Investigating the Role of the Tibetan Plateau in the Formation of Atlantic Meridional Overturning Circulation

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ABSTRACT

The Tibetan Plateau (TP) over the Eurasian continent has significant effects on both regional and global climate. It can even affect the remote Atlantic meridional overturning circulation (AMOC), as shown in this study. Through coupled modeling experiments, we demonstrate that removing the TP immediately weakens the meridional wind over East Asia, resulting in stronger westerlies in the midlatitudes. The stronger westerlies enhance the southward Ekman flow and surface latent and sensible heat losses in the subpolar North Atlantic, cooling the surface ocean and leading to stronger North Atlantic deep-water formation and stronger AMOC during the first few decades after the TP removal. At the same time, accompanying the weakened trade winds in the tropical Pacific, more moisture is transported from the tropical Pacific to the North Atlantic, freshening the surface ocean and triggering a weakening of the AMOC. The AMOC weakening in turn results in southward expansion and melting of sea ice, providing more freshwater to the North Atlantic, which furthers the weakening of the AMOC. The positive feedback between the AMOC and sea ice eventually leads to AMOC shutdown. We illustrate that there would be no AMOC without the TP. These results call for a revisiting of how ocean circulation and global climate may have responded to the TP uplift and other tectonic changes on the geological time scale.

1. Introduction

The Tibetan Plateau (TP), with an average elevation of more than 4000 m above sea level and a total area of 2.3 million square kilometers, is one of the most prominent features on Earth. The significant uplift of the TP can be traced back to 10–8 Ma (million years ago) (Harrison et al. 1992; Molnar et al. 1993). Paleoclimatic data suggest that the TP uplift might have resulted in a progressively drier winter over the northern Great Plains and the Eurasian interior, and an intensified winter westerly jet in the Northern Hemisphere (NH) during the past 10 Ma (Ruddiman and Kutzbach 1989; Kutzbach et al. 1993; An et al. 2001). Records from the North Pacific Ocean Drilling Program show a major dust peak about 8–7 Ma, which is thought to be due to the accumulated wind-blown dust from Asia accompanying the TP uplift (Rea et al. 1998). By attenuating the

downward sunlight and increasing the nucleus of condensation for atmospheric water vapor, the accumulated dust over the Pacific could have resulted in the cooling and freshening of the North Pacific and the weakening of the Pacific meridional overturning circulation (PMOC; Rea et al. 1998). Model results show that the elevation increase along the TP and Himalayas is sufficient to alter the thermally forced circulation and establish strong continent-scale summer and winter monsoons and central Asian aridity (An et al. 2001). The TP uplift may also have influenced global climate by accelerating rock weathering and withdrawing carbon dioxide from the atmosphere (Ruddiman et al. 1997).

The presence of the TP acts as a huge “heat engine” in the boreal summer and exerts tremendous impact not only on Asian monsoon system but also on North America’s weather and climate through planetary waves (e.g., Hoskins and Karoly 1981; Held and Ting 1990; Zhou et al. 2009; Wu et al. 2012; Liu et al. 2013). The TP effects on Asian weather and climate have been of particular concerns to Chinese scientists (Zhou et al. 2009). For example, scientists generally agree that the heating effect over the TP in summer drives the thermal circulation of

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the South Asian summer monsoon (Chou and Neelin 2001). The TP acts as a “sensible heat pump,” driving local ascending air that causes the low-level moist air south of the plateau to converge and condense, producing latent heating that in turn drives the large-scale monsoon circulation (e.g., Wang 2006; Wu et al. 2012). Both observations and numerical experiments suggest that thermal forcing controls the Asian summer monsoon system, whereas large-scale orographic mechanical forcing is not essential (Wu et al. 2012). The thermal forcing of the TP can also influence remote areas, such as the western Pacific subtropical high (Wu et al. 1999; Rodwell and Hoskins 2001; Yanai and Wu 2006; Boos and Kuang 2010; Wu et al. 2012; Lee et al. 2013; Duan et al. 2017) and the upper-level jet over the North Pacific (Lee et al. 2013). Together, they intensify the East Asian winter monsoon circulation and weaken storm-track activities. A study using recent reanalysis data suggested a link between the interannual variability of spring atmospheric heating over the TP and the early-spring tripolar pattern of sea surface temperature (SST) anomalies over the North Atlantic (Cui et al. 2015).

In the last two decades, the development of coupled atmosphere–ocean models has led modeling groups to revisit the effect of continental orography on ocean circulation (Kitoh 1997, 2002; Fallah et al. 2016; Maffre et al. 2018). Many modeling results show that the present-day configuration of mountains is an important background to the modern-state ocean thermohaline circulation (i.e., deep-water formation in the North Atlantic); such deep-water formation would have occurred in the Pacific in a flat world (Schmittner et al. 2011; Sinha et al. 2012). A coupled modeling study of Schmittner et al. (2011) showed that higher mountains reduce water vapor transport from the North Pacific to the Atlantic Ocean, contributing to increased (decreased) salinity and deep-water formation in the Atlantic (Pacific). A recent study by Maffre et al. (2018) showed that in a world with a globally flat continent, there would be strong westward freshwater transfer across Africa, accompanying the collapse of the Asian summer monsoon, which is critical to the freshening of the Atlantic and the increased salt content in the Pacific. However, these studies did not pinpoint the individual roles of different mountains in shaping the modern-day ocean circulations. Although the equilibrium responses of ocean circulation to the orographic forcing are more or less consistent among different studies, the transient responses of ocean circulation to the orographic forcing have not been investigated thoroughly.

The Atlantic meridional overturning circulation (AMOC) is one of the key elements of the global climate system. It is commonly recognized that the AMOC is

sustained by the North Atlantic Deep Water (NADW) formation (Delworth et al. 1993; Delworth and Mann 2000; Latif et al. 2004; Swingedouw et al. 2009). The strong NADW formation has persisted for the past several million years, which is thought to be maintained by strong net evaporation over the Atlantic and northward saline water transport (Ferreira et al. 2018). Paleoclimatic studies suggest that the AMOC first appeared about 12 Ma and was fully established in the late Pliocene (4–3 Ma), which occurred under a basin configuration similar to that of today (Ferreira et al. 2018, and references therein). The significant uplift of the TP occurred about 10–8 Ma (Harrison et al. 1992; Molnar et al. 1993), whose timing was close to that of AMOC establishment. In contrast, the Rocky Mountains had fully developed by 45 Ma, which was far earlier than the AMOC appearance. Therefore, the role of the TP in the formation of modern AMOC deserves an in-depth investigation, which is our focus in this study.

