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Possible Thermal Effect of Tibetan Plateau on the Atlantic Meridional Overturning Circulation

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Key Points:

- The Tibetan Plateau (TP) heating (cooling) can cause enhanced moisture transport from the North Atlantic (Pacific) to Pacific (North Atlantic)
- The TP heating (cooling) can strengthen (weaken) the Atlantic meridional overturning circulation (AMOC)
- The TP warming at the end of 21st century under the RCP8.5 scenario can lead to a 10% AMOC enhancement

Supporting Information:

Supporting Information may be found in the online version of this article.

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Abstract

The Tibetan Plateau (TP) is important in shaping the global climate. To understand how its thermal condition may affect the Atlantic meridional overturning circulation (AMOC), we conduct a series of albedo forcing experiments to simulate the TP warming and cooling conditions. The results show that the TP warming can immediately generate anticyclone in the subtropical North Atlantic, leading to more water vapor transporting from the North Atlantic to eastern tropical Pacific. This results in less precipitation over the North Atlantic and increases the sea surface salinity. The latter enhances the deep-water formation in the subpolar North Atlantic and thus the AMOC. In contrast, the TP cooling weakens the AMOC via opposite processes. These findings imply that the AMOC change can be affected not only by local factors but also remote forcing from Asia.

Plain Language Summary

The Tibetan Plateau (TP) plays a vital role in regulating the regional and global climate. By far, most studies focus on the impacts of TP thermal effect on atmospheric change. Few efforts have been devoted to investigate the TP thermal effect on ocean circulation. In this study, we artificially reduce and increase the surface albedo over the TP region to simulate the TP warming and cooling conditions. The results show that the TP thermal conditions can affect the Atlantic meridional overturning circulation (AMOC) strength by altering the inter-basin moisture transport between the North Atlantic and tropical Pacific. The TP heating immediately generates a westward Rossby wave train that can arrive at the North Atlantic, resulting in anomalous high-pressure system over there. This helps to bring more moisture from the North Atlantic to eastern tropical Pacific. The moisture divergence over the North Atlantic increases the sea surface salinity, which is responsible for the enhanced deep-water formation and thus the AMOC. In contrast, the TP cooling leads to AMOC weakening via opposite processes. Our study implies that the 7°C TP warming at the end of 21st century under the representative concentration pathway 8.5 (RCP8.5) scenario can enhance the AMOC by about 10%.

1. Introduction

The Tibetan Plateau (TP), known as the highest plateau in the world now, began to rise in late Cenozoic and sped up at 10–8 million years ago (Harrison et al., 1992; Molnar et al., 1993), causing substantial changes in the earth's climate (An et al., 2001; Kutzbach et al., 1993; Ramstein et al., 2019; Ruddiman, 1997).

The impact of TP on regional and global climate has been studied for around 70 years beginning by using simple one-dimensional model (Charney & Eliassen, 1949), to atmosphere only general circulation model (Manabe & Terpstra, 1974; Ruddiman and Kutzbach, 1989), and then coupled ocean atmosphere models (Rind et al., 1997; Wen et al., 2021). Using one-dimensional barotropic model, Charney and Eliassen first suggest that the big mountains can produce disturbances of the middle-latitude westerlies (Charney & Eliassen, 1949). Subsequently, numerical experiments with atmosphere only model have shown that the mountains may alter the global distribution of precipitation (Manabe & Terpstra, 1974), change the propagation speed of large-scale waves along the westerlies (Boyle, 1986). Specifically, the existence of TP can intensify the Asian monsoon (Prell, 1984; Ruddiman and Kutzbach, 1989), bring wetter climates to the south and east of the TP and drier climates to the north and west of the TP, and cause northern polar latitudes cooling (Kutzbach et al., 1993). During recent years, the development of coupled ocean atmosphere models provides more comprehensive and thoroughly understanding of TP's effect on global climate. For example, Vavrus and Kutzbach (2002) show that the Atlantic Meridional Overturning Circulation (AMOC) could exist in a world with lower topography and higher CO_2 concentration. Fallah et al. (2016), simply by removing TP, demonstrate that the absence of TP can modulate wind-driven circulation over the North Atlantic and lead to a decrease in the AMOC. Su et al. (2018) show that the removal of TP can weaken the AMOC and enhance the Pacific meridional overturning circulation (PMOC) via redistribution of atmospheric water vapor. Furthermore, Yang and Wen (2020), as well as Wen and Yang (2020) further suggest that in addition to atmospheric water vapor reorganization, the sea ice feedback and ocean subduction are both important for TP's influence on AMOC and PMOC.

While the above studies that focus on the effects of TP include both mechanical and thermal forcing, other studies have emphasized the impact of the dominant thermal forcing of the TP (Rajagopalan & Molnar, 2013; Wu et al., 2015). Flohn (1957) first show that the TP is a heat source during boreal summer. The ascending air over the TP can subside zonally over Iran and meridionally over its north and south region to generate a dry climate over there (Sato & Kimura, 2005; Yeh, 1982). In addition, the thermal effect of the TP can even exceed its mechanical effect to dominate over Indian monsoon onset (Chou & Neelin, 2001; Wu et al., 2015). Boos and Kuang (2010) demonstrate that the plateau insulation, instead of its thermal heating, sustains the Indian monsoon strength. Moreover, the thermal effect of the TP can also reach the remote areas, such as western Pacific and North America by generating Rossby waves propagating eastward or westward (Duan et al., 2017; Lee et al., 2013; Liu et al., 2013; Zhao et al., 2012). Inspired by rapid warming over TP during recent decades and drastic warming projected in the future (Figure S1 in Supporting Information S1; Guo et al., 2019; Zhou & Zhang, 2021), the heating effect of TP on weather and climate become an active area of research (Wang et al., 2008). Some works focus on TP's heating effect on its upstream climate, and show that it can significantly influence temperature and precipitation over Atlantic-African-European sector in boreal summer (Lu et al., 2018, 2019). In general, most of prior works focus on TP's thermal role in atmospheric change. How it affects the ocean circulation receives much less attention.

The AMOC plays a vital role in shaping the Earth's climate by transporting large heat northward in the Atlantic (IPCC, 2013; Zhu and Liu, 2020). Theoretically, the AMOC is affected by surface wind at short timescale, while its climatological strength is determined by surface buoyancy flux (Kuhlbrodt et al., 2007). The North Atlantic deep-water formation is known as the most direct factor to determine the AMOC strength at present-day, which is maintained by strong surface evaporation and northward salty water transport over North Atlantic (Ferreira et al., 2018). The phase 5 and 6 of the Coupled Model Intercomparison Project (CMIP5 and CMIP6) have already projected that the AMOC will weaken by 15%–60% under the representative concentration pathway 8.5 (RCP8.5) scenario (Figure S2 in Supporting Information S1; Cheng et al., 2013; Weijer et al., 2020). Most works have focused on greenhouse gases (Zhu et al., 2015), aerosol (Menary et al., 2020), and sea ice (Liu et al., 2019) on AMOC variations. How TP's thermal condition may influence the AMOC is an emerging scientific issue that worth studying.

In this study, we conduct a series of sensitivity experiments to investigate the thermal effect of TP on AMOC by imposing albedo forcing over TP to simulate the TP surface warm and cold conditions. These idealized simulations with a state-of-the art Community Earth System Model (CESM) allow us to see detailed processes regarding how the TP surface thermal conditions affect the AMOC, which is very difficult to be extracted from such a complex climate system and is rarely addressed in previous works.

2. Model and Simulations

In this study, we use a coupled climate system model, CESM version 1.0.4, developed by the U.S. National Centre for Atmospheric research. This model includes dynamic atmosphere, land, ocean, sea ice components and one coupler (Hunke & Lipscomb, 2010; Lawrence et al., 2012; Park et al., 2014). No flux adjustments are used in CESM. The CESM can well simulate the observed climatology (Figure S3 in Supporting Information S1). It has

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been widely used and validated to study the Earth's past, present, and future climate (e.g., Hurrell et al., 2013; Yang et al., 2017).

2.1. Coupled Climate Simulation

We choose the low-resolution configuration (T31_g37) for fully coupled simulations. The horizontal grid of atmospheric model (CAM5) and land model (CLM4) are roughly 3.75° × 3.75° with 26 vertical levels. The ocean model (POP2) and sea ice model (CICE4) adopt a finer oceanic horizontal grid, with 60 vertical levels, a uniform 3.6° spacing in the zonal direction, and a non-uniformly spacing in the meridional direction: It is 0.6° near the equator, gradually increasing to the maximum 3.4° at 35°N/S and then decreasing poleward. The first experiment is CTRL, which uses standard configuration with a preindustrial CO_2 concentration of 285 ppm. The CTRL simulation is integrated for 1,500 years to reach an equilibrium state (Yang et al., 2017) and is integrated for additional 200 years for comparison with the sensitivity experiments. It is well known that the albedo effect can influence surface temperature. When surface albedo decreases, land surface absorbs more solar radiation and surface temperature increases accordingly, and vice versa. Thus, following the previous works, for example, Wang et al. (2008) and Boos and Kuang (2010), we artificially reduce the surface albedo over the TP regions (25°–45°N, 70°E–104°E) by 50% and 80% to mimic the TP surface warming and artificially increase the surface albedo over the TP regions by 80% and 150% to simulate the TP surface cooling. These four sensitivity experiments are named TPW_−0.5 A, TPW_−0.8 A, TPC_0.8 A, and TPC_1.5 A. The albedo changes are shown in Figure S4 in Supporting Information S1. The sensitive experiments start from the year 1501 of CTRL and are integrated for 200 years. Except for land surface albedo over the TP, all other boundary conditions remain the same as that in CTRL. The last 50 years of these simulations are used to do the analysis. We focus on the annual mean climate responses since the climate responses are similar in boreal winter and summer.

2.2. Slab Ocean Simulations

In order to confirm the atmospheric circulation change in CESM low resolution runs, we further use CAM5 with higher-resolution coupled with a slab ocean model (SOM). The CAM5 are roughly 1.9° × 2.5° in the horizontal with 26 vertical levels. The SOM simulations contain a control run and two albedo forcing run. The control run is named CTRL_SOM, which is integrated for 200 years using the same configuration as CTRL, except that the dynamical ocean is replaced by a mixed layer. The albedo forcing runs are named TPW0.5_SOM and TPC0.5_SOM, in which the surface albedo over TP region is decreased and increased by 50%, respectively. The sensitive experiments are also run 200 years and the last 50 years are used to deduce the equilibrium change.

3. AMOC Change

In the climatology state, the AMOC consists of a wind-driven subtropical cell in the tropical Atlantic and a thermohaline circulation (Weaver et al., 1993; Yang and Wen, 2021). The mean state of AMOC is well simulated in CTRL compared with that from ECCO-v4 ocean reanalysis (Ferreira et al., 2018), with a maximum value of about 15–20 Sv (1 Sv = 10^6 m^3/s) at the depth range of 1,000–1,500 m (contours in Figures 1c–1f). The AMOC is intensified in response to the TP heating while it is weakened under the TP cooling with the amplitude of AMOC response proportional to the magnitude of albedo forcing. For example, the AMOC is enhanced by 1 Sv in TPW_−0.5 A, which is 1 Sv weaker than AMOC response in TPW_−0.8 A (Figure 1c vs. 1d). Similarly, the weakened AMOC in TPC_1.5 A is much more pronounced than that in TPC_0.8 A (Figure 1e vs. 1f). These indicate a robust AMOC response to TP's thermal effect. These changes occur after more than 50 years when the forcing is imposed, indicating a slow oceanic thermohaline adjustment to TP heating (Figure 1b). Note that the AMOC change in response to TP heating or cooling mainly occurs in its thermohaline branch, while the wind-driven cell, manifested by the streamfunction contours in the upper 200 m of the tropics (30°S–30°N), is nearly unchanged. This suggests that the thermohaline branch of the AMOC is more sensitive to the TP thermal forcing.

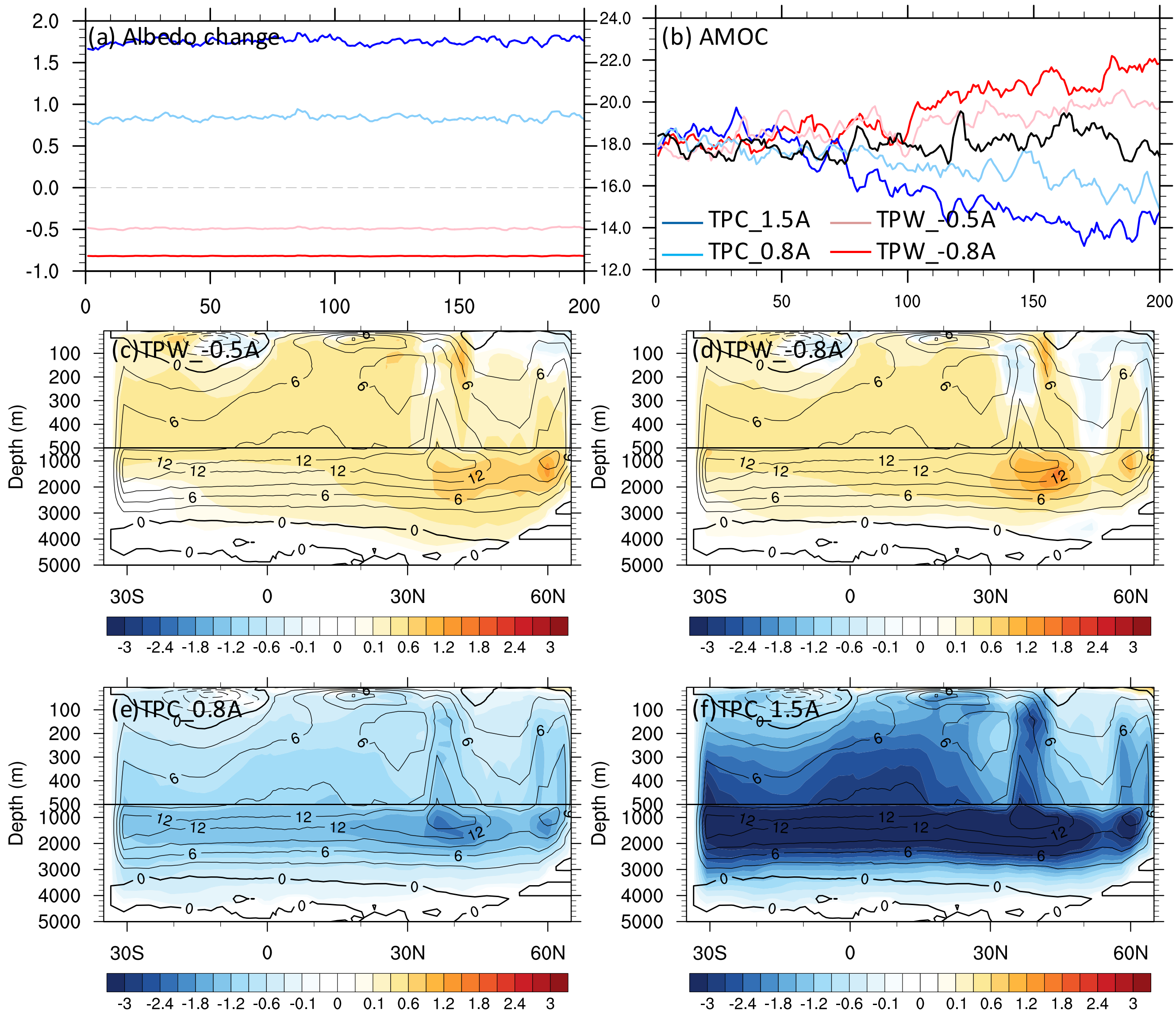

Figure 1. (a) Temporal evolution of changes in surface albedo over Tibetan Plateau region. The values are obtained by sensitive runs minus CTRL run and divide CTRL run. (b) Temporal evolution of Atlantic meridional overturning circulation (AMOC) index. The AMOC index is defined as the maximum streamfunction in the range of 0°–10°C over 20°–70°N in the Atlantic. In (a) and (b), the black line for CTRL, the blue line for TPC_1.5 A, the light blue line for TPC_0.8 A, the pink line for TPW_−0.5 A, and the red line for TPW_−0.8 A. (c)–(f) are the spatial distribution of AMOC in CTRL simulation (contours) and its change in TPW_−0.5 A, TPW_−0.8 A, TPC_0.8 A, and TPC_1.5 A, respectively. Units: Sv. 1 Sv = 10^9 kg/m3.

4. Atmospheric Responses

The AMOC change is determined by the changes in surface buoyancy flux in the North Atlantic, which is mainly attributed to the atmospheric response. The 80% reduction of surface albedo results in 7°C warming over the TP region (Figure 2a), with much stronger warming occurred in boreal winter (Figure not shown). The warming magnitude over TP in TPW_−0.8 A is very similar to that under the RCP8.5 scenario (Figure S13 in Supporting Information S1). The TP warming forces an anomalous ascending motion (Figure 3d), resulting in the local cyclonic circulation at lower level and anticyclonic circulation at upper troposphere (Figures 2b and 2c). This baroclinic atmospheric response is consistent with those of air pumping effect of the TP (Wu et al., 2007). The impact of the TP heating on the North Atlantic climate is via the Rossby wave trains. This teleconnection can be illustrated by horizontal wave activity flux (Takaya & Nakamura, 1997). Annually, the wave flux propagates eastward in the mid-latitudes along the westerlies (Figure S5a in Supporting Information S1; Yang et al., 2021).

