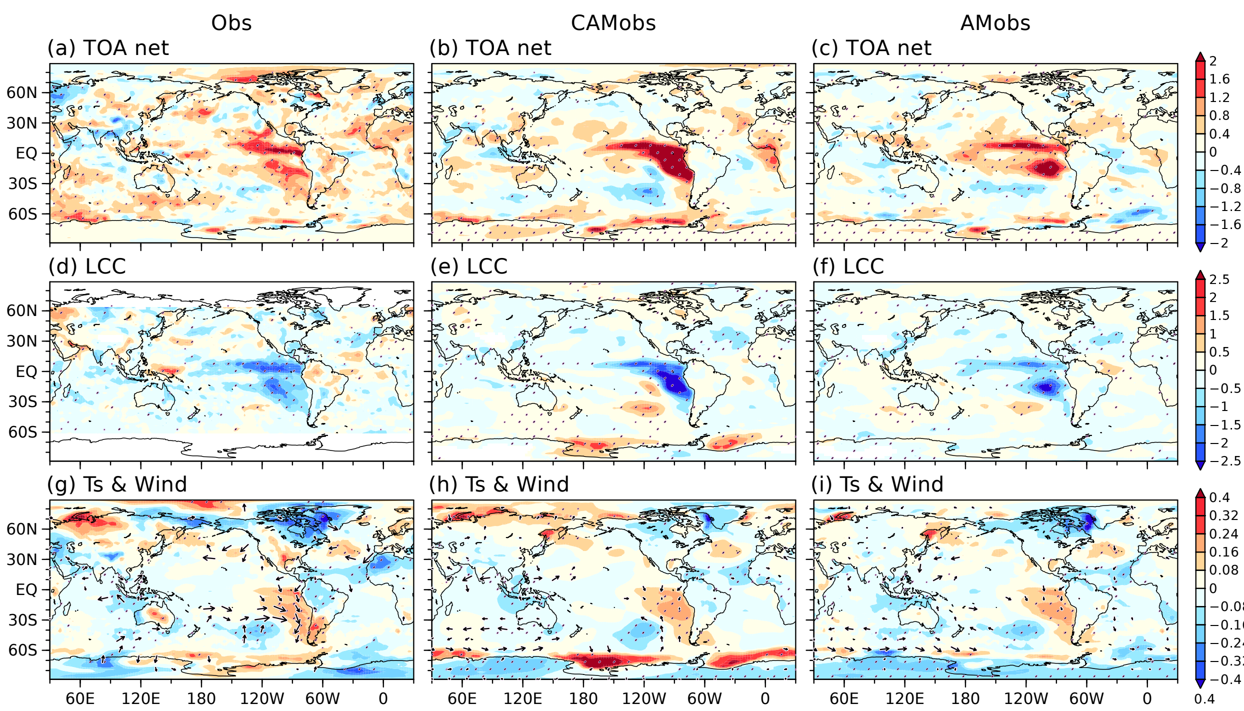


FIG. 8. Regression maps of (a,d,g) Observed, (b,e,h) CAMobs, and (c,f,i) AMobs anomalies onto CERES GMTOA at lag 0. (a-c) TOA radiation (W m^{-2}). (d-f) Low cloud cover (%). (g-i) surface temperature (shading; °C) and wind (arrows; m s^{-1}; only points with the 90% confidence are drawn). Stippling indicates the 90% confidence.

b. Origin of timescale dependence

This study highlights two drivers of GMTOA variations via low cloud changes: ENSO and extratropical atmospheric variability. In CESM, the former dominates on interannual timescales whereas the latter dominates on decadal timescales. It is not surprising that ENSO, the strongest natural mode of variability, plays a major role in GMTOA fluctuations. ENSO is essentially an interannual oscillation arising from redistribution of tropical ocean heat content (Jin 1997), with a peak period of 2-8 years in both observations and CESM2 (Capotondi et al. 2020). Meanwhile, SST variations driven by extratropical atmospheric forcing become more important on decadal and longer timescales through stochastic reddening (Hasselmann 1976). This diminishes the relative importance of equatorial Pacific-forced GMTOA variability on decadal timescales. It is noteworthy that the ENSO effect may be exaggerated in CESM because of the excessive ENSO strength and regularity (Capotondi et al. 2020). This may explain the strong ENSO effect on GMTOA and its lead-lag relationship with GMST in CESM compared with the observations (Figs. 2j,o).

Still, TPDV has statistically significant lagged correlations with GMTOA (Figs. 2e and 6). In addition to the TPDV forcing on GMTOA, TPDV may be driven in part by extratropical atmospheric variability associated with GMTOA variations. Previous studies argued that North and South Pacific SST anomalies forced by atmospheric stochastic variability can modulate TPDV via Pacific meridional modes (Okumura 2013; Di Lorenzo et al. 2015; Sun and Okumura 2019). Indeed, such meridional mode-like patterns are found in the Pacific after the GMTOA peak (Figs. 6f,h,j).

It is striking that the decadal TOA radiation changes are concentrated in subtropical low cloud decks without pronounced concurrent SST changes in the deep tropics (Figs. 6e-h). This coherent pattern likely emerges as a superposition of independent stochastic forcings on the low cloud-SST variations. In fact, the interbasin temporal correlation of radiation over low cloud regions (e.g., North Pacific versus South Pacific) is limited (not shown). Nevertheless, the inter-basin coupling of low cloud decks cannot be ruled out. The meridional mode-like variability and developing TPDV could play a role in the inter-basin coupling via changes in the deep tropics. Further studies—including those using partially coupled runs—are needed to better understand the cause and effect of low cloud variability, particularly the relative contributions of tropical and extratropical forcings and the possibility of interbasin influences.

6. Conclusion

This study investigates the natural variability of GMTOA based on a 500-year CESM2 preindustrial simulation and a corresponding perfect-model AMIP simulation. We show that the low-cloud radiative effect plays a dominant role in the GMTOA variations, with spatial patterns that differ markedly between interannual and decadal timescales. This difference reflects the relative influence of two distinct drivers: equatorial Pacific variability (e.g., ENSO) on interannual timescales and extratropical atmospheric variability on decadal timescales. The influence of both drivers on GMTOA is hinted at by CERES observations.

On interannual timescales, low cloud anomalies are distributed across tropical and extratropical oceans, with maxima over the equatorial eastern Pacific in the transition phase of ENSO. During positive GMTOA anomalies, reduced low cloud cover over the northeastern equatorial Pacific arises from anomalous warm advection due to a developing El Niño. Meanwhile, the remaining broad decrease in low clouds aligns with weakened stability primarily during the decaying phase of La Niña, which leaves an imprint on free-troposphere temperature and SST nonlocally through teleconnections. In contrast, decadal GMTOA variability features more localized radiation anomalies in the eastern subtropical low cloud decks without concurrent SST changes in the deep tropics. These cloud anomalies are collocated with underlying SST anomalies, which allow AMIPs to reproduce the CRE changes through low cloud-SST feedback. The low cloud-SST co-variations are triggered by stochastic wind anomalies associated with extratropical atmospheric variability. This timescale dependence likely reflects the nature of these drivers: ENSO peaks on interannual timescales due to tropical ocean dynamics, while extratropical forcing becomes increasingly important on longer timescales. This study for the first time emphasizes the importance of the extratropical-forced subtropical low cloud-SST variations on GMTOA.

The importance of SST patterns for GMTOA demonstrated by the perfect model framework affirms the use of AMIP experiments for simulating GMTOA variability during the CERES and pre-satellite era (Andrews et al. 2022; Schmidt et al. 2023) and for attributing it to regional SST anomalies (Zhou et al. 2017; Bloch-Johnson et al. 2024). Still, this study has shown that low cloud anomalies are not solely formed by tropical moist adiabat adjustment but also involve coupled low cloud-SST variations triggered by ENSO and stochastic atmospheric forcing in the extratropics. Disentangling multiple drivers requires isolating the individual contributions to the

SST anomalies. Given that GMTOA exhibits lagged correlations with ENSO and TPDV—each inherently irregular—it remains challenging to separate the equatorial Pacific-forced signal from others. To address this, ocean-atmosphere coupled modeling or advanced statistical techniques are likely to be required.