A fully coupled climate model is used here to study the TP effect on the AMOC. Through sensitivity experiments with and without the TP, we demonstrate that removing the TP would eventually result in AMOC collapse. The evolution of the AMOC shows a stronger AMOC during the first few decades after the TP removal, and then a weakening AMOC. The atmosphere processes are important during the transient stage of the AMOC change. Without the TP, there would be stronger westerlies in the midlatitudes, which enhance southward Ekman flow and surface latent and sensible heat losses in the subpolar North Atlantic, leading to surface ocean cooling that enhances the NADW formation and results in a stronger AMOC. At the same time, accompanying weakened trade winds in the tropical Pacific, more water vapor is transported from the central tropical Pacific to the North Atlantic via the atmosphere river, freshening the upper ocean of the North Atlantic and triggering a weakening of the AMOC. During the later stage of the AMOC evolution, the positive feedback between the AMOC and sea ice eventually leads to AMOC shutdown.

This study focuses on the transient evolution of the AMOC change. We try to answer the following questions: How do the atmospheric changes induced by the TP removal affect the North Atlantic? Why does the NADW formation change with time? And what is the role of sea ice? This paper is organized as follows. We introduce the coupled model and experiments in section 2. The transient and equilibrium responses of the North Atlantic are shown in section 3. The mechanisms for AMOC changes are discussed in section 4. The atmospheric responses are described in section 5. A summary and discussion are given in section 6.

2. Model and experiments

The model used in this study is the Community Earth System Model (CESM1.0) of the National Center for Atmospheric Research (NCAR). CESM is a fully coupled global climate model that provides state-of-the-art simulations of Earth's past, present, and future climate states (<http://www2.cesm.ucar.edu/>). It consists of five components and one coupler: the Community Atmosphere Model (CAM5; [Park et al. 2014](#)), the Community Land Model (CLM4; [Lawrence et al. 2012](#)), the Community Ice Code (CICE4; [Hunke and Lipscomb 2010](#)), the Parallel Ocean Program (POP2; [Smith and Gregory 2009](#)), the Community Ice Sheet Model (Glimmer-CISM), and the CESM Coupler (CPL7). CESM1.0 has been widely used and validated by researchers in the community (e.g., [Yang and Dai 2015](#); [Yang et al. 2017](#)). The model grid employed in this study is T31_gx3v7. The CAM5 has 26 vertical levels, with the finite volume nominally $3.75^\circ \times 3.75^\circ$ in the horizontal. The CLM4 has the same horizontal resolution as the CAM5. The POP2 has 60 vertical levels, and a uniform 3.6° spacing in the zonal direction. In the meridional direction, the grid is nonuniformly spaced: it is 0.6° near the equator, gradually increasing to the maximum 3.4° at 35°N/S and then decreasing poleward. The CICE4 has the same horizontal grid as the POP2. No flux adjustments are used in CESM1.0.

To explore the role of the TP in the formation of the AMOC, two parallel experiments with different topography are carried out ([Fig. 1](#)). One is a 2400-yr control run and the other is a 400-yr experiment with the TP removed at the beginning. The control run starts from the rest with standard configuration and a preindustrial CO_2 concentration of 285 ppm. The model geometry, topography, and continents of the control run are realistic ([Fig. 1a](#)). This control run is named “Real.” The model climate in Real reaches quasi-equilibrium (QE) after 1000 years of integration ([Yang et al. 2015](#)). The experiment without the TP, named “NoTibet,” starts from year 2001 of Real and is integrated for 400 years, with the topography of the TP reset as 50 m above mean sea level ([Fig. 1b](#)). To minimize the effect of internal variability on the transient change of the AMOC, we conducted 10 ensemble runs. These experiments were integrated for 100 years each, having exactly the same configuration as NoTibet but starting from different initial state of Real in year 2001. The AMOC change due to the TP removal is obtained by subtracting the results of Real from those of NoTibet. (Alternatively, the AMOC change due to the presence of the TP is obtained by subtracting the results of NoTibet from those of Real.) The AMOC change is quantified by the AMOC

FIG. 1. Topography configuration (m) in coupled experiments: (a) the realistic topography used in the control simulation (Real) and (b) the modified topography without the Tibetan Plateau used in the experiment NoTibet.

index, which is defined as the maximum streamfunction in the range of 0° – 10°C over 20° – 70°N in the Atlantic. For the convenience of discussion, two stages are defined according to the evolution of the AMOC index. Stage I is from model year 10 to year 50, representing the earlier fast change stage of the AMOC in NoTibet. Stage II is from model year 300 to year 400, representing the QE stage of the AMOC in NoTibet.

3. Changes in the Atlantic

a. Responses of AMOC

Removing the TP can affect the AMOC significantly. The transient change of the AMOC shows a slight intensification in the first few decades after the TP removal and then a significant decline. [Figure 2](#) shows the AMOC index and its spatial patterns in both stages. The AMOC is enhanced during the first 50 years, reaching its peak value in about 30 years (enhanced by 15%). The stronger AMOC is validated by 10 ensemble runs (gray curves in [Fig. 2a](#)), ensuring that it is a robust climate response rather than that caused by internal variability of the climate system. After 50 years of the TP removal, the AMOC shows a roughly linear decline for about 200

FIG. 2. (a) Temporal evolution of percentage change in Atlantic meridional overturning circulation (AMOC), with gray curves representing results from 10 ensemble runs. The AMOC index is defined as the maximum streamfunction in the range of 0° – 10° C over 20° – 70° N in the Atlantic. (b) Temporal changes in sea surface salinity (SSS; psu; blue), sea surface temperature (SST; $^{\circ}$ C; red), and sea surface density (SSD; kg m^{-3} ; black). SSD changes due to SSS and SST are plotted as dashed blue and dashed red curves, respectively. The y axis on the left is for the density change. The red and blue y axes on the right side are for SST and SSS, respectively. All variables are averaged over the North Atlantic Deep Water (NADW) formation region, which is outlined by the gray contour in Fig. 3c. (c) AMOC pattern (Sv) in Real (black contour) and its changes (shading) in stage I. (d) As in (c), but for stage II.

