Roles of the Rocky Mountains in the Atlantic and Pacific Meridional Overturning Circulations

RUI JIANG^a AND HAIJUN YANG^{b,c}

^a *Department of Atmospheric and Oceanic Sciences, School of Physics, Peking University, Beijing, China*

^b *Department of Atmospheric and Oceanic Sciences, Institute of Atmospheric Science, Fudan University, Shanghai, China*

^c *CMA-FDU Joint Laboratory of Marine Meteorology, Fudan University, Shanghai, China*

(Manuscript received 24 October 2020, in final form 20 May 2021)

ABSTRACT: The effect of the Rocky Mountains (RM) on meridional overturning circulations (MOCs) is investigated using a fully coupled climate model. Located between the Atlantic and Pacific Oceans, the RM is a major mountain chain in North America. The presence of RM plays an important role in atmospheric moisture transport between the two oceans. Adding the RM to a flat global continent (OnlyRocky) leads to a weakening of the atmospheric moisture transport from the North Pacific to the North Atlantic, which is consistent with previous findings. However, the simulation also shows more atmospheric moisture is transported from the tropical Pacific and Atlantic to the North Atlantic. The net effect of moisture transport leads to a slight freshening of the North Atlantic. The Atlantic MOC (AMOC) is hardly changed, but the Pacific MOC (PMOC) declines by 40% due to more moisture retained in the North Pacific. The sensitivity experiment of removing the RM from a realistic global topography (NoRocky) gives roughly opposite atmospheric changes to the OnlyRocky experiment. The AMOC in NoRocky declines slightly and then recovers, while the PMOC is nearly unchanged. The paired experiments conducted in this study demonstrate that the presence of the RM plays a trivial role in Northern Hemisphere deep-water formation.

KEYWORDS: Meridional overturning circulation; Topographic effects; Atmosphere-ocean interaction; Climate models

1. Introduction

The Rocky Mountains (RM) is a major mountain range located in western North America between the Pacific and Atlantic Oceans. The RM uplifted gradually from about 80 to 55 million years ago (Ma) during the Laramide orogeny, and reached the modern elevation about 45 Ma (Ferreira et al. 2018, and the references therein); it has a meridional extension of about 4800 km. Paleoclimatic studies suggest that the Atlantic meridional overturning circulation (AMOC) first appeared about 12 Ma and was fully established in the late Pliocene of about 4–3 Ma when the basin configuration was similar to that of the modern day. The RM uplift was far earlier than the AMOC appearance. It is commonly recognized that the AMOC is sustained by the North Atlantic deep-water (NADW) formation (Latif et al. 2004; Swingedouw et al. 2009) and the Ekman pumping in the Southern Ocean (Wunsch and Ferrari 2004; Nikurashin and Vallis 2012). There is a lack of paleoclimatic evidence to support a strong NADW at 45 Ma when the RM reached its modern-day height. Paleoclimatic evidence also suggests that the North Pacific deep-water (NPDW) formation nearly disappeared since 10 Ma (Ferreira et al. 2018). Due to its meridional extension, the RM has long been thought to have great influence on the zonal moisture transport between the North Pacific and North Atlantic (Kittel et al. 2002), and thus may play a role in

the evolutions of the AMOC and Pacific meridional overturning circulation (PMOC).

Many studies have shown that the continental topography has a profound effect on the formation of global ocean circulations (Schmittner et al. 2011; Sinha et al. 2012; Maffre et al. 2018). The configuration of continental topography may well contribute to the salinity contrast between the Pacific and Atlantic via their effect on atmospheric circulation (Weyl 1968). Modern-day topography may have determined the deep-water formation in the North Atlantic rather than in the North Pacific (Sinha et al. 2012; Maffre et al. 2018). Warren (1983) suggested that the deep-water formation exists in the North Atlantic rather than in the North Pacific because the gyre circulation in the North Atlantic carrying saline water poleward more efficiently than in the North Pacific. Emile-Geay et al. (2003) also showed that key factors affecting the deep-water formation region depend on the modern-day sea-land configuration.

Using coupled climate models, the effect of continental topography on ocean circulation can be studied by removing all mountains or a particular mountain. Through an experiment with all mountains removed, Schmittner et al. (2011) found no NADW formation; and they attributed this to the absence of the RM. In the presence of the RM, the atmospheric moisture transport from the Pacific to the Atlantic is blocked at the midlatitudes, contributing to the high salinity and NADW formation. Through a similar experiment but using a different model, Sinha et al. (2012) found similar AMOC decline and similar mechanism as in Schmittner et al. (2011); that is, in the absence of the RM, more moisture-laden air can travel from the Pacific to the Atlantic, which increases precipitation in the North Atlantic, suppressing the deep-water formation there.

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

Corresponding author: Haijun Yang, yanghj@fudan.edu.cn

DOI: 10.1175/JCLI-D-20-0819.1

© 2021 American Meteorological Society. For information regarding reuse of this content and general copyright information, consult the AMS Copyright Policy (www.ametsoc.org/PUBSReuseLicenses).

Maffre et al. (2018) also performed a similar experiment and concluded that removing entire orography can switch the overturning circulation from the Atlantic to the Pacific. In addition to the freshwater transport mechanism proposed in Schmittner et al. (2011) and Sinha et al. (2012), Maffre et al. (2018) found that the river runoff changes in the tropics also contribute significantly to the North Atlantic freshening. These studies emphasized the role of the RM in regulating the AMOC and PMOC based on flat global continental experiments, instead of using experiments that modify the RM only. Before the individual role of different mountains is isolated, it is hard to identify the role of the RM in the seesaw change of the AMOC and PMOC.

Recent studies have emphasized the role of the Tibetan Plateau (TP) in the AMOC and PMOC. Su et al. (2018) found that removing the TP alone leads to AMOC collapse and PMOC establishment. The initial slowdown of the AMOC is due to intensified westerly wind and increased freshwater flux over the North Atlantic. The further decreased heat flux and increased sea ice area cause the AMOC collapse. The final establishment of the PMOC is due to the increased heat flux over the North Pacific. Yang and Wen (2020) and Wen and Yang (2020) also found similar AMOC and PMOC responses to the TP removal; however, the mechanisms are different. They found that the wind effect increases the AMOC initially and then freshwater effect decreases the AMOC, rather than the heat flux dominating. They emphasized that the atmospheric moisture relocation from the tropical Pacific to the North Atlantic is the key trigger for AMOC weakening, and the positive feedback between the sea ice and AMOC leads to the AMOC collapse.