We find that the spatial pattern of TOA radiation anomalies associated with GMTOA variability markedly differs from that associated with GMST variability, which is characterized by pronounced signals in the equatorial Pacific and high latitudes (e.g., Xie et al. 2016). This discrepancy implies a redistribution of heat by atmospheric and oceanic circulations. The energy input may not only be passively advected but interact with the circulations. One plausible mechanism inferred from the decadal variations is that anomalous heat uptake in the North and South Pacific may propagate equatorward via the joint low cloud-WES feedback (Bellomo et al. 2014; Miyamoto et al. 2023) and subsurface ocean adjustment (Luongo et al. 2025), potentially modulating TPDV and consequently GMST. The role of coupled dynamics in linking global energy imbalance to temperature patterns warrants further investigation.

CERES data reveal a marked positive trend of GMTOA over the past two decades. The associated SST warming is pronounced in the Northeast Pacific, the South Indian Ocean, and the South Atlantic (Fig. 4 in Loeb et al. 2024). This pattern bears some resemblance to that of the unforced GMTOA variations identified in this study. Given the importance of subtropical low clouds in both forced and unforced GMTOA variability, it is essential to carefully attribute the observed changes.

Although partially supported by the observational datasets, this study is primarily based on a single model and subject to model biases, including excessively strong ENSO in CESM2 (discussed in Section 5). In addition to the bias in equatorial Pacific variability, Tsuchida et al. (2023) suggested that the sensitivity of tropical atmosphere to anomalous equatorial SST also contributes to the intermodel spread in GMTOA variations. Moreover, many climate models underestimate subtropical low cloud-SST feedback (Kim et al. 2022; Kang et al. 2023), which is, however, strong in CESM2 (Kang et al. 2023; Larson et al. 2024; Miyamoto and Xie 2025). As low clouds are a critical factor in the uncertainty of climate feedback (Zelinka et al. 2020), natural GMTOA variations may be related to the spread of projected warming (Zhou et al. 2015; Lutsko and Takahashi 2018). Improvements of these biases promise a unified understanding of natural GMTOA variability in a multi-model framework. Together with a better understanding of radiatively forced response, this will ultimately help us understand the historical GMTOA variations and constrain future projections.

Acknowledgments. We thank Dr. Isla Simpson for setting up the perfect-model CAM6 simulation and CESM CVCWG for making CESM2/CAM6 simulations publicly available. The AM4 simulation is supported by the Japanese Ministry of Education, Culture, Sports, Science and Technology (MEXT) program for the advanced studies of climate change projection (JPMXD0722680395). The National Center for Atmospheric Research (NCAR) is sponsored by the National Science Foundation under Cooperative Agreement 1852977.

Data availability statement. The observational data used in this study are available online (ERA5: `https://cds.climate.copernicus.eu`; CERES-EBAF: `https://ceres.larc.nasa.gov/data`; MODIS: `https://ladsweb.modaps.eosdis.nasa.gov/archive/allData/61/MOD08_D3`; OISST: `https://psl.noaa.gov/data/gridded/data.noaa.oisst.v2.highres.html`). CESM2 and CAM6 simulations were downloaded from the casper system in National Center for Atmospheric Research. The authors can provide AM4 experiments upon reasonable requests.

APPENDIX

Decomposition of surface heat flux

Anomalous surface heat flux can be decomposed into latent heat (LH), sensible heat (SH), shortwave (SW), and longwave (LW) components. LH is a mixture of atmosphere-driven and SST-damping components. Following Xie et al. (2010), the SST damping term may be cast as

$$\mathrm{LH}_\mathrm{o}' = \overline{\mathrm{LH}}\left(\frac{1}{\overline{q_s}}\frac{\mathrm{d}\overline{q_s}}{\mathrm{d}T_a}\right)\mathrm{SST}' \tag{A1}$$

where T_a and q_s are air temperature and saturation specific humidity following the Clausius-Clapeyron equation, respectively. Overbar and prime denote monthly climatology and anomaly, respectively. The residual of anomalous latent heat flux represents the atmosphere-driven component (LH_a') related to anomalous atmospheric conditions,

$$\mathrm{LH}_\mathrm{a}' = \mathrm{LH}' - \mathrm{LH}_\mathrm{o}'. \tag{A2}$$

On interannual and longer timescales, the heat storage becomes negligible and downward net surface heat flux is almost balanced with ocean heat transport in the ocean mixed-layer heat budget. We therefore use negative of net surface heat flux as a proxy of the ocean dynamical effect (D_o).

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