years. It is finally weakened by more than 80% and practically shut down in about 300 years. The curves in Fig. 2a are normalized by their mean values in Real and demonstrate percentage changes of the AMOC. The spatial patterns of the AMOC changes averaged in stage I and stage II are plotted in Figs. 2c and 2d, respectively. The patterns exhibit a weak positive anomaly in stage I, and a significant weakening of its downward mass transport in the subpolar North Atlantic in stage II, that is, a collapse of the AMOC. Note that the AMOC consists of two parts: the wind-driven subtropical cell (STC) in the tropical Atlantic and the thermohaline circulation (Weaver et al. 1993). The wind-driven STC, denoted by the streamfunction contours in the upper 200 m of the tropics (30°S – 30°N) (Figs. 2c,d), is nearly unaffected by the TP removal. This suggests that the major influence of the TP is on the thermohaline branch of the AMOC. The model experiments imply a critical role of the TP in maintaining the AMOC in modern-day climate.

b. Responses of ocean buoyancy

As the first step to understand the AMOC change, the temporal evolution of surface buoyancy change averaged in the NADW region (defined in Fig. 3c) is

examined in Fig. 2b. The sea surface density (SSD) increases during the first 50 years and then decreases linearly (black curve), consistent with the change of the AMOC index (Fig. 2a). The SSD change consists of SST-induced change (dashed red) and sea surface salinity (SSS)-induced change (dashed blue). Note that the surface ocean keeps cooling in the whole 400 years (red curve), which always increases the SSD (dashed red). In contrast, the SSS increases first and then decreases (blue curve), followed closely by the SSD change (dashed blue and black curves). Based on Fig. 2b, we quantify that the increase of SSD during stage I is contributed by surface cooling (30%) and surface salinization (70%), while the SSD decrease later is totally contributed by surface freshening. It clearly illustrates that the weakening of the AMOC should be attributed to the freshwater increase in the NADW region.

The definition of the NADW region in this work is based on the pattern of March mixed layer depth (MLD). The site of the deepest vertical mixing and convection can be found in March MLD (Brady and Otto-Bliesner 2011). Figure 3 shows the March MLD in Real, which is calculated following Large et al. (1997), and its change in NoTibet (with respect to Real). In Real, the maximum mean MLD is located in the Greenland–Iceland–Norwegian

FIG. 3. (a) Climatological March mixed layer depth (MLD; m) in Real and its changes in (b) stage I and (c) stage II of NoTibet. The MLD is defined the same way as in Large et al. (1997). In (b) and (c), the thick gray contour denotes zero MLD change within 40° – 70° N, 60° W– 20° E. The gray contour in (c) roughly outlines the NADW region, which is used in Fig. 2b and will be used later for term balance analyses.

(GIN) seas (Fig. 3a), which is largely consistent with the findings of other modeling studies and observations (Yeager and Danabasoglu 2014; Yang et al. 2016; Lozier et al. 2019). In NoTibet, the MLD becomes deeper in stage I (Fig. 3b) and shallower in stage II (Fig. 3c), signaling the enhanced and weakened NADW formation, respectively. Consistent with the AMOC collapse in stage II, the MLD in the GIN seas is reduced by more than 200 m (Fig. 3c). The NADW region used in Fig. 2b is defined as the region where the MLD becomes shallower in stage II of NoTibet (Fig. 3c),

enclosed by the gray zero contour of the MLD change. (Later in Figs. 6 and 8, all the variables are averaged within this NADW region.)

Figure 4 shows the horizontal patterns of buoyancy change in the North Atlantic in both stages. In stage I, the SST change has a tripolar structure (Fig. 4a): a significant cooling (more than 2°C) occurs in the mid-latitudes between 40° and 60° N, saddled by two warming regions located in the GIN seas and the tropics, respectively. The SSS change has an east–west dipole structure (Fig. 4b), with significant salinization in the Labrador Sea and south of the Greenland Sea and a weak freshening in the eastern North Atlantic. The combined effect of changes in SST and SSS results in SSD increase (Fig. 4c), particularly in the NADW region; this increases the MLD (Fig. 3b) and thus the deep-water formation, leading to a stronger AMOC in stage I. In stage II, the QE changes in SST, SSS, and SSD show rather simple structures (Figs. 4d–f), that is, significant cooling, freshening, and thus a lighter surface ocean in the entire North Atlantic, consistent with the “off” state of the AMOC. It is clear that the SSD changes in the North Atlantic in both stages are mostly determined by SSS change. The mechanisms for the buoyancy changes will be discussed in section 4.

The upper-ocean change in the North Atlantic suggests a profound impact of the TP (Fig. 5). In stage I, the salinity change is weak and confined to the North Atlantic and southern tropical Atlantic (Fig. 5a). The North Atlantic shows salinization in the top 100 m and freshening down to 1000 m, a baroclinic structure caused by enhanced upward salinity diffusion (Fig. 8b in section 4) (Oka et al. 2001; Zhang and Steele 2007). The upper-ocean temperature shows a uniform cooling over 40° – 60° N (Fig. 5b), triggered mainly by surface heat loss (Fig. 6). The combined effect of salinity and temperature changes leads to an increased density in the North Atlantic and decreased density in the tropical upper ocean (Fig. 5c). In stage II, the QE changes in salinity and temperature show bipolar seesaw structures in the Atlantic basin, namely strong freshening and cooling in the North Atlantic and weak salinization and warming in the South Atlantic (Figs. 5d,e). This structure has been well recognized in many previous studies (e.g., Manabe and Stouffer 1995; Zhang and Delworth 2005; Stocker et al. 2007; Stouffer et al. 2007; Oppo and Curry 2012; Yang et al. 2013) and occurs in company with the shutdown of the AMOC. The upper-ocean density in the entire Atlantic is, as expected, decreased (Fig. 5f), due to freshening in the North Atlantic and warming in the tropical Atlantic. Once again, we can see that the salinity change dominates the density change in the high latitudes (because thermal expansion becomes very small in the low

FIG. 4. Changes in (a) SST ($^{\circ}\text{C}$), (b) SSS (psu), and (c) SSD (kg m^{-3}) averaged over stage I. (d)–(f) As in (a)–(c), but for stage II. The white contours in (a) and (d) denote the SSD change induced by SST change, while those in (b) and (e) denote the SSD change induced by SSS change. Dashed contours are for negative change, and solid contours are for positive change.

temperature regime), while both temperature and salinity changes contribute to the density change in the tropics.

4. Mechanisms for ocean buoyancy change

To gain insight into the AMOC change, we perform term balance analyses on SST and SSS equations in this section. All the terms except the source terms are calculated online (during the model integration). The SST and SSS equations can be written as follows:

$$\frac{\partial T}{\partial t} = - \left(u \frac{\partial T}{\partial x} + v \frac{\partial T}{\partial y} + w \frac{\partial T}{\partial z} \right) + A_H \nabla^2 T + \frac{\partial}{\partial z} \kappa \frac{\partial T}{\partial z} + \text{Heatflux}, \quad (1)$$

$$\frac{\partial S}{\partial t} = - \left(u \frac{\partial S}{\partial x} + v \frac{\partial S}{\partial y} + w \frac{\partial S}{\partial z} \right) + A_H \nabla^2 S + \frac{\partial}{\partial z} \kappa \frac{\partial S}{\partial z} + Q_s, \quad (2)$$

where

$$\text{Heatflux} = \text{SW} + \text{LW} + \text{SH} + \text{LH}, \quad (3)$$

$$Q_s = \text{VSF}_{\text{EMP}} + \text{VSF}_{\text{ice}}, \quad \text{and} \quad (4)$$

$$\text{VSF}_{\text{ice}} = \text{MeltF} + \text{IoffF} + \text{SaltF}. \quad (5)$$

The terms on the right-hand side of Eq. (1) are temperature advection, horizontal diffusion, vertical diffusion, and the net surface heat flux, respectively. The source term Heatflux in Eq. (1) consists of shortwave radiative flux (SW), longwave radiative flux (LW), sensible heat flux (SH), and latent heat flux (LH) [Eq. (3)]. The terms on the right-hand side of Eq. (2) are salinity advection, horizontal diffusion, vertical diffusion, and the source term. Also, u , v , and w are zonal, meridional, and vertical velocities, respectively; A_H is the horizontal diffusivity coefficient; and κ is vertical diffusivity coefficient, which is calculated using the KPP model in CESM. The KPP model is detailed in Large et al. (1994). The terms in Eqs. (1) and (2) are calculated over the top grid cell (with a depth of 30 m) of the ocean domain. The source term in Eq. (2) includes the virtual salt flux due to evaporation, precipitation, and river runoff (VSF_{EMP}) and that due to sea ice formation (VSF_{ice}) [Eq. (4)]. Sea ice formation consists of sea ice generation or melting in

FIG. 5. Depth–latitude section of changes in Atlantic (left) salinity (psu), (center) temperature ($^{\circ}\text{C}$), and (right) potential density (kg m^{-3}) averaged over (a)–(c) stage I and (d)–(f) stage II.

the high latitudes (MeltF), continental ice runoff flux (IoffF), and salt flux (SaltF) due to the salinity gradient between ice and water [Eq. (5)]. Usually, IoffF and SaltF are less than 1% of the total MeltF, and can be neglected. Here, we want to emphasize that the vertical diffusion terms of temperature and salinity include the effects of vertical convection and mixing. The coupled model used in this study is incapable of resolving convection, which is parameterized as part of vertical diffusion in the ocean model.

a. Mechanism for SST change

Figure 6 shows the time series of the terms on the right-hand side of Eq. (1), averaged over the upper 30-m ocean in the NADW region (defined in Fig. 3c). As shown before, the SST changes rapidly after the TP removal, showing a cooling trend through the whole 400 years (Fig. 2b). In stage I, the SST cooling is caused by more net surface heat loss (black curve in Fig. 6b) as well as anomalous temperature advection (blue curve in Fig. 6a). In stage II, the SST cooling is maintained by reduced incoming SW (red curve in Fig. 6b) and by weakened vertical mixing and diffusion (gray curve in Fig. 6a), as well as anomalous temperature advection (blue curve in Fig. 6a). A probe into the heat flux components reveals that both LH and SH (green and orange curves in Fig. 6b) control the SST cooling in stage I. Although the change in temperature advection always contributes to SST cooling, the sources of advection water are different in different stages. In stage I, it is the enhanced cold-water advection from the subpolar ocean

that cools the NADW region, whereas in stage II it is the weakened warm-water advection from the south that helps to maintain the cooling in the North Atlantic. Details are presented in Fig. 7.

Figure 7a shows the enhanced LH and SH losses in the NADW region and GIN seas, which lead to more than $3^{\circ}\text{C yr}^{-1}$ cooling there in stage I. The southward cold-water advection from the subpolar ocean, carried mainly by the enhanced Eastern Greenland Current and Labrador Current, contributes about $3^{\circ}\text{C yr}^{-1}$ to the surface cooling in the subtropical Atlantic (Fig. 7b). Changes in these factors are all attributed to the intensified westerly wind (Fig. 7c). The stronger wind has three consequences. First, it directly enhances the LH and SH losses. Second, it increases the surface evaporation and thus the LH loss. Third, it enhances the southward Ekman flow (Fig. 7b), which brings more cold water southward to lower latitudes [this is consistent with the study of Cessi (2018)]. For stage I, we focus on atmosphere dynamics and emphasize that the change in the surface wind plays the most important role in the North Atlantic surface cooling. The cause for the intensified westerly wind will be discussed in section 5a.