It is well recognized that removing the continental topography can lead to profound changes in the atmosphere, such as changes in surface wind, moisture transport, and river runoff. Maroon (2016) examined the effects of wind stress and river runoff on ocean circulation through an experiment without the RM. The AMOC is weakened by about 30% in response to the RM removal. In another experiment, the RM topography is unchanged but North American runoff is artificially rerouted to the North Atlantic; similar AMOC weakening appears. They concluded that the impact of the RM on global meridional overturning circulation (MOC) is through its effect on hydrology rather than on surface wind. The authors also examined precipitation asymmetry between the Atlantic and Pacific. The zonal variation of net precipitation [precipitation minus evaporation (PME)] is an important factor affecting the difference of surface salinity between the Atlantic and Pacific. By calculating the atmospheric water budget using atmospheric reanalyze data, Wills and Schneider (2015) and Craig et al. (2020) both found that zonal variation of the basin-integrated PME contribute greatly to the surface salinity contrast between the Atlantic and Pacific, so it can affect the location of the deep-water formation.

In this work, we investigate the RM's role in the formations of the AMOC and PMOC. Through sensitivity experiments with and without the RM, we demonstrate that the RM does not play an important role in either AMOC or PMOC,

distinctly different from the conclusions in many previous studies. In our control run, the AMOC has the strength of about 20 Sv ($1 \text{ Sv} = 10^6 \text{ m}^3 \text{ s}^{-1}$), which is in line with the observed AMOC (Ganachaud and Wunsch 2000; Talley et al. 2003; Longworth et al. 2011; Zhang and Wang 2013). When the RM is removed from a model with realistic topography, the AMOC only declines slightly during the first 100 years and then recovers; the PMOC is hardly changed. When the RM is added to a flat global continent, the AMOC is again barely changed, while the PMOC declines by about 40%. Although perturbing the RM can cause clear changes in North Hemisphere (NH) surface ocean temperature, salinity, and density, the magnitude of these changes is inadequate to cause significant change in the AMOC.

This paper is organized as follows: An introduction to the fully coupled climate model and experiments is given in section 2. The transient and equilibrium responses of the AMOC and PMOC are shown in section 3. The changes of North Atlantic Ocean buoyancy are described in section 4. The atmospheric moisture transport between the Pacific and Atlantic Oceans is analyzed in section 5. Summary and discussion are given in section 6.

2. Model and experiments

In this study, the Community Earth System Model (CESM1.0) is employed. The coupled model version used here consists of an atmosphere model with 26 vertical levels and T31 horizontal resolution ($3.75^\circ \times 3.75^\circ$), a land model that has the same horizontal resolution as the atmosphere model, an ocean model with 60 vertical levels and gx3v7 horizontal resolution, and a sea ice model that has the same horizontal resolution as the ocean model. More detailed information about these model components can be found in Smith and Gregory (2009), Hunke and Lipscomb (2010), Lawrence et al. (2012), Park et al. (2014).

To explore the role of the RM in the formation of the AMOC, two groups of topography experiments are carried out (Fig. 1). The first group includes a 1200-yr flat global topography experiment (Flat) and a 400-yr experiment with only the RM added (OnlyRocky). Specifically, Flat has a flat topography that is 50 m above the sea level globally (Fig. 1a), which starts from the rest and is integrated for 1200 years. OnlyRocky is similar to Flat, except that the realistic topography of the RM has been added (Fig. 1b), which starts from year 801 of Flat and is integrated for 400 years. The second group includes a 2400-yr control run (Real) and a 400-yr experiment with the RM removed (NoRocky). The model geometry, topography, and continents of Real are realistic (Fig. 1c); it starts from the rest with standard configuration and is integrated for 2400 years. NoRocky is similar to Real, except that the topography of the RM is set to 50 m above the sea level (Fig. 1d). It starts from year 2001 of Real and is integrated for 400 years. In all four experiments, the CO_2 concentration is set to the preindustrial value of 285 ppm. Changes in river routing and vegetation type are not considered in these experiments.

The climate changes due to the RM removal are obtained by subtracting the results of Real from those of NoRocky.

FIG. 1. Topography configuration in coupled experiments. (a) Flat, (b) OnlyRocky, (c) Real, and (d) NoRocky. The integration lengths of these experiments are shown at the top. Units: m.

In comparison, the changes due to the presence of the RM are obtained by subtracting the results of Flat from those of OnlyRocky. By comparing these two groups of experiments, we can obtain a quantitative estimate of the RM's role in global climate. The climate changes in NoRocky (with respect to Real) are opposite to those in OnlyRocky (with respect to Flat), though the magnitude is slightly different. In this paper, we focus on the changes due to the presence of the RM. The RM is an ancient mountain range that uplifted about 45 Ma (Ferreira et al. 2018); thus, we can explore the role of the RM in the formation of global thermohaline circulation. All of the above experiments have reached quasi-equilibrium (QE) states in their final 100 years. We focus on the QE changes, which are defined as the climate changes averaged over the last 100 years of each integration, though we also describe the evolution of the MOCs when necessary. In addition, only annual-mean climate changes are presented in this paper.

3. Changes of AMOC and PMOC

The mean climate states in Flat and Real are quite different. Previous studies have shown that the present-day configuration of mountains is an important background to the present-day ocean thermohaline circulation (Schmittner et al. 2011; Sinha et al. 2012; Cessi 2019). The same is true in our experiments. For example, the NADW occurs in Real whereas the NPDW forms in Flat. In the modern climate, deep water in the North Atlantic forms mainly in winter when the heat loss to the

atmosphere destabilizes the water column in the Labrador Sea and Nordic seas, resulting in intense convection that can extend down to 2000 m deep (Marshall and Schott 1999; Weaver et al. 1999). However, in the climate with a flat global continent, some studies showed that the Asian summer monsoon collapses, accompanied by the westward water vapor transport across Africa, which leads to the freshening of the North Atlantic and salinity increase in the North Pacific (Maffre et al. 2018). Our experiments show similar outcomes. In Flat, the NPDW formation occurs while the NADW formation is absent.