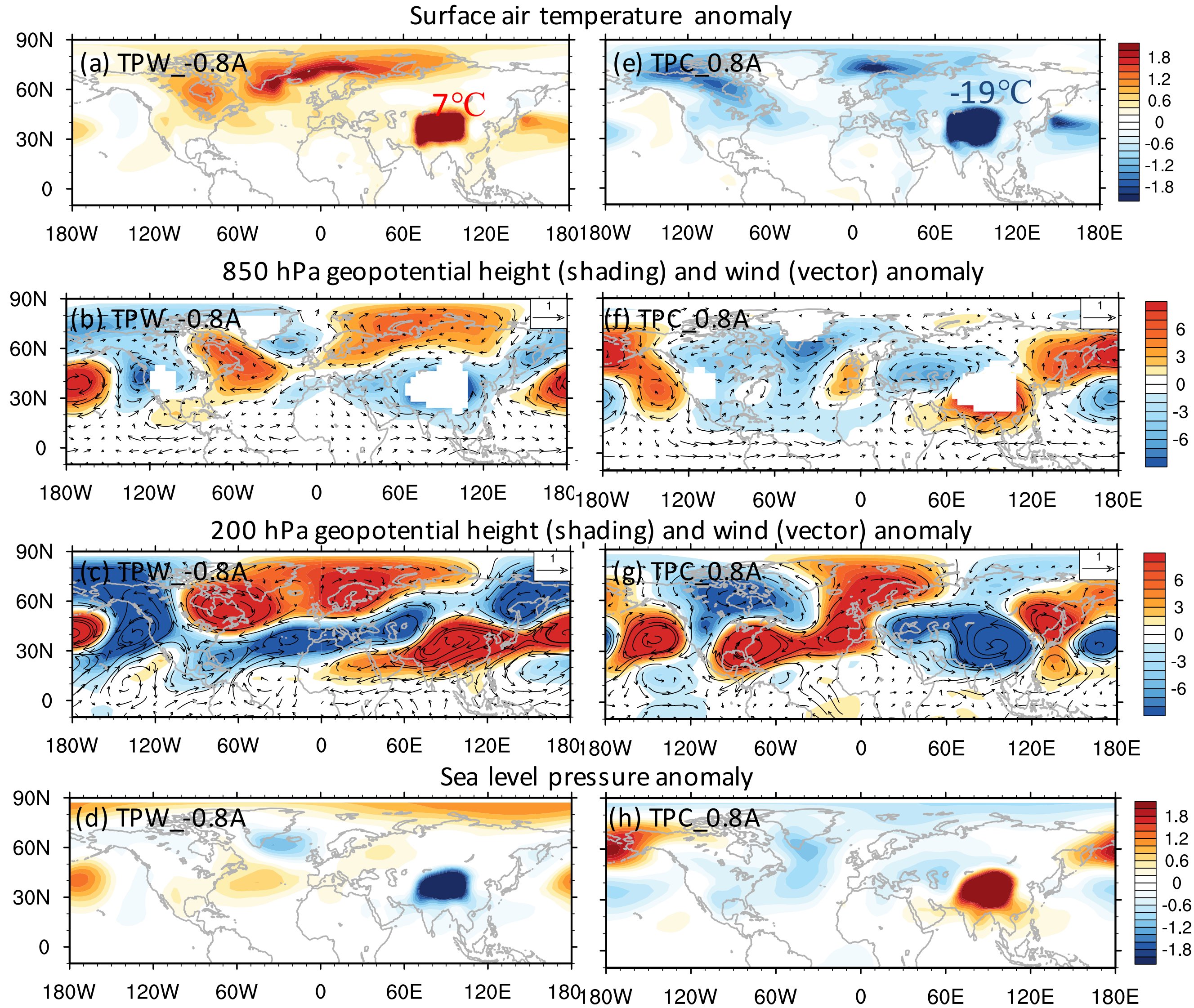

Figure 2. Annual mean changes in (a) and (e) surface air temperature (SAT; °C), (b)and (f) geopotential height (shading; m) and wind (vector; m/s) at 850 hPa, (c) and (g) geopotential height (shading; m) and wind (vector; m/s) at 200 hPa, and (d) and (h) sea level pressure (hPa). The left panel is for TPW_−0.8 A minus CTRL and right panel is for TPC_0.8 A minus CTRL. The numbers in (a) and (e) denote the maximum temperature change. Only difference exceeding the 95% significance level according to the Mann-Kendall test are plotted in the figures.

However, the TP heating can weaken the eastward wave flux, resulting in anomalous westward propagation of Rossby waves (Figure S5b in Supporting Information S1), which can finally cause baroclinic response over the subtropic North Atlantic, with anomalous high-pressure at lower troposphere and anomalous low-pressure at tropopause (Figures 2b and 2c). Note that in the climatological state, the North Atlantic is dominated by high pressure and anticyclonic circulation. Thus, the high-pressure anomaly over the North Atlantic indicates a strengthened subtropical high over there, which further prohibits the convection, resulting in decreased low clouds by 2% (Figure S6e in Supporting Information S1), increased incoming shortwave by 1.3 W/m^2 (Figure S6a in Supporting Information S1) and warmer sea surface temperature (SST; Figures 4a). The warmer SST leads to increased latent heat flux and outgoing longwave radiative flux into atmosphere (Figure S6b and S6c in Supporting Information S1), which finally warm the surface air (Figure 2a). In addition, the TP surface warming can also propagate westward with the Rossby waves (Figure S5b in Supporting Information S1), contributing to the North Atlantic surface warming as well. The westward propagation of TP surface warming has been validated by Zhao et al. (2012), in which they suggest that tropospheric warming over the Asian continent can travel toward the west over the extratropical Atlantic-Eurasian section with the anomalous westward propagation of the wave energy.