Figure 7d shows the reduced incoming SW in the North Atlantic in stage II, which is the dominant cooling effect as shown in Fig. 6a. The solid and dashed red curves in Fig. 7d represent the sea ice margin in Real and NoTibet, respectively. The ocean between these two curves is covered partly by sea ice in the absence of the TP, which increases surface albedo significantly and reduces SW absorption. In comparison, the reduced SW

FIG. 6. Temporal evolutions of terms in (a) SST equation ($^{\circ}\text{C yr}^{-1}$) and (b) surface heat flux (W m^{-2} ; positive for downward anomaly) averaged in the NADW region. In (a), the gray curve is for vertical diffusion; blue is for horizontal advection; purple is for horizontal diffusion; green is for the sum of latent heat flux (LH), sensible heat flux (SH), and longwave radiative flux (LW); and red is for net downward shortwave radiative flux (SW). In (b), the red curve is for SW; blue is for LW; orange and green are for SH and LH, respectively; and black is for net surface heat flux (NET). All curves are for changes in NoTibet, with respect to Real.

south of the sea ice region is attributed mainly to the increased low clouds (Fig. 7f). Note that the sea ice margin (dashed red curve in Figs. 7d,f) in the absence of the TP coincides well with the zero contour of low-cloud change (Fig. 7f). The increased low clouds in the subtropical Atlantic are in turn due to the colder ocean. Here, the low clouds are defined as the total clouds below the level of the 680-hPa isobar, as defined in Zhang et al. (2010). In summary, the SW over the North Atlantic is reduced by 21 W m^{-2} , in which 6 W m^{-2} is due to the increased low clouds south of the sea ice margin and 15 W m^{-2} is due to the increased sea ice reflection north of the sea ice margin.

The secondary cooling effect on the ocean comes from anomalous temperature advection (Fig. 7e). As mentioned before, different from that in stage I, it is the weakened warm-water advection from the south that helps maintain the cooling in the subtropical Atlantic. Figure 7e shows the anomalous southwestward surface current in the subtropics (i.e., the weakened North Atlantic Current), which is particularly clear along the

pathway of the Gulf Stream and its extension, manifesting the weakened surface branch of the AMOC. In stage II, we focus on ocean dynamics and sea ice change. From stage I to stage II, the gradual decline of the AMOC is the key to the changes discussed above. Why the AMOC reverses to the “off” state in stage II will be answered in the following section.

b. Mechanism for SSS change

The SSS change plays a more important role in the changes of SSD and AMOC. Figure 8a shows that the enhanced vertical salinity diffusion (solid gray curve) is the main factor that increases the SSS in stage I. The response of salinity diffusion is very fast, contributing about 0.3 psu yr^{-1} to the SSS increase in the beginning. The horizontal diffusion and the net freshwater flux due to evaporation minus precipitation (EMP) have freshening effects on the surface ocean. The effects of sea ice melting and salinity advection are small during this stage. After stage I, the EMP change always reduces surface salinity. Vertical diffusion gradually becomes a freshening effect on upper-ocean salinity. Sea ice melting becomes a dominant factor for the surface freshening in 200 years. Both horizontal diffusion and salinity advection contribute to salinity increase instead. In the following discussion, we focus on vertical diffusion, EMP, and sea ice melting, since each of these three factors plays a deterministic role in different stages of the AMOC evolution after the TP removal.

The vertical salinity diffusion in the NADW formation region changes its role from having a salinization effect to having a freshening effect on the surface ocean. This can be attributed to the changes in the surface wind and surface freshwater flux. In Eqs. (1) and (2), the vertical diffusivity κ in CESM is calculated using the KPP model, which has included the effects of surface wind and buoyancy flux forcings on the vertical mixing and diffusion in the mixed layer (Large et al. 1994). Previous studies (e.g., Oka et al. 2001; Yang et al. 2016) have also suggested that wind stress is crucial in maintaining the AMOC, through sustaining vertical convection and stirring up the upper ocean, as well as the Ekman pumping in the subpolar gyre. The intensified westerly wind immediately enhances vertical mixing and diffusion (Fig. 8c), resulting in enhanced vertical salinity diffusion (Fig. 8b), which brings more saline water upward to the upper ocean in stage I. Later on, although the westerly wind is roughly unchanged compared to that in stage I, the accumulated surface freshwater flux ($\text{EMP} < 0$) would finally overcome the wind effect, weakening vertical mixing and diffusion (Fig. 8c), freshening the surface ocean, and triggering a decline of the AMOC. The vertical diffusivity in the surface ocean

FIG. 7. Changes in (a) temperature tendency ($^{\circ}\text{C yr}^{-1}$) induced by LH and SH, (b) ocean surface current (vector; cm s^{-1}) and temperature advection (shading; $^{\circ}\text{C yr}^{-1}$), (c) surface wind stress (vector; dyne cm^{-2}) and percentage change of low cloud (shading; %) in stage I, and (d) temperature tendency ($^{\circ}\text{C yr}^{-1}$) induced by SW in stage II. Solid and dashed red curves show the sea ice margin in Real and NoTibet, respectively. In (a) and (d), a positive (negative) value represents heat gain (loss) by the ocean. The sea ice margin is defined by the 15% sea ice fraction in the North Atlantic, which is used throughout the paper. (e), (f) As in (b) and (c), but for stage II.

(Fig. 8c) actually determines the vertical salinity diffusion (Fig. 8b). The source of the surface freshwater flux will be discussed in section 5b.

Accompanying the weakening of the AMOC, the sea ice in the subpolar ocean expands southward gradually, providing additional freshwater to the NADW region, which in turn leads to further weakening of the AMOC. Vertical salinity diffusion is thus weakened significantly and becomes a freshening factor to the upper ocean (Fig. 8b). Figure 9a shows the temporal evolution of sea ice coverage in the North Atlantic. The sea ice shows a significant expansion during years 150–250 after the TP removal, consistent with the great amount of freshwater flux into the North Atlantic (orange curve, Fig. 8a). Figures 9b and 9c further show the changes in sea ice velocity and formation as well as sea ice margin. Blue color in Figs. 9b and 9c represents negative sea ice formation (i.e., sea ice melting). The curves in Figs. 9b and 9c illustrate the locations of the sea ice margin. In stage I, the sea ice margin retreats slightly in the GIN seas and expands southward in the Labrador Sea (dashed red curve, Fig. 9b). The former is due to the enhanced northward

heat transport caused by the stronger AMOC, and the latter is due to the intensified westerlies that drive the sea ice southward, illustrated by the sea ice velocity in Fig. 9a. The MLD becomes deeper (Fig. 3b), in association with stronger surface wind and colder SST, signaling the strengthening of convection, vertical mixing, and diffusion, and thus the deep-water formation (Brady and Otto-Bliesner 2011). After stage I, the sea ice margin expands southward and eastward toward the region of the Gulf Stream extension, and roughly reaches QE in 300 years (dashed red curve, Fig. 9c), accompanied by much sea ice melting (blue color, Fig. 9c). The MLD becomes shallower (Fig. 3c), signaling the weakening of convection, vertical mixing, and diffusion, and thus further weakening of the AMOC.

The decline of the AMOC appears to be triggered by the surface freshwater flux related to EMP (Fig. 8a). This will be discussed further in section 5b. The sea ice melting takes effect later than the AMOC weakening (Fig. 9 vs Fig. 2a) and therefore is a result, not a driving factor, of the AMOC weakening. However, based on the evolutions of the AMOC (Fig. 2a) and meltwater provided by sea ice

FIG. 8. (a) Temporal changes of the terms in SSS equation (psu yr⁻¹). The gray curve is for vertical diffusion; blue is for horizontal advection; purple is for horizontal diffusion; red is for virtual salt flux (VSF) due to evaporation minus precipitation (EMP); orange is for VSF due to sea ice melt/formation; and dashed black is for the sum of horizontal and vertical diffusion. (b) Temporal change in vertical salinity diffusion $\kappa \partial^2 S / \partial z^2$ (psu yr⁻¹). (c) Temporal change in vertical diffusivity coefficient κ (cm² s⁻¹). All variables are for changes in NoTibet, with respect to Real.

(Fig. 8a), as well as the sea ice coverage and margin shown in Fig. 9, we can conclude that there is a positive feedback between the AMOC weakening and sea ice expansion/melting. This positive feedback finally shuts down the AMOC. A similar positive feedback between sea ice and AMOC has been shown in many previous studies (e.g., Brady and Otto-Bliesner 2011; Yang et al. 2016).

FIG. 9. (a) Temporal evolutions of sea ice coverage (10^6 km²) in the North Atlantic (40°–80°N) in Real (black curve) and NoTibet (blue curve). (b) Changes in sea ice formation (shading; psu yr⁻¹) and sea ice velocity (vector; cm s⁻¹) in stage I of NoTibet. Positive (negative) value means sea ice formation (melting). Solid and dashed red curves represent the sea ice margin in Real and NoTibet, respectively. (c) As in (b), but for stage II. In (c), orange, green, and dashed red curves show the sea ice margin in the 100th year, 200th year, and QE stage of NoTibet, respectively. Solid red curve represents the sea ice margin in Real.

The change in salinity advection in the NADW region tends to increase salinity, especially in stage II (blue curve, Fig. 8a). Many studies have shown that wind stress maintains the meridional salt transport (Oka et al. 2001) and that the salinity advection due to wind-driven circulation has a significant salinization effect in the high latitudes (Yang et al. 2016). In our study, the westerly wind becomes stronger after the TP removal, which intensifies the wind-driven circulation and thus the surface salinity advection.

FIG. 10. Changes in geopotential height (shading; m) and wind (vector; m s^{-1}) at (a),(c) 850 hPa and (b),(d) 500 hPa in NoTibet with respect to Real, for (top) stage I and (bottom) stage II. To better see the wave structure, the zonal-mean value of geopotential height has been removed.

5. Atmospheric responses

a. Stationary wave and teleconnection

Figure 10 shows the spatial patterns of atmospheric circulation change at 850 and 500 hPa in the two stages after the TP removal. The intensified westerlies can be clearly seen over the North Atlantic and North Pacific. It is a classic scenario that lowering the Northern Hemisphere (NH) orography can cause midlatitude westerlies to be more zonal (Manabe and Terpstra 1974). This is also true in our experiments. The intensified westerlies, accompanied by weakened meridional winds, suggest an enhanced annular mode, consistent with the finding of previous studies (Manabe and Terpstra 1974; Ruddiman and Kutzbach 1989; Thompson and Wallace 2001; Naiman et al. 2017). The changes of atmospheric circulation are roughly barotropic, based on similar changes at 850 and 500 hPa (Figs. 10a,c vs Figs. 10b,d). These atmospheric responses do not change much from stage I to stage II (Figs. 10a,b vs Figs. 10c,d), suggesting a short response time scale of atmospheric circulation to the TP removal. More importantly, it also implies that oceanic changes in the later years do not have considerable feedback to the atmosphere circulation. This is extremely important, since the separation of response time scales in atmosphere and ocean makes it easy to separate the effects of atmosphere dynamics and ocean dynamics on certain change in a model experiment.

The pattern change of atmospheric circulation shows clearly a planetary wave structure in the NH mid-high latitudes (Fig. 10). Removing the TP induces cyclonic

geopotential height anomalies to the north of the TP area and over the subpolar Atlantic, and anticyclonic anomalies to the south of the TP area and over the subpolar Pacific. The local dipole pattern around the TP region is the common feature in all topography perturbation experiments (e.g., Brayshaw et al. 2009; White et al. 2017), which can be understood by linear potential vorticity dynamics (Valdes and Hoskins 1991; White et al. 2017). The remote responses over the subpolar Pacific and Atlantic exhibit clearly a wave train structure, suggesting a northeastward propagation of the wave energy, which establishes a robust teleconnection between the perturbation over the TP area and the atmospheric circulation over the Pacific and Atlantic. The dynamics of this wave train can be well understood by using the classic planetary wave theory in a quasigeostrophic system (Hoskins and Karoly 1981; Hoskins and Ambrizzi 1993). Detailed atmospheric circulation changes are also discussed in Yang et al. (2019), which states that the teleconnection patterns shown in Fig. 10 agree well with those in previous studies (Zhao et al. 2007; Cui et al. 2015). The teleconnection between the North Atlantic and the Asian monsoon region has been confirmed by paleoclimate records (Wang et al. 2001, 2005; Liu et al. 2013), which are established mainly via atmospheric processes (Chiang and Bitz 2005; Broccoli et al. 2006; Zhao et al. 2007; Liu et al. 2013). The ocean process does not contribute to the teleconnection; instead, it responds to the teleconnection. In other words, the TP affects the ocean circulation and buoyancy fields via

FIG. 11. Changes in (a) surface wind speed (m s^{-1}) at 10-m height above mean sea level and (b) annual mean wave activity flux (vector; $\text{m}^2 \text{s}^{-2}$) and its divergence (shading; 10^{-6}m s^{-2}) at 850 hPa during stage I. In (b), positive (negative) value is for divergence (convergence). The situation during stage II is similar to that in stage I, and thus is not shown here.

atmospheric processes. Similar conclusions were presented in Fallah et al. (2016).

The anomalous cyclonic circulation over the subpolar Atlantic contributes greatly to the enhanced westerlies over the North Atlantic (Figs. 10a,c). Figure 11a shows the enhanced surface wind. The changes in wave activity flux (vector) and its divergence (shading) at 850 hPa due to the TP removal are plotted in Fig. 11b, to explain the dynamics of the enhanced westerlies over the North Atlantic. The eastward wave activity flux manifests an eastward energy propagation from the Pacific to the Atlantic. The divergence (positive value) of the wave flux over the North Atlantic suggests that the mean background westerlies are enhanced (Fig. 11b). Here, the calculation of the topography-induced wave flux follows the approach used in Takaya and Nakamura (1997, 2001); the Takaya–Nakamura stationary wave activity flux can be used to examine the horizontal energy propagation of stationary waves. The flux magnitude is related to the phase-independent Rossby wave amplitude following the wave's group velocity (Takaya and Nakamura 2001). In our calculations, the perturbations in the wave activity flux are the differences created by the TP (i.e., NoTibet minus Real). Therefore, the divergence (positive value) of the wave activity flux over the North Atlantic denotes where the wave perturbations are emitted and converted to the mean background flow (Qiao and Feng 2016). Similar work was conducted by White et al. (2017) to examine the effect of the Mongolian mountains on the Pacific wintertime atmospheric circulation.

b. Atmospheric moisture transport and EMP

The change of surface freshwater flux EMP over the North Atlantic freshens the upper ocean at a stable rate of about $0.05\text{--}0.10 \text{ psu yr}^{-1}$, as shown by red curve in Fig. 8a. This freshwater supply is very important, and the accumulated freshwater in the North Atlantic in the first

several decades triggers the weakening of the AMOC. For a steady state, the EMP across the ocean surface is equivalent to the vertically integrated moisture transport divergence $\nabla \cdot \mathbf{v}q$ over the entire atmosphere column, when neglecting the freshwater flux at the land surface and river runoff (Yang et al. 2015). We plot the atmospheric moisture transport $\mathbf{v}q$ (vector) and its convergence $-\nabla \cdot \mathbf{v}q$ (color shading) in Fig. 12 so that we can trace where the anomalous freshwater over the North Atlantic comes from (after the TP removal).

Removing the TP results in more freshwater relocating from the tropical Pacific to the North Atlantic (Fig. 12a). Since moisture transport $\mathbf{v}q$ is determined by both atmospheric circulation $\mathbf{v}$ and specific humidity q , the $\mathbf{v}q$ pattern can be different from the circulation pattern in Fig. 10a. Figure 12 shows the moisture

FIG. 12. Changes in (a) vertically integrated moisture transport $\rho_a \mathbf{v}q$ (vector; $\text{kg m}^{-1} \text{s}^{-1}$) and its convergence $[-\rho_a \nabla \cdot (\mathbf{v}q)]$; shading; $10^{-5} \text{kg m}^{-2} \text{s}^{-1}$] during stage I in NoTibet; $\rho_a = 1.29 \text{kg m}^{-3}$ is air density. (b) As in (a), but for stage II. Here, positive value is for convergence, which suggests freshwater from the atmosphere into the ocean.

FIG. 13. Schematic diagram summarizing the main processes after the TP removal. The upward (downward) arrows represent increase (decrease) of a quantity. NA: North Atlantic; SST and SSS: sea surface temperature and salinity; LH and SH: latent and sensible heat fluxes; AMOC: Atlantic meridional overturning circulation; MLD: mixed layer depth; HT: heat transport; PM_E: Precipitation minus evaporation.

transport pathway and the gain (or loss) of freshwater in both atmosphere and ocean. The atmospheric moisture convergence is plotted as positive (i.e., $-\nabla \cdot \mathbf{v}q > 0$), representing a loss of atmosphere freshwater to the ocean ($\text{EMP} < 0$, or $E < P$). The moisture divergence is plotted as negative, representing a gain of atmosphere freshwater from the ocean. Without the TP, more atmospheric moisture will be transported all the way from the central tropical Pacific to the North Atlantic, converging over the North Atlantic, leading to a fresher North Atlantic upper ocean. The atmosphere over the North Atlantic will thus become drier, which can be demonstrated by the reduced atmospheric specific humidity (figure not shown here; Yang et al. 2019). Here, we see once again that the atmospheric moisture transport and its convergence do not change much from stage I to stage II (Fig. 12a vs Fig. 12b), suggesting that oceanic changes in the later years do not have considerable feedback to the atmospheric moisture transport.

6. Summary and discussion

In this study, we investigate the role of the TP in the formation of the modern-day AMOC. Through sensitivity experiments with and without the TP, we demonstrate that removing the TP would eventually result in an AMOC collapse. The main processes that are responsible for the AMOC change are summarized in Fig. 13. The AMOC change can be divided into two stages: the fast transient stage during which the atmospheric processes dominate,

and the slow evolution stage in which the oceanic processes dominate. In response to the TP removal, a planetary wave train is excited immediately in the NH mid-high latitudes and an anomalous low pressure develops over the subpolar Atlantic. The latter leads to enhanced westerlies over the North Atlantic. In the fast transient stage, the oceanic responses to the enhanced westerlies include more LH and SH losses to the atmosphere, an anomalous southward cold-water advection from the subpolar ocean, and a deepening of the MLD. These factors cause a colder SST and an elevated SSS, which temporarily increase the deep-water formation and thus the AMOC. Since there is continuous moisture transport from the tropical Pacific to the North Atlantic, the accumulated freshwater in the North Atlantic will finally trigger the weakening of the AMOC. In the slow evolution stage, the weakening of the AMOC results in a southward expansion and melting of the sea ice from the subpolar region, which in turn further the weakening of the AMOC and eventually leads to AMOC collapse.