Figure 2 shows the temporal evolutions of AMOC and PMOC indexes in the two groups of experiments. The MOC examined in this paper is the sum of Euler mean circulation and those related to the Gent–McWilliams (GM) effect (Gent and McWilliams 1990), submesoscale effect (Fox-Kemper and Ferrari 2008) and dissipation. The AMOC (PMOC) index is defined as the maximum streamfunction in the range of 0–2000 m of 20°–70°N in the Atlantic (Pacific). The AMOC strength in Real is about 18 Sv (dashed black curve; Fig. 2a), in line with the estimate based on observations (Cunningham et al. 2007; Lumpkin and Speer 2007; Kanzow et al. 2010; Talley 2013; McCarthy et al. 2015; Meinen et al. 2018). In NoRocky, the AMOC only declines slightly by about 3 Sv during the first 100 years, and then recovers to the value in Real (dashed blue curve; Fig. 2a). In Flat, the AMOC is completely shut down (solid black curve; Fig. 2a). In OnlyRocky, the AMOC is not established either (solid blue curve; Fig. 2a),

FIG. 2. Temporal evolution of the indices of AMOC and PMOC in Real (dashed black), NoRocky (dashed blue), Flat (solid black), and OnlyRocky (solid blue). Only the last 400 years' evolution of all indices is plotted. (a) The AMOC index and (b) the PMOC index, which are defined as the maximum streamfunction in the range of 0–2000 m of 20°–70°N in the North Atlantic and North Pacific, respectively. Units: Sv.

similar to that in Flat. Figure 2b shows the temporal evolutions of the PMOC index in the two groups of experiments. The PMOC is about 3–4 Sv in Real and is nearly unchanged in NoRocky (dashed curves; Fig. 2b). The PMOC in Flat can be established (solid black curve; Fig. 2b) with its QE value of about 11 Sv. The PMOC in OnlyRocky declines by about 40% with respect to that in Flat, but it does not collapse (solid blue curve; Fig. 2b). The presence of the RM has almost no influence on the formation of AMOC, but has moderate effect on the PMOC.

The effect of the RM on the AMOC and PMOC can be seen more clearly in Fig. 3, which shows the spatial patterns of the AMOC and PMOC averaged in the QE state in Flat and OnlyRocky, and their difference. There is almost no deep overturning circulation in the Atlantic, that is, the AMOC does not exist, in both Flat and OnlyRocky (Figs. 3a,b). Note that the MOC consists of both wind-driven subtropical cell (STC) and thermohaline circulation (Weaver et al. 1993; Toggweiler and Samuels 1995; Nikurashin and Vallis 2012). The wind-driven STC in the Atlantic is clear, which is denoted by the streamfunction contours in the upper 200 m of the tropics (30°S–30°N). In other words, the RM uplift only leads to a very weak intermediate water formation in the upper 400 m of the North Atlantic, which is not strong enough to establish the AMOC (Fig. 3c). In contrast, there is a strong PMOC in Flat, the depth

of the maximum PMOC occurs near 2000 m (Fig. 3d). The PMOC can be weakened significantly by about 40% in response to the RM uplift (Figs. 3e,f). This reduction is mainly attributed to the destruction of the interhemispheric structure of the PMOC (Figs. 3d,e). The Pacific STC is nearly unaffected in the two groups of experiments. Figure 4 is the same as Fig. 3, except for showing the results from Real and NoRocky. In Real, a strong AMOC and a very weak PMOC exist (Figs. 4a,d), which are hardly changed in NoRocky (Figs. 4b,c,e,f).

These experiments demonstrate that in this model changing RM does not affect the AMOC. This is different from most previous findings. Schmittner et al. (2011) concluded that the presence of the RM can block the atmospheric moisture transport from the Pacific to the Atlantic at the midlatitudes, contributing to the high surface salinity and NADW. Maffre et al. (2018) also found an increased freshwater export from the Pacific to the North Atlantic through North America in a flat global topography, which tends to freshen the North Atlantic and weakens the AMOC significantly. They both emphasized the important role of the RM in the AMOC. However, in their model experiments, all global mountains are flattened, instead of flattening the RM only. In our work, only the RM topography is changed, and we find no change in the AMOC. In section 5, we will show that the moisture transport between the Pacific and Atlantic would not cause significant salinity change in the North Atlantic.

4. Changes of ocean buoyancy

The horizontal patterns of sea surface salinity (SSS) change is very similar to that of sea surface density change (SSD) in both OnlyRocky and NoRocky (Fig. 5), suggesting the dominant role of SSS change in SSD change. The patterns of changes of sea surface temperature (SST), SSS, and SSD in the two experiments are generally opposite to each other, suggesting a roughly linear response of the surface ocean to the RM change. Based on the equation of state of the ocean, we further calculate individual contributions of temperature and salinity to density. The salinity-induced density change can be roughly estimated using the salinity in OnlyRocky (NoRocky) and temperature in Flat (Real). Similarly, the temperature-induced density change can be roughly estimated using the temperature in OnlyRocky (NoRocky) and salinity in Flat (Real). Note that the equation of state includes high-order nonlinear terms that may cause some errors, whereas this kind of truncation errors is very small compared to the linear term in the equation.

In both the North Atlantic and North Pacific, the SSS-induced SSD change has roughly the same magnitude and same evolution as the total SSD change (blue and green curves, Fig. 6), suggesting further the dominant role of SSS in changing SSD, and thus the deep-water formation there. Figure 6 shows the temporal evolution of SSD change averaged in the North Atlantic (10°–60°N) and North Pacific (40°–60°N) in OnlyRocky. In the North Atlantic, the SSD decreases during the first 200 years and reaches a QE state in about 400 years (green curve, Fig. 6a), which is not favorable for the AMOC formation. The SSD change consists of SST-induced change

FIG. 3. Patterns of (top) AMOC and (bottom) PMOC in (a),(d) Flat and (b),(e) OnlyRocky and (c),(f) their changes in OnlyRocky with respect to Flat. Units: Sv.

(red curve, Fig. 6a) and SSS-induced change (blue curve, Fig. 6a). It is clear that the SSS change dominates the SSD change in the North Atlantic. In the North Pacific, the SSD decreases rapidly in the first 40 years and then reaches a QE state in about 50 years, which is also not favorable for the PMOC formation.

The North Atlantic upper-ocean changes in OnlyRocky and NoRocky are shown in Fig. 7. In OnlyRocky, the North Atlantic shows cooling and freshening down to 500 m (Figs. 7a,b). The combined effect of temperature and salinity changes leads to a slightly decreased density in the upper 500 m of the North Atlantic (Fig. 7c), with the maximum decrease of

FIG. 4. Patterns of (top) AMOC and (bottom) PMOC in (a),(d) Real and (b),(e) NoRocky and (c),(f) their changes in NoRocky with respect to Real. Units: Sv.

FIG. 5. Changes in (left) SST ($^{\circ}\text{C}$), (center) SSS (psu), and (right) SSD (kg m^{-3}) in (a)–(c) OnlyRocky with respect to Flat and (d)–(f) NoRocky with respect to Real. White contours in (a) and (b) and in (d) and (e) denote the SSD change induced by SST change and SSS change, respectively.

about 0.6 kg m^{-3} . In NoRocky, the combined effect of the North Atlantic upper-ocean warming (Fig. 7d) and saline (Fig. 7e) leads a negligible change in the upper-ocean density (less than 0.3 kg m^{-3}) (Fig. 7f). In both experiments with or without the RM, the temperature and salinity changes tend to compensate each other, leading to density change in the North Atlantic that is too small to cause significant change in the AMOC.

The North Pacific upper-ocean changes are shown in Fig. 8. Similar to those occurred in the North Atlantic, the compensation changes between the upper-ocean temperature and salinity are clear in both experiments. The resultant SSD change in the upper 500 m of the North Pacific is also too small to cause significant change in the PMOC. Note that, consistent with PMOC evolution shown in Fig. 2b, the magnitude of the upper-ocean density decrease in OnlyRocky (Fig. 8c) is much larger than the magnitude of the upper-ocean density increase in NoRocky (Fig. 8f), suggesting a skew response of the North Pacific Ocean to the RM perturbation.

To illustrate the effect of ocean buoyancy change on deep-water formation more clearly, we calculate mixed layer depth (MLD), following the method in Large et al. (1997). The site of the deepest vertical mixing and convection can be found in March MLD (Brady and Otto-Bliesner 2011). Figure 9 shows the March MLD patterns and their changes in OnlyRocky and NoRocky. In Flat, the maximum MLD is located in the North Pacific rather than in the North Atlantic (Fig. 9a), corresponding to a strong PMOC. In OnlyRocky, the MLD becomes shallower in both the North Pacific and North Atlantic (Figs. 9b,c). The former is consistent with the weakened PMOC and the latter suggests a more unfavorable condition

for the AMOC formation. As seen in Fig. 3e, the interhemispheric component of the PMOC collapses in OnlyRocky, which is attributed to the deeper MLD in the Southern Ocean

FIG. 6. Temporal changes in SSD (kg m^{-3} ; green) in OnlyRocky with respect to Flat. SSD changes due to SST (red) and SSS (blue) are also plotted. All variables are averaged in (a) the North Atlantic (10° – 60°N) and (b) the North Pacific (40° – 60°N).

FIG. 7. Depth–latitude section of changes in (left) temperature ($^{\circ}\text{C}$), (center) salinity (psu), and (right) density (kg m^{-3}) in (a)–(c) OnlyRocky with respect to Flat and in (d)–(f) NoRocky with respect to Real. All variables are zonally averaged over the Atlantic.

(Fig. 9c) when the RM is introduced. This suggests a nontrivial role for RM in controlling the Southern Ocean MLD. In Real, the maximum MLD is located in the Greenland–Iceland–Norwegian (GIN) Seas (Fig. 9d), corresponding to a strong AMOC, which is largely consistent with the findings of observations and modeling studies (Yeager and Danabasoglu 2014; Yang et al. 2016; Lozier et al. 2019). In NoRocky, the MLD is hardly changed (Figs. 9e,f), consistent with the unaffected

AMOC and PMOC. Perturbing the height of the RM has no serious consequence to both the AMOC and PMOC.

Since the SSS change controls the SSD change, we only need to focus on the freshwater budget in the Northern Hemisphere. Note that the background SST in the North Atlantic in OnlyRocky is still above the freezing point (-1.8°C), even after the significant cooling between 40° and 60°N (Fig. 5a). The sea ice in the GIN seas expands southward and melts,

FIG. 8. As in Fig. 7, but averaged over the Pacific.

FIG. 9. March MLD (m) in (a) Flat and (b) OnlyRocky and (c) the change in OnlyRocky with respect to Flat. (d)–(f) As in (a)–(c), but in (d) Real and (e) NoRocky and (f) the change in NoRocky with respect to Real.

leading to freshening in the North Atlantic (Fig. 10a). This southward sea ice expansion is mainly along the eastern coast of Greenland and driven by the anomalous cyclonic circulation over the subpolar Atlantic (Fig. 11a). The solid black and red curves in Fig. 10a represent the sea ice margin in Flat and OnlyRocky, respectively. The sea ice coverage is defined as

the area of the ocean with ice concentration above 15%. The sea ice coverage in the North Atlantic expands southward in OnlyRocky, providing freshwater to the North Atlantic and contributing about -0.3 psu yr^{-1} to the SSS tendency of 40° – 60°N or about $-0.08 \text{ psu yr}^{-1}$ of 10° – 60°N (the latter is the same as the region used for calculating the moisture transport

FIG. 10. Changes in (a),(d) virtual salt flux (VSF) due to sea ice (psu yr^{-1}). Positive (negative) value represents sea ice formation (melting). (b),(e) VSF due to EMP (psu yr^{-1}). Positive (negative) value represents evaporation larger (smaller) than precipitation. The sum of (c) sea ice formation and (f) EMP. (top) Changes in OnlyRocky with respect to Flat; (bottom) changes in NoRocky with respect to Real. Solid black and red curves show the sea ice margin in Flat (Real) and OnlyRocky (NoRocky), respectively. The sea ice margin is defined by the 15% sea ice fraction, which is used throughout the paper.

FIG. 11. Changes in (a) geopotential height (shading; m) and wind (vectors; m s^{-1}) at 850 hPa and (b) vertically integrated moisture transport ($\langle \rho \mathbf{v} q \rangle$; vectors; $\text{kg m}^{-1} \text{s}^{-1}$) and its convergence ($-\nabla \cdot \langle \rho \mathbf{v} q \rangle$; shading; $10^{-5} \text{ kg m}^{-2} \text{s}^{-1}$) in OnlyRocky with respect to Flat. (c),(d) As in (a) and (b), but for changes in NoRocky with respect to Real. The angle brackets denote the density-weighted vertical integral, and ρ is air density. In (b) and (d), positive value denotes convergence, and suggest that ocean gains freshwater from the atmosphere. The yellow dashed box denotes the region for freshwater budget calculation.

in section 5). The evaporation minus precipitation (EMP) flux over the North Atlantic freshens the upper ocean at a stable rate of about $-0.12 \text{ psu yr}^{-1}$ (Fig. 10b).

In contrast, the salinity budget in the North Pacific is different from that of the North Atlantic. The North Pacific sea ice in OnlyRocky generally retreats northward (the red and black curve in Fig. 10a) contributes about 0.04 psu yr^{-1} to the SSS tendency in $40^{\circ}\text{--}60^{\circ}\text{N}$. The EMP flux contributes about 0.07 psu yr^{-1} to the SSS tendency in $10^{\circ}\text{--}60^{\circ}\text{N}$ (Fig. 10b). The effect of the moisture flux to salinity in the North Pacific is much weaker than that in the North Atlantic. The main factors to freshen the North Pacific upper ocean come from the Ekman pumping term ($-0.08 \text{ psu yr}^{-1}$) and horizontal diffusion term ($-0.03 \text{ psu yr}^{-1}$) (figure not shown), which lead to the weakened PMOC in OnlyRocky. In NoRocky, the changes in sea ice coverage and EMP with respect to Real (Figs. 10d–f) are roughly opposite to those in OnlyRocky, but with a much smaller magnitude. This is consistent with negligible changes in the ocean thermohaline circulation in NoRocky.

5. Atmospheric moisture transport

It is well recognized that the presence of the RM weakens the midlatitude westerlies and eastward atmospheric moisture transport from the North Pacific to the North Atlantic

(e.g., Schmittner et al. 2011). However, the presence of the RM also enhances the northeastward atmospheric mass and moisture transports from the eastern tropical Pacific to the North Atlantic. Figure 11a shows that in OnlyRocky there are an anomalous anticyclonic and a cyclonic circulations to the north and south of the RM, respectively, which lead to an anomalous westward moisture transport from the North Atlantic to the North Pacific around 40°N , and an anomalous northeastward moisture transport from the eastern tropical Pacific to the North Atlantic (Fig. 11b). Note that there is an anomalous cyclonic circulation in the subpolar Atlantic (Fig. 11a), which helps to enhance the southward sea ice expansion along eastern Greenland as seen in Fig. 10a. In NoRocky (Figs. 11c,d), the changes in atmospheric circulation and moisture transport are roughly opposite to those in OnlyRocky; that is, there are an anomalous eastward moisture transport from the North Pacific to the North Atlantic around 40°N and southwestward moisture transport from the North Atlantic to the eastern tropical Pacific. The two moisture transport pathways between the Pacific and Atlantic have the opposite effects on the SSS in the North Atlantic. Their net effect can be quantified, and will be discussed later (Fig. 12). Here, the atmospheric moisture transport ($\langle \mathbf{v} q \rangle$) is calculated by vertically integrating product of atmospheric circulation and specific humidity. The moisture fluxes across four boundaries of the dashed yellow box outlined in Figs. 11b and 11d will be quantified in Fig. 12 as well.

In fact, the atmospheric divergence pattern shows that the North Atlantic surface ocean gains (loses) freshwater from (to) the atmosphere in OnlyRocky (NoRocky) (shading; Figs. 11b,d). This is consistent with the virtual salt flux change due to EMP shown in Figs. 10b and 10e. This is opposite to the studies that suggested a more saline North Atlantic upper ocean in the presence of the RM, or a fresher upper ocean in the absence of the RM (Schmittner et al. 2011; Maffre et al. 2018). The atmospheric moisture convergence ($-\nabla \cdot \mathbf{v}q$) can be used to assess the net freshwater gain or loss over certain region. In Figs. 11b and 11d, the positive (negative) value of the shading represents atmospheric moisture convergence (divergence), that is, a loss (gain) of atmosphere freshwater to (from) the ocean and a freshening (saline) of the upper ocean. Note that the net freshwater budget over the ocean due to atmospheric moisture convergence can also be easily estimated by EMP, since for a steady state, the vertically integrated moisture transport divergence for an entire atmosphere column is equivalent to the net freshwater flux across the ocean surface, when neglecting the freshwater flux of the continental groundwater and river runoff. Table 1 lists the net freshwater flux and its change over the North Atlantic and North Pacific in different experiments. In both Flat and Real, the EMP over both the North Atlantic and North Pacific is about -0.4 Sv, acting to increase SSS. In OnlyRocky, the North Atlantic upper ocean gains about 0.11 Sv freshwater, tending to reduce SSS; the North Pacific upper ocean loses about

0.13 Sv freshwater, tending to increase SSS further. In NoRocky, the freshwater flux changes are opposite to those in OnlyRocky, with a smaller magnitude. In all the experiments, the river runoff and its change are much smaller than EMP, and play a negligible role in SSS change.

To further identify how the atmospheric moisture transport affects the net freshwater budget in the North Atlantic, we calculate the moisture flux across the rectangle boundaries outlined in Fig. 11. First of all, the net moisture flux changes across the four boundaries of the rectangle are about $+0.116$ Sv in OnlyRocky and -0.069 Sv in NoRocky (Fig. 12), roughly equal to the net EMP changes listed in Table 1, which are about $+0.110$ and -0.055 Sv, respectively. Second, the moisture flux across the western boundary of the rectangle box, which reflects the direct effect of the RM on the atmospheric moisture transport, does indicate a slightly saline effect on the North Atlantic SSS. Consistent with Schmittner et al. (2011), the presence (absence) of the RM does block (enhance) the eastward freshwater transport through the northern pathway, that is, from the North Pacific to the North Atlantic, which is estimated to be about 0.175 Sv in OnlyRocky (Fig. 12a), and about 0.065 Sv in NoRocky (Fig. 12b). However, the southern pathway, connecting the freshwater transport between the eastern tropical Pacific and North Atlantic, plays an opposite role to the northern pathway. It leads to an enhanced (a weakened) eastward transport of about 0.122 Sv (0.117 Sv) in OnlyRocky (NoRocky) (Figs. 12a,b). The net freshwater

FIG. 12. Changes in vertically integrated (a),(b) zonal moisture transport ($\langle uq \rangle$; $\text{kg m}^{-1} \text{s}^{-1}$) and (c),(d) meridional moisture transport ($\langle vq \rangle$; $\text{kg m}^{-1} \text{s}^{-1}$). (left) Changes in OnlyRocky with respect to Flat; (right) changes in NoRocky with respect to Real. The angle brackets denote the density-weighted vertical integral. Positive value denotes eastward or northward transport. Numbers at boundaries of the yellow rectangle denote the transports across the boundaries. Units: Sv.

TABLE 1. EMP and river runoff calculated in the North Atlantic (10°–60°N, 80°W–0°) and North Pacific (10°–60°N, 120°E–100°W) in all simulations. Positive value indicates freshwater flux into the ocean. Units: Sv.

Expt	North Atlantic		North Pacific	
	EMP	River runoff	EMP	River runoff
Flat	−0.381	0.085	−0.380	0.076
OnlyRocky-Flat	0.110	−0.012	−0.130	+0.011
Real	−0.409	0.089	−0.375	0.190
NoRocky-Real	−0.055	0.001	0.121	−0.012

transport of these two pathways is about -0.05 Sv in both OnlyRocky and NoRocky, that is, having a slightly saline effect on the North Atlantic SSS regardless of the presence of the RM. As far as the zonal moisture transport ($\langle uq \rangle$) over the North American continent is concerned, our experiments show a saline effect of the RM on the North Atlantic SSS (Figs. 12a,b).

However, the RM affects not only the atmospheric circulation over the North American continent but also the remote region significantly. The remote effect of the RM on the North Atlantic freshwater budget can be roughly estimated by looking at the meridional moisture transport ($\langle vq \rangle$) as well as the zonal moisture transport in the eastern Atlantic (Figs. 12c,d). In OnlyRocky, the meridional moisture convergence contributes a total of 0.058 Sv freshwater to the North Atlantic (Fig. 12c), of which about 0.039 Sv comes from the equatorial Atlantic and 0.019 Sv from the polar region. In addition, there is about 0.111 Sv freshwater coming from the European continent (Fig. 12a). These remote moisture transports finally lead to a freshwater surplus of about 0.116 Sv in the North Atlantic. In NoRocky (Fig. 12d), the meridional moisture divergence leads to about 0.011 Sv freshwater out of the North Atlantic, mainly exporting to the polar ocean. Also there is about 0.006 Sv freshwater exporting to the European continent. Together, they contribute to the total 0.069 Sv freshwater deficit in the North Atlantic.

In general, our experiments show that the atmospheric moisture transport change in response to the RM uplift (removal) leads to a fresher (more saline) North Atlantic. The amount of moisture convergence (divergence) over the North Atlantic (10°–60°N) decreases (increases) SSS at a stable rate of about -0.12 psu yr^{-1} ($+0.07$ psu yr^{-1}) in OnlyRocky (NoRocky). Similar SSS tendency due to the atmospheric moisture convergence occurs in the North Pacific (10°–60°N). This amount of freshwater forcing cannot cause significant change in the deep-water formation in both the North Atlantic and North Pacific. We have also carefully calculated the effect of the changes in river runoff, horizontal and vertical salinity advection, and dissipation on the SSS fields of the North Atlantic and North Pacific (figure not shown). The combined effect of all these processes results in an SSD change of less than ± 1.5 kg m^{-3} (Figs. 5c,f), which is inadequate to cause significant change in the deep-water formation in the Northern Hemisphere.

6. Summary and discussion

In this study, we use sensitivity experiments with and without the RM to demonstrate a trivial role of the RM in the formation of both AMOC and PMOC. In the two groups of experiments, the AMOC is barely changed, and the PMOC is unchanged in NoRocky with respect to Real and but is weakened by about 40% in OnlyRocky with respect to Flat. The latter is attributed to the destruction of the interhemispheric structure of the PMOC caused by the enhanced Southern Ocean sinking in OnlyRocky (Figs. 3e, 9b). We conclude that the RM does not play a significant role in the Northern Hemisphere sinking.

The climate changes in NoRocky are the opposite of those in OnlyRocky, though the magnitude is marginally different. In the presence of the RM, less atmospheric moisture is transported from the North Pacific to the North Atlantic, but at the same time less moisture is transported from the tropical Atlantic to the eastern tropical Pacific. Surely the RM can help to relocate the moisture transport between the Pacific and Atlantic. However, as far as the net impact of atmospheric moisture transport on the North Atlantic salinity is concerned, the role of the RM appears to be negligible. In the presence of the RM, we see no sustained NADW formation.

The findings in this work are different from those reported in Schmittner et al. (2011) and Maffre et al. (2018), in which the authors concluded an important contribution of the RM to the higher SSS in the North Atlantic. We found that the net effect of the moisture transport tends to freshen the North Atlantic in the presence of the RM. We think the different conclusions are mainly due to that fact that they modified all continental topography in their model experiments, but attributed the North Atlantic SSS change to the RM. In our experiments, only the RM is perturbed, so that the individual contribution of the RM to the global climate can be quantified cleanly, by analyzing anomalies of OnlyRocky-Flat (or NoRocky-Real) rather than contrasting Flat and Real. Maffre et al. (2018) also found that the river runoff changes in the tropics contribute to the freshening of the North Atlantic. In our work, the changes in river runoff are very small in the North Atlantic and North Pacific (Table 1).

As far as the atmospheric moisture exchange between the Pacific and Atlantic is concerned, both Schmittner et al. (2011) and our work identify two moisture transport pathways between the two oceans. The zonal moisture transport pathway around 40°N are the same in both studies; that is, there would be weakened (enhanced) eastward moisture transport from the North Pacific to the North Atlantic in the presence (absence) of the RM. The southern moisture transport pathway in our study acts to bring more moisture from the eastern tropical Pacific to the North Atlantic in the presence of the RM, while Schmittner et al. (2011) concluded that this pathway acts to increase moisture export from the Atlantic to the tropical Pacific by the trade winds in the presence of the RM. Again, this different occurs because Schmittner et al. (2011) perturbed all continental topography. In fact, our previous study did suggest an enhanced southwestward moisture

transport through the southern pathway as in Schmittner et al. (2011), when perturbing the topography of the TP (Yang and Wen 2020). The RM alone plays an opposite role in moisture transport direction between the Atlantic and tropical Pacific to that the TP does. And the presence of the RM does not help the NADW.

Paleoclimatic studies have dated the AMOC establishment back to about 12 Ma, which was close to the time of the TP uplift (10–8 Ma). The TP uplift may have affected the formation of the AMOC under modern-day geometry. Our previous study revealed that removing the TP can lead to the AMOC shutdown and PMOC establishment (Yang and Wen 2020). The atmospheric moisture relocation between the Pacific and North Atlantic is a critical factor to trigger the initial weakening of the AMOC. The final AMOC collapse is attributed to the positive feedback between sea ice melting and AMOC weakening. This work further shows that although perturbing the RM can cause clear changes in SST, SSS and SSD in the Northern Hemisphere oceans, the magnitude of these changes, particularly the magnitude of SSS change, is not strong enough to cause clear change in the AMOC and PMOC. Although the distance between the RM and the North Atlantic is much shorter than that between the TP and the North Atlantic, the TP plays a more important role in regulating the global hydrological cycle (Yang and Wen 2020). Thus, we conclude that it is the TP, rather than the RM, that plays more important role in the formation of the modern global thermohaline circulation.

The conclusions drawn in this paper may be model dependent and subject to model limitations. The location of the deep-water formation simulated in the model may be related to model's resolution and different from observation. Background CO₂ concentration in our experiments is set at the preindustrial level, which is different from that when the RM uplifted at 45 Ma. Topography change in our experiments is instantaneous and different from the actual evolution process. River routing scheme and runoff can be very different in different background continental topography. The results of using different models or different background climate parameters could differ. Nevertheless, this highly idealized modeling study provides a quantitative way to understand the role of the giant continental topography in the real world. Our modeling has been used to explore the formation of global thermohaline circulation with a gradual evolution of the continental terrain. We are conducting more sensitivity experiments to investigate ocean circulation evolution in response to transient changes in continental orography, geometry, and CO₂ at geological time scale. More interesting results are expected in the near future.

Acknowledgments. This work is supported by the National Natural Science Foundation of China (91737204, 41725021, and 91937302). We greatly appreciate discussion with Prof. Z. Liu at The Ohio State University, and invaluable suggestions from three reviewers. The experiments were performed at the National Supercomputer Centre in Tianjin (Tian-He No.1).

REFERENCES

- Brady, E. C., and B. L. Otto-Bliesner, 2011: The role of meltwater-induced subsurface ocean warming in regulating the Atlantic meridional overturning in glacial climate simulations. *Climate Dyn.*, **37**, 1517–1532, <https://doi.org/10.1007/s00382-010-0925-9>.
- Cessi, P., 2019: The global overturning circulation. *Annu. Rev. Mar. Sci.*, **11**, 249–270, <https://doi.org/10.1146/annurev-marine-010318-095241>.
- Craig, P. M., D. Ferreira, and J. Methven, 2020: Monsoon-induced zonal asymmetries in moisture transport cause anomalous pacific precipitation minus evaporation. *Geophys. Res. Lett.*, **47**, <https://doi.org/10.1029/2020GL088659>.
- Cunningham, S. A., and Coauthors, 2007: Temporal variability of the Atlantic meridional overturning circulation at 26°N. *Science*, **317**, 935–938, <https://doi.org/10.1126/science.1141304>.
- Emile-Geay, J., M. A. Cane, N. Naik, R. Seager, A. C. Clement, and A. van Geen, 2003: Warren revisited: Atmospheric freshwater fluxes and “Why is no deep water formed in the North Pacific.” *J. Geophys. Res.*, **108**, 3178, <https://doi.org/10.1029/2001JC001058>.
- Ferreira, D., and Coauthors, 2018: Atlantic-Pacific asymmetry in deep-water formation. *Annu. Rev. Earth Planet. Sci.*, **46**, 327–352, <https://doi.org/10.1146/annurev-earth-082517-010045>.
- Fox-Kemper, B., and R. Ferrari, 2008: Parameterization of mixed layer eddies. Part II: Prognosis and impact. *J. Phys. Oceanogr.*, **38**, 1166–1179, <https://doi.org/10.1175/2007JPO3788.1>.
- Ganachaud, A., and C. Wunsch, 2000: Improved estimates of global ocean circulation, heat transport and mixing from hydrographic data. *Nature*, **408**, 453–457, <https://doi.org/10.1038/35044048>.
- Gent, P., and J. C. McWilliams, 1990: Isopycnal mixing in ocean circulation models. *J. Phys. Oceanogr.*, **20**, 150–155, [https://doi.org/10.1175/1520-0485\(1990\)020<0150:IMOCM>2.0.CO;2](https://doi.org/10.1175/1520-0485(1990)020<0150:IMOCM>2.0.CO;2).
- Hunke, E. C., and W. H. Lipscomb, 2010: CICE: The Los Alamos Sea Ice Model documentation and software user's manual, version 4.1. Doc. LACC-06-012, 76 pp., https://cdms.colorado.edu/w/images/CICE_documentation_and_software_user%27s_manual.pdf.
- Kanzow, T., and Coauthors, 2010: Seasonal variability of the Atlantic meridional overturning circulation at 26.5°N. *J. Climate*, **23**, 5678–5698, <https://doi.org/10.1175/2010JCLI3389.1>.
- Kittel, T. G. F., P. E. Thornton, J. A. Royle, and T. N. Chase, 2002: Climates of the rocky mountains: Historical and future patterns. *Rocky Mountain Futures: An Ecological Perspective*, J. S. Baron, Ed., Island Press, 59–92.
- Large, W. G., G. Danabasoglu, S. C. Doney, and J. C. McWilliams, 1997: Sensitivity to surface forcing and boundary layer mixing in a global ocean model: Annual-mean climatology. *J. Phys. Oceanogr.*, **27**, 2418–2447, [https://doi.org/10.1175/1520-0485\(1997\)027<2418:STSFA>2.0.CO;2](https://doi.org/10.1175/1520-0485(1997)027<2418:STSFA>2.0.CO;2).
- Latif, M., and Coauthors, 2004: Reconstructing, monitoring, and predicting multidecadal-scale changes in the North Atlantic thermohaline circulation with sea surface temperature. *J. Climate*, **17**, 1605–1614, [https://doi.org/10.1175/1520-0442\(2004\)017<1605:RMAPMC>2.0.CO;2](https://doi.org/10.1175/1520-0442(2004)017<1605:RMAPMC>2.0.CO;2).
- Lawrence, D. M., and Coauthors, 2012: The CCSM4 land simulation, 1850–2005: Assessment of surface climate and new capabilities. *J. Climate*, **25**, 2240–2260, <https://doi.org/10.1175/JCLI-D-11-00103.1>.

- Longworth, H. R., H. L. Bryden, and M. O. Baringer, 2011: Historical variability in Atlantic meridional baroclinic transport at 26.5°N from boundary dynamic height observations. *Deep-Sea Res. II*, **58**, 1754–1767, <https://doi.org/10.1016/j.dsr2.2010.10.057>.
- Lozier, M. S., and Coauthors, 2019: A sea change in our view of overturning in the subpolar North Atlantic. *Science*, **363**, 516–521, <https://doi.org/10.1126/science.aau6592>.
- Lumpkin, R., and K. Speer, 2007: Global ocean meridional overturning. *J. Phys. Oceanogr.*, **37**, 2550–2562, <https://doi.org/10.1175/JPO3130.1>.
- Maffre, P., J.-B. Ladant, Y. Donnadieu, P. Sepulchre, and Y. Goddris, 2018: The influence of orography on modern ocean circulation. *Climate Dyn.*, **50**, 1277–1289, <https://doi.org/10.1007/s00382-017-3683-0>.
- Maroon, E. A., 2016: The roles of land and orography on precipitation and ocean circulation in global climate models. Ph.D. thesis, University of Washington, 155 pp.
- Marshall, J., and F. Schott, 1999: Open-ocean convection: Observations, theory, and models. *Rev. Geophys.*, **37**, 1–64, <https://doi.org/10.1029/98RG02739>.
- McCarthy, G. D., and Coauthors, 2015: Measuring the Atlantic meridional overturning circulation at 26°N. *Prog. Oceanogr.*, **130**, 91–111, <https://doi.org/10.1016/j.pocean.2014.10.006>.
- Meinen, C. S., and Coauthors, 2018: Meridional overturning circulation transport variability at 34.5°S during 2009–2017: Baroclinic and barotropic flows and the dueling influence of the boundaries. *Geophys. Res. Lett.*, **45**, 4180–4188, <https://doi.org/10.1029/2018GL077408>.
- Nikurashin, M., and G. Vallis, 2012: A theory of the interhemispheric meridional overturning circulation and associated stratification. *J. Phys. Oceanogr.*, **42**, 1652–1667, <https://doi.org/10.1175/JPO-D-11-0189.1>.
- Park, S., C. S. Bretherton, and P. J. Rasch, 2014: Integrating cloud processes in the Community Atmosphere Model, version 5. *J. Climate*, **27**, 6821–6856, <https://doi.org/10.1175/JCLI-D-14-00087.1>.
- Schmittner, A., T. A. M. Silva, K. Fraedrich, E. Kirk, and F. Lunkeit, 2011: Effects of mountains and ice sheets on global ocean circulation. *J. Climate*, **24**, 2814–2829, <https://doi.org/10.1175/2010JCLI3982.1>.
- Sinha, B., A. T. Blaker, J. J.-M. Hirschi, S. Bonham, M. Brand, S. Josey, R. S. Smith, and J. Marotzke, 2012: Mountain ranges favour vigorous Atlantic meridional overturning. *Geophys. Res. Lett.*, **39**, L02705, <https://doi.org/10.1029/2011GL050485>.
- Smith, R. S., and J. M. Gregory, 2009: A study of the sensitivity of ocean overturning circulation and climate to freshwater input in different regions of the North Atlantic. *Geophys. Res. Lett.*, **36**, L15701, <https://doi.org/10.1029/2009GL038607>.
- Su, B., D. Jiang, R. Zhang, P. Sepulchre, and G. Ramstein, 2018: Difference between the North Atlantic and Pacific meridional overturning circulation in response to the uplift of the Tibetan Plateau. *Climate Past*, **14**, 751–762, <https://doi.org/10.5194/cp-14-751-2018>.
- Swingedouw, D., T. Fichefet, H. Goosse, and M. F. Loutre, 2009: Impact of transient freshwater releases in the Southern Ocean on the AMOC and climate. *Climate Dyn.*, **33**, 365–381, <https://doi.org/10.1007/s00382-008-0496-1>.
- Talley, L. D., 2013: Closure of the global overturning circulation through the Indian, Pacific, and Southern Oceans: Schematics and transports. *Oceanography*, **26**, 80–97, <https://doi.org/10.5670/oceanog.2013.07>.
- , J. L. Reid, and P. E. Robbins, 2003: Data-based meridional overturning streamfunctions for the global ocean. *J. Climate*, **16**, 3213–3226, [https://doi.org/10.1175/1520-0442\(2003\)016<3213:DMOSFT>2.0.CO;2](https://doi.org/10.1175/1520-0442(2003)016<3213:DMOSFT>2.0.CO;2).
- Toggweiler, J. R., and B. Samuels, 1995: Effect of Drake Passage on the global thermohaline circulation. *Deep-Sea Res. I*, **42**, 477–500, [https://doi.org/10.1016/0967-0637\(95\)00012-U](https://doi.org/10.1016/0967-0637(95)00012-U).
- Warren, B. A., 1983: Why is no deep water formed in the North Pacific? *J. Mar. Res.*, **41**, 327–347, <https://doi.org/10.1357/00224083788520207>.
- Weaver, A. J., J. Marotzke, P. F. Cummins, and E. S. Sarachik, 1993: Stability and variability of the thermohaline circulation. *J. Phys. Oceanogr.*, **23**, 39–60, [https://doi.org/10.1175/1520-0485\(1993\)023<0039:SAVOTT>2.0.CO;2](https://doi.org/10.1175/1520-0485(1993)023<0039:SAVOTT>2.0.CO;2).
- , C. M. Bitz, A. F. Fanning, and M. M. Holland, 1999: Thermohaline circulation: High-latitude phenomena and the difference between the Pacific and the Atlantic. *Annu. Rev. Earth Planet. Sci.*, **27**, 231–285, <https://doi.org/10.1146/annurev.earth.27.1.231>.
- Wen, Q., and H. Yang, 2020: Investigating the role of the Tibetan Plateau in the formation of Pacific meridional overturning circulation. *J. Climate*, **33**, 3603–3617, <https://doi.org/10.1175/JCLI-D-19-0206.1>.
- Weyl, P. K., 1968: The role of the oceans in climate change: A theory of the ice ages. *Causes of Climatic Change, Meteor. Monogr.*, No. 30, Amer. Meteor. Soc., 37–62.
- Wills, R. C., and T. Schneider, 2015: Stationary eddies and the zonal asymmetry of net precipitation and ocean freshwater forcing. *J. Climate*, **28**, 5115–5133, <https://doi.org/10.1175/JCLI-D-14-00573.1>.
- Wunsch, C., and R. Ferrari, 2004: Vertical mixing, energy, and the general circulation of the oceans. *Annu. Rev. Fluid Mech.*, **36**, 281–314, <https://doi.org/10.1146/annurev.fluid.36.050802.122121>.
- Yang, H., and Q. Wen, 2020: Investigating the role of the Tibetan Plateau in the formation of Atlantic meridional overturning circulation. *J. Climate*, **33**, 3585–3601, <https://doi.org/10.1175/JCLI-D-19-0205.1>.
- , K. Wang, H. Dai, Y. Wang, and Q. Li, 2016: Wind effect on the Atlantic meridional overturning circulation via sea ice and vertical diffusion. *Climate Dyn.*, **46**, 3387–3403, <https://doi.org/10.1007/s00382-015-2774-z>.
- Yeager, S., and G. Danabasoglu, 2014: The origins of late-twentieth-century variations in the large-scale North Atlantic circulation. *J. Climate*, **27**, 3222–3247, <https://doi.org/10.1175/JCLI-D-13-00125.1>.
- Zhang, L., and C. Wang, 2013: Multidecadal North Atlantic sea surface temperature and Atlantic meridional overturning circulation variability in CMIP5 historical simulations. *J. Geophys. Res. Oceans*, **118**, 5772–5791, <https://doi.org/10.1002/jgrc.20390>.