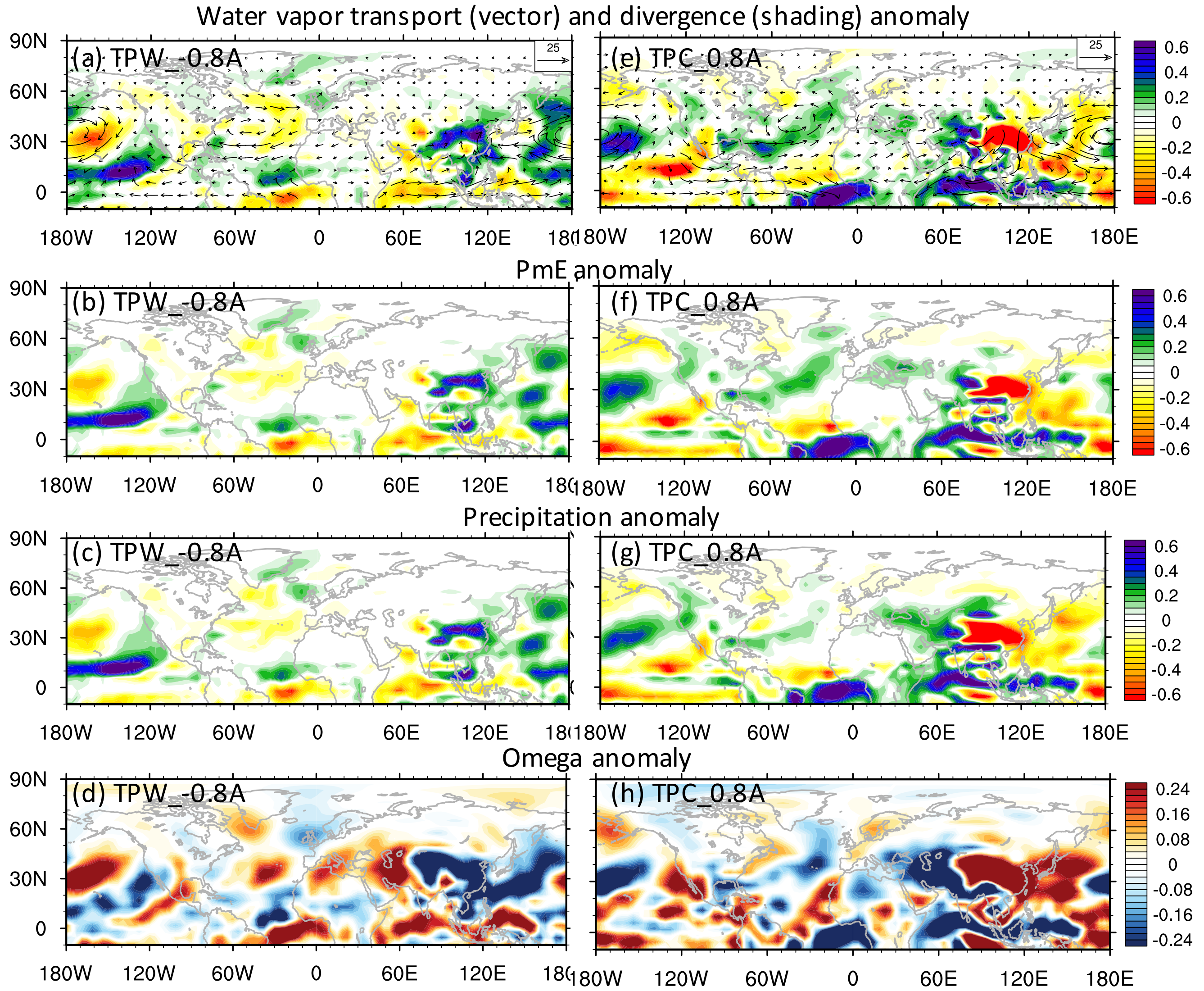

Figure 3. Annual mean changes in (a) and (e) vertically integrated moisture transport (vector; kg/m/s) and its convergence (shading; negative for convergence; 10–5 kg/m2/s), (b) and (f) precipitation minus evaporation (mm/day), (c) and (g) precipitation (mm/day), and (d) and (h) vertical velocity (0.01 Pa/s; positive for downward motion). The left panel is for TPW_−0.8 A minus CTRL and right panel is for TPC_0.8 A minus CTRL. Only difference exceeding the 95% significance level according to the Mann-Kendall test are plotted in the figures.

The warm surface air temperature (SAT)-ridge relationship over the North Atlantic found in this study indicate that the surface air temperature change is forced by the atmosphere instead of ocean surface (Bjerknes, 1964). On the contrary, the TP surface cooling generates low-level cyclonic circulation and thus the cold temperature anomaly over the North Atlantic (right panel in Figure 2).

The global hydrological cycle is also changed in response to the atmospheric adjustment. The vertically integrated water vapor transport roughly follows the low-level wind change (Figure 3a vs. Figure 2b). In TPW_−0.8 A, the low-level cyclonic circulation over the TP region and anticyclonic circulation over the western North Pacific result in enhanced southerlies over East Asia, which brings substantial water vapor from Indian Ocean and western Pacific into the East Asia, and increases the East Asian subtropical rainfall north of 30°N (Figures 3a and 3c). This phenomenon mainly occurs in boreal summer and is also captured in previous works (e.g., Wang et al., 2008; Yang and Wen, 2020). Over the North Atlantic, the high-pressure anomaly results in northeasterlies at the south tip of it, which transports more water vapor from the North Atlantic to eastern Pacific (Figure 3a). The divergence of water vapor in the North Atlantic causes less precipitation there (Figure 3c) and thus the less freshwater flux into ocean (Figure 3b). In the tropical Atlantic, previous work suggests that the TP heating can

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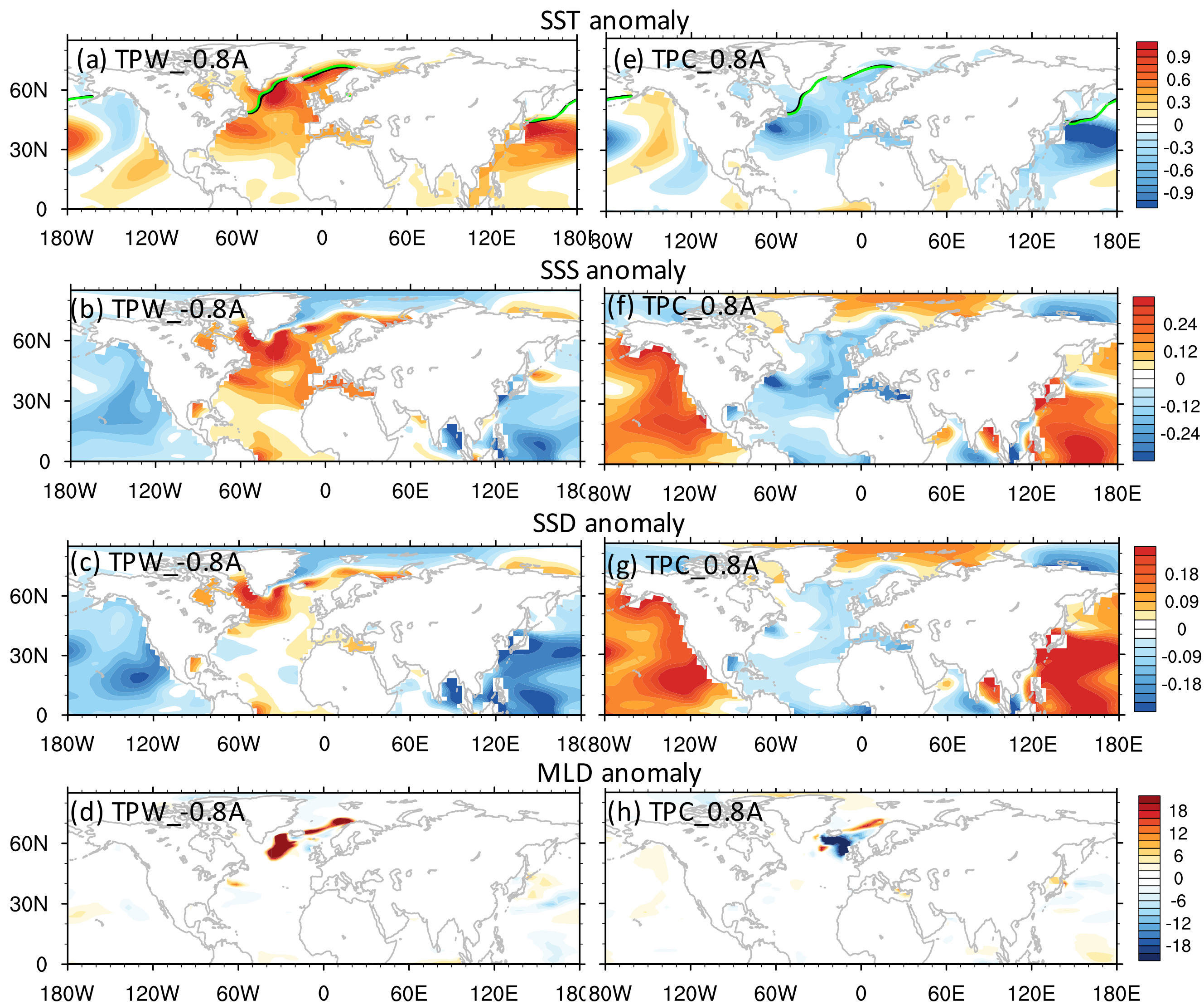

Figure 4. Annual mean changes in (a) and (e) sea surface temperature (SST; °C), (b) and (f) sea surface salinity (SSS; psu), (c) and (g) sea surface density (SSD; kg/m3), and (d) and (h) mixed layer depth (MLD; m). The left panel is for TPW_−0.8 A minus CTRL and right panel is for TPC_0.8 A minus CTRL. In (a) and (e), the black contours denote the sea ice margin in CTRL while the green contours denote the sea ice margin in thermal forcing experiments. Only difference exceeding the 95% significance level according to the Mann-Kendall test are plotted in the figures.

lead to a northward shift of the intertropical convergence zone (ITCZ; Lu et al., 2018). Here, we show that more water vapor transports from the Southern Hemisphere to the NH, resulting in precipitation increase (decrease) in the tropical North (South) Atlantic (Figure 3c). The north-south dipole pattern of precipitation change over the tropical Atlantic also implies the northward shift of the ITCZ, which is opposite to its southward shift under global warming (Mamalakis et al., 2021). As expected, the response of hydrological cycle exhibits opposite feature under TPC_0.8 A, with more water vapor transporting from eastern Pacific to the North Atlantic and thus more precipitation over there (left panel in Figure 3).

The atmospheric response to TP thermal effect is robust confirmed by our albedo forcing experiments with different magnitude (Figures S7 and S8 in Supporting Information S1). That is, the TP surface warming can cause warmer SAT, anticyclonic circulation, water vapor divergence and thus less precipitation over the North Atlantic while the TP surface cooling results in colder SAT, cyclonic circulation, water vapor convergence and thus more precipitation over the North Atlantic. Moreover, we conduct SOM simulations with higher-resolution to further check the atmospheric response. Since the atmospheric response is symmetric to TP warming and cooling in our low-resolution experiments, we only show its difference between TPW0.5_SOM and TPC0.5_SOM in Figures

S9 and S10 in Supporting Information S1. As expected, the SOM runs also reproduce the main features of atmospheric change in our low-resolution simulations.

5. Oceanic Responses

It is commonly recognized that the NADW formation and Southern Ocean surface winds are two key elements in determine the strength of the AMOC. In this study, the Southern Ocean surface winds are nearly unchanged during 200 years (Figure not shown), so the changes in NADW formation plays a critical role in AMOC change. The NADW formation is enhanced in the Greenland–Iceland–Norwegian Sea (GIN seas), contributing to the AMOC enhancement in response to the TP heating. This enhanced NADW formation is occurred in boreal winter (Figure not shown). The changes in surface buoyancy flux in the North Atlantic are shown in Figure 4. The SST is increased due to the enhanced incoming solar radiation (Figure S6a in Supporting Information S1). The increased SST is accompanied by a slight sea ice retreat (contours in Figure 4a). The SST warming in the GIN seas reduces the sea surface density (SSD) and weakens the deep-water formation. Concurrently, driven by the precipitation reduction (Figure 3c), the sea surface salinity (SSS) is increased over the large area of the North Atlantic (Figure 4b). The positive contribution from the SSS increase, dominating over the negative contribution from the SST, leads to SSD increase over the NADW region that enhances the NADW formation (Figure 4c). This is also manifested by the deepened mixed layer depth (Figure 4d). However, since the atmospheric circulation and hydrological change exhibits opposite features under TP cooling, the SSD is decreased, so does the mixed layer depth (right panel in Figure 4). The North Atlantic SST cooling under TP cooling condition is accompanied by a slight sea ice expansion (contours in Figure 4e). We also check the response of ocean surface properties in TPW_−0.5 A and TPC_1.5 A in Figure S11 in Supporting Information S1. It shows that the climate response in TPW_−0.5 A resembles to that in TPW_−0.8 A, that is, the warmer SST, increased SSS and SSD, and deepened mixed layer depth, except that the magnitude is weaker (Figures S11a–S11d in Supporting Information S1 vs. Figures 4a–4d). Similarly, the climate response in TPC_1.5 A is closely resembled to that in TPC_0.8 A but with much stronger amplitude (Figures S11e–S11h in Supporting Information S1 vs. Figures 4e–4h).

6. Conclusion and Discussion

In this study, we conduct a series of albedo forcing experiments using state-of-the-art climate model to understand the possible thermal effect of TP on the AMOC. The results show that the TP surface heating can immediately cause an anomalous high-pressure and SAT warming over the subtropical North Atlantic via the westward propagation of Rossby waves. The northeasterlies at the southeast side of the anomalous high can transport more water vapor from the North Atlantic to eastern Pacific, resulting in water vapor divergence over the North Atlantic, which reduces the precipitation and increases the SSD in the North Atlantic. The NADW formation is thus enhanced, so does the AMOC. The opposite conditions are occurred under TP cooling, that is, the TP cooling is associated with anomalous low-pressure, SAT cooling, water vapor convergence, lighter surface waters over the North Atlantic, which lead to weakened AMOC. The relationship between TP's thermal condition and AMOC is quite robust and is confirmed by our albedo forcing experiments with different magnitude. The correlation coefficient between AMOC and SAT over TP is 0.97 (Figure S12 in Supporting Information S1). The atmospheric response to TP warming in our study is consistent with the results from Lu et al. (2018), in which they conduct higher-resolution simulation with fully coupled configuration and find that the TP warming results in SAT warming over the North Atlantic, as well as more water vapor transporting from North Atlantic to Pacific. Although highly idealized, this work provides detailed processes regarding how the TP surface thermal conditions affect the AMOC, which is very difficult to be extracted from such a complex climate system and is rarely addressed in previous works.

Here, we show that the TP heating effect on AMOC is through the adjustment of inter-basin moisture transport. Atmospheric moisture transport from the Atlantic to the Pacific basin plays a vital role in regulating the salinity of North Atlantic and thus the strength of the AMOC (Dey & Döös, 2020), which has been validated in previous works that focus on the AMOC response to the TP uplift, that is, Yang and Wen (2020), and Su et al. (2018). However, simply removing the TP in previous works cannot understand the mechanisms clearly (Vavrus & Kutzbach, 2002). We show that the mechanical and thermal effect induced by the TP uplift during geological time cause an opposite behavior of AMOC evolution (Yang & Wen 2020 and our results). For future climate, the increased CO_2 concentration will introduce a consistent warming in the world with continent warming stronger than ocean warming. The TP region will experience 7°C warmer condition at the end of 21st century under RCP8.5 (Figure S13 in Supporting Information S1; Su et al., 2013). Moreover, the hydrological cycle is enhanced, with wet regions getting wetter and dry regions getting drier (Figure S13 in Supporting Information S1; Held & Soden, 2006). The AMOC is consistently projected to weaken under global warming (Liu et al., 2020; Roberts et al., 2020) due to hydrological response and sea ice decline (Sévellec et al., 2017). According to our estimation, the 7°C of SAT warming over the TP alone can lead to about 10% AMOC enhancement (Figure S13 in Supporting Information S1) by generating water vapor divergence out of the deepwater formation region, competing the effects of the global warming that thermally cause more precipitation over the deepwater formation region. Bearing in mind that this work does not suggest that the AMOC will become stronger under global warming due to TP warming. Indead, the global warming effect dominates over the TP heating effect and will finally lead to the weakened AMOC at the end of 21 century. This work provides clues that the AMOC variation under climate change can be affected by not only local process, that is, sea ice or ice sheet, but also remote forcing from the Asian regions.

These idealized simulations allow us to see signals triggered by TP thermal forcing to reach at North Atlantic clearly. However, our study still has some limitations. For example, the thermal structure over the TP has strong seasonal and spatial features (Hua et al., 2018) that have not been considered in our experiment design, which may trigger discrepancies in AMOC response. Due to the computation resource, the lower resolution (T31_g37) are used for fully coupled simulations. This resolution will underestimate the deep-water formation over the Labrador Sea (Wen and Yang, 2020; Yang & Wen, 2020) and may also underestimate the AMOC response. In addition, the sea ice margin is more extensive compared to observations. Moreover, our experiments are performed with CO_2 concentration set at preindustrial level, whereas it was much higher under global warming. Due to these limitations, the conclusions drawn in this study may be model-dependent. Studies using more coupled models, with more deliberately designed thermal forcing experiments, are still extremely needed.

Data Availability Statement

Reanalysis data can be found on official websites. The model data to produce the main figures in this paper can be found at: https://doi.org/10.5281/zenodo.5215899. The data of CMIP5 models can be downloaded at https://esgf-node.llnl.gov/search/cmip5/. The precipitation data of GPCP and other variables from NCEP-DOE Reanalysis II can be found at https://psl.noaa.gov/data/gridded/.

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