Our modeling results are qualitatively consistent with those in previous studies of Fallah et al. (2016), Maffre et al. (2018), and Su et al. (2018), that is, removing the TP will result in the AMOC collapse. However, the detailed processes that lead to the AMOC collapse are different in different studies. Fallah et al. (2016) emphasized the role of reduced northeastward heat advection from the North Atlantic in the weakening of the AMOC. Maffre et al. (2018) concluded that the westward freshwater transfer across Africa is critical to

the freshening of the Atlantic and thus the AMOC weakening. [Su et al. \(2018\)](#) showed the critical factor of northward moisture transport over the North Atlantic. These studies did not show the transient change in the AMOC. Our study shows an initial strengthening, followed by a decline of the AMOC, in response to the TP removal. Moreover, we emphasize that the atmospheric moisture relocation from the tropical Pacific to the North Atlantic is the key that triggers the weakening of the AMOC, and the positive feedback between the southward expansion of sea ice and AMOC leads to the AMOC shutdown. These mechanisms are different from those in previous studies. The discrepancies in mechanisms suggest the complexity of the TP in affecting global ocean circulations. More sensitivity experiments using different models are needed.

Although this is a highly idealized modeling study, this work helps explain the quantitative role of the TP in the real world. Our modeling results may have applications for paleoclimate study. For example, we show an anomalous southward cross-equatorial jet in the lower atmosphere and southward moisture transport over the Indian and western Pacific in the *absence* of the TP ([Figs. 10 and 12](#)). This provides a counterexample supporting the idea that the rapid uplift of the TP at 10–8 Ma might have pushed the establishment of the monsoon system in southeastern Asia ([Kutzbach et al. 1993](#); [An et al. 2001](#)). Moreover, geological evidence shows that the strong NADW formation was established about 10 Ma ([Woodruff and Savin 1989](#)), roughly in the same period when the rapid TP uplift occurred. Our results imply that in the *presence* of the TP, there would be enhanced atmospheric moisture transport from the North Atlantic to the tropical Pacific, which helps increase surface salinity and density in the North Atlantic, leading to strong NADW formation and thus the AMOC formation.

The results from the extreme topography experiments may not be comparable to any observed variability. However, a better understanding of detailed processes of the global impact of the TP can be fulfilled via these unrealistic experiments. The impact of TP uplift on global ocean circulation is difficult to study based on observational data, although there is evidence showing that the tectonic changes in the Miocene were associated with large-scale changes in the global ocean circulation ([Barker and Burrell 1977](#); [Kennett 1977](#)). The ocean circulation can be modified by surface wind stress and freshwater flux. These two factors will be changed along with the TP uplift. Often cited in support of this idea is the fact that the intensity and direction of wind and moisture transport will change in response to topography change ([Fallah et al. 2016](#); [Maffre et al. 2018](#)). Through extreme topography, we can see more clearly

the signals in both local and remote regions, and understand better the mechanisms in essence.

In this work, we show that removing the TP can lead to AMOC shutdown. We have also done experiments showing that with globally flattened continents, a sudden uplift of the TP can lead to the AMOC formation (figure not shown). However, the presence of the TP does not have to be a necessary condition for the existence of the AMOC. Many idealized coupled model experiments suggested that the deep meridional overturning circulation exists in the Atlantic because of its smaller width (e.g., [Ferreira et al. 2010](#); [Nilsson et al. 2013](#)), regardless of the continental topography used. By planting very thin meridional continents in an aquaplanet, [Ferreira et al. \(2010\)](#) showed that the small and large basins exhibit distinctive Atlantic-like and Pacific-like characteristics, respectively. The small basin is warmer, saltier, and denser at the surface than the large basin, and is the main site of deep-water formation with a deep overturning circulation and strong northward ocean heat transport. [Nilsson et al. \(2013\)](#) further showed that the southward extent of the land barrier can affect the deep-water formation because the length of meridional barrier controls the wind-driven Sverdrup circulation, and thus the interbasin salt transport. These idealized aquaplanet experiments suggest the fundamental roles of the basin geometry in World Ocean circulations. However, our work suggests that under modern-day basin geometry, the TP uplift may have affected the AMOC formation.

The conclusions drawn in this study are subject to model limitations, and may be model-dependent. For example, our model results show a relatively weak convection in the Labrador Sea ([Fig. 3](#)), when compared to observations and the observed convection in the GIN seas. In addition, the sea ice margin is more extensive compared to observations. These limitations may be related to model's horizontal resolution. The ocean model used in this paper cannot fully resolve the convection in the Labrador Sea. We have examined the convection site and sea ice margin in a high-resolution CESM control run. Strong convection occurs in both the Labrador Sea and GIN seas (figure not shown), consistent with observations and other modeling studies ([Yeager and Danabasoglu 2014](#); [Lozier et al. 2019](#)). In addition, our experiments are performed with constant atmospheric CO₂ concentration (285 ppm) at the pre-industrial level, whereas it was higher during the TP uplift phase in reality ([Lowenstein and Demicco 2006](#)). During the late Eocene, the climate conditions were featured with higher atmospheric CO₂ concentration (up to 1000 ppm) ([Lowenstein and Demicco 2006](#)), which could affect the sea ice to a great extent. The most

critical point people should bear in mind is that the climate changes in response to a sudden removal of the TP from the modern-day topography may not have to be consistent with the climate evolution with the gradual uplift of the TP from a paleo-topography viewpoint. In other words, this work unveils the critical role of the TP in the formation of the AMOC using an apagogical method. How the global meridional overturning circulations have evolved with the tectonic changes, particularly the gradual TP uplift at the geological time scale, remains uncertain. Studies using more coupled models, with more deliberately designed topography experiments, are still extremely needed.

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