Fast and Slow Responses of Atlantic Meridional Overturning Circulation to Antarctic Meltwater Forcing

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March 06, 2024

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Key Points:

- The response of the AMOC to perennial meltwater-induced cooling is investigated by GFDL CM2.1 experiments
- Before the cooling spreads out the Atlantic, tropical atmospheric teleconnection could weaken the Greenland ocean convection
- The fast and slow responses imply the importance of the atmospheric teleconnection to regulate polar climate

Abstract

Antarctic meltwater discharge has been largely emphasized for its potential role in climate change mitigation, not only by reducing global warming, but also by stabilizing the Atlantic Meridional Overturning Circulation (AMOC). Despite the tremendous impact of the AMOC on the climate system, its temporal evolution in response to the meltwater remains poorly understood. Here, we investigate the meltwater impacts on the AMOC based on the GFDL CM2.1 experiments and discover its fast weakening and slow strengthening to the Antarctic meltwater discharge. Cold ocean surface caused by meltwater spread throughout the globe and eventually strengthened the AMOC. However, in the early stages, the tropical temperature response could stimulate the Rossby wave teleconnection, modulating atmospheric circulation in the North Atlantic, and weakening convection and even the AMOC. This counterintuitive evolution implies a potential destabilizing effect of Antarctic meltwater, underscoring the importance of the atmospheric dynamics in the interaction between the two poles.

Plain Language Summary

The climate system is made up of interactions between different subsystems, so that regional climate changes can have global effects. The freshwater discharge from Antarctica would increase in the future and result in regional cooling. Atmospheric and oceanic dynamics extend this local effect to the globe, reducing global surface temperature and strengthening the large-scale ocean circulation in the Atlantic. These mitigating effects naturally put the spotlight on Antarctic meltwater. Our study however suggests that the mitigation effect depends on the time scale. Although the global mean temperature is always reduced, the ocean circulation in the Atlantic surprisingly slows down as an early response to Antarctic meltwater; fast atmospheric teleconnection enables it. This non-monotonic evolution emphasizes the importance of the atmospheric teleconnection between the two poles, which should be carefully considered to understand the polar climate.

1. Introduction

The growing discharge of freshwater into the Southern Ocean (SO) from the Antarctic ice melt is one of the undeniable observations and the anticipated consequences of global warming. Recent satellite data indicate that Antarctic mass loss has increased sharply over the past 40 years (e.g., Rignot et al., 2019; Shepherd et al., 2018), and this trend is projected to continue in the next century (DeConto & Pollard, 2016; Hansen et al., 2016). The mass redistribution has been largely highlighted because it will contribute not only to the global sea level rise (Hanna et al., 2020) but also to a substantial response of the regional climate system associated with distinct structures of the SO (e.g., Bintanja et al., 2013). The ocean circulation surrounding Antarctica is characterized by a cold surface layer and a warm circumpolar deep water (CDW). As meltwater flows into the salty ocean, the low density further reduces ocean mixing, thereby inhibiting CDW upwelling (Fogwill et al., 2015), even though the horizontal gradient of ocean temperature and salinity under ice shelf could induce small-scale mixing and horizontal intrusion (Na et al., 2023). The CDW isolation modulates the biogeochemical properties of the SO (Bronselaer et al., 2020; Oh et al., 2022) and leads to increased sea-ice extent and decreased SO surface temperature (Park & Latif, 2019; Pauling et al., 2016).

Even local changes in extratropical regions can have noticeable effects in other regions: the tropics (e.g., Kang et al., 2020; Shin et al., 2021) and even the opposite extratropics (e.g., Cabré et al., 2017; England et al., 2020a; Shin & Kang, 2021). The meltwater-induced cooling is a representative example of such a global teleconnection, causing a northward shift of the Intertropical Convergence Zone (ITCZ)—a narrow band of rainfall near the equator—and global-wide cooling (Bakker & Prange, 2018; Bronselaer et al., 2018). This cooling pattern is accompanied by local cooling minima in East Asia and the Subpolar Northern Atlantic (SPNA). The former has been proposed to be explained by the atmospheric Rossby wave teleconnection mechanism (Oh et al., 2020). The latter, the so-called cooling hole, is generally associated with the strengthening of the Atlantic Meridional Overturning Circulation (AMOC) (e.g., Buckley & Marshall, 2016; Keil et al., 2020), which suggests that the freshwater injection into the Southern Ocean eventually leads to a positive AMOC response (Li et al., 2023; Weaver et al., 2003). Particularly, this response may help to delay the AMOC collapse and its associated impacts (Sinet et al., 2023; Wunderling et al., 2021). Therefore, the Antarctic meltwater has been widely emphasized for its potential role in mitigating both gradual and abrupt climate change.

While much attention has been paid to the impacts of Antarctic meltwater, less has been paid to its temporal evolution. It may be acceptable to pay less investigation given the slow time scale of ocean adjustment. However, the interplay between the atmospheric and the oceanic pathways could be of great importance in formulating climate responses on different time scales. For example, the Arctic sea-ice loss could lead to dramatically different impacts between longer multidecadal and shorter decadal time scales (Liu & Fedorov, 2019). They showed that considerable time is required for the slow AMOC response to overcome the global atmospheric teleconnection induced by sea-ice loss, ultimately resulting in split climate patterns with respect to the time scale. Given that Antarctic meltwater input could lead to AMOC responses and also to atmospheric Rossby wave teleconnection, it is natural and reasonable to investigate the potential role of the Antarctic meltwater forcing in the AMOC and regional climate variations across various time scales.

The primary purpose of this study is accordingly to investigate the impact of Antarctic meltwater input on the AMOC and its temporal evolution. We conduct a series of experiments to identify the impact of meltwater input. Our findings reveal that an interplay between the atmospheric and oceanic pathways can result in a non-monotonic response of the AMOC to the Antarctic meltwater input: early weakening and late strengthening. This non-monotonic response emphasizes the connection between the two polar regions in regulating climate response to external forcing.

2. Model and experiment

In this study, we employ a fully coupled model, CM2.1, developed at the Geophysical Fluid Dynamics Laboratory (Delworth et al., 2006). The atmosphere and land models have a nominal 2° horizontal resolution, and the ocean and ice models have a nominal 1° horizontal resolution. We start with a 2000-year control simulation (CTL) at an atmospheric carbon dioxide concentration of 353 ppm, representing the 1990 level. The first 1,000 years are discarded to avoid long-term drift and the last 1,000 years are used as initial conditions for forced ensemble experiments. We conducted two forced experiments for 12 ensemble members, each integrated from every 50 years of the last 1000 years of the control experiment. One is the MW_off experiment, which is solely forced by atmospheric carbon dioxide (CO_2) concentration that increases at a rate of 1% yr^{-1} for 70 years and then remains at doubled CO_2 concentration (706 ppm). The other is the MW_on experiment, which is not only forced by the same radiative forcing, but also by an idealized Antarctic meltwater input. The meltwater forcing is introduced at the surface, assuming that it results from ice-sheet/shelf melting. The meltwater forcing is a time-invariant 0.2 Sv freshwater discharge, equivalent to a sea level rise of 1.6 cm per year. This aligns with the expected amount around 2050 under the Representative Concentration Pathway 8.5 scenarios (DeConto & Pollard, 2016). Considering contributions from large icebergs, the anticipated timeframe could be earlier than 2050 (e.g., England et al., 2020b). The meltwater distribution is injected proportional to the climatological runoff of Antarctica into the Southern Ocean. Thus, the majority of the meltwater is concentrated around West Antarctica, as indicated by recent observations (Rignot et al., 2019; Shepherd et al., 2018). The difference between the two experiments indicates the impact of Antarctic meltwater: $\delta = \mathrm{MW\_on} - \mathrm{MW\_off}$.

The statistical significance of meltwater impacts, δ, is measured by a bootstrap analysis that calculates the 95% confidence level between the 25th and 975th values among randomly generated 1,000 bootstrap samples. Note that this set of experiments shares a general design with previous studies investigating Antarctic meltwater impacts (Oh et al., 2022; Park & Latif, 2019).

3. Result

The impact of Antarctic meltwater on global climate is shown in Figure 1. Doubling CO_2 warms the global mean surface temperature by 1.6 K without meltwater input (red in Figure 1a). As previous studies suggested, the Antarctic meltwater reduces surface warming (blue in Figure 1a). The global cooling effect peaks at year 31, reaching -0.41 K, and then gradually weakens (Figure 1b). Note that this gradual weakening of the cooling effect is commonly reported in studies using time-invariant meltwater input (e.g., Park and Latif 2019; Oh et al. 2020), regarded as the compensation of subsurface warming with limited heat reservoir of the deep ocean (Martin et al., 2013; L. Zhang & Delworth, 2016). Nevertheless, the meltwater input always reduces global mean surface warming.

Meltwater-induced cooling is strongest in the SO following geographic adjacency, especially near West Antarctica (Figure 1e). This cooling extends to the entire Southern Hemisphere and tropics, even in the Arctic (e.g., Bronselaer et al., 2018). In addition, the tropical precipitation shows a distinct pattern: a zonal-mean northward shift and a weakening of walker circulation (green-brown contour in Figure 1e). The former has been relatively well-established since the early 2000s, first noticed by paleoclimate proxies and modeling experiments (e.g., Chiang & Bitz, 2005; Peterson et al., 2000). Theoretical studies suggest that the zonal-mean precipitation is shifted to the north which reduces the meltwater-induced interhemispheric energy asymmetry (e.g., Kang et al., 2018a). The latter is more related to the Walker cell response to the forcing. The employed model produces weakened convection at the warm pool to the Antarctic meltwater, which is consistently shown in other models imposing the Antarctic meltwater forcing (see Figure 2 in Bronselaer et al., 2018; Figure 2 in Oh et al., 2020).

In climate models, the AMOC is generally expected to weaken under global warming (e.g., Levang & Schmitt, 2020; Reintges et al., 2017), and the Antarctic meltwater is expected to counteract the weakening. Consistently, in our experiments, the meltwater input tends to strengthen the AMOC, measuring the maximum meridional stream function at 45°N, after the cooling peaks. However, although the meltwater ultimately mitigates the impacts of global warming on the AMOC (Figures 1c-d), the meltwater discharge unexpectedly weakens the AMOC beforehand (Figure 1d). Internal variability may obscure the weakening in some members. Nevertheless, it is evident that the meltwater-forced AMOC response depends significantly on the time scale, which is consistently shown in the SPNA temperature (Figure S1a). Considering that the temperature response generally retains its pattern with respect to the time (Figure S1b), the AMOC response to the meltwater-induced cooling shows not a simple linear but a non-monotonic relationship, which is counterintuitive to the large inertia of the ocean circulation that would be expected to produce a more gradual response to the perennial cooling.

To understand this non-monotonic response in detail, we look at the periods before and after the sign reversal of δAMOC, referred to as weak (year 6-20) and strong (year 31-45) periods (orange and gray shading in Figure 1). We first note that, although the AMOC responses are opposite, the zonal-mean ocean circulation response in the SO and tropics is somewhat similar (Figures 1f-g). In the southern extratropics, the freshwater input reduces the formation of Antarctic Bottom Water, while the cold surface intensifies a meridional temperature gradient and westerly winds, accompanied by enhanced Deacon cell (Park & Latif, 2019). In the tropics, the northward ITCZ shift indicates a weakening of the northern Hadley cell, and vice versa. Since the atmospheric and oceanic circulation are mechanically coupled through surface wind stress, subtropical cell responses mirror the changes occurring on the Hadley cell (Held, 2001), compensating meltwater-induced interhemispheric asymmetry (e.g., Green & Marshall, 2017; Kang et al., 2018b; Schneider, 2017). Only in the northern extratropics the oceanic circulation is dependent on the time scale. During the strong period, the AMOC generally strengthens (Figure 1g), which is consistent with previous studies reporting the long-term response of the AMOC to meltwater input and/or the surface cooling hole (Bronselaer et al., 2018; Oh et al., 2020; Park & Latif, 2019). However, just a few years after the freshwater forcing, the AMOC experiences a significant weakening. Note that it closely resembles the response of the AMOC to freshwater input from Greenland, the antipode of Antarctica (Figure 1f; Figures 6a-c in Li et al., 2023).

Focusing on the weak period, we examine the temporal evolution of upper-ocean density in the Atlantic basin (Figure 2). The density responses are most pronounced above 500 m, suggesting that upper-level stratification plays an important role in regulating the AMOC (Figure S2; Zhang et al., 2017). The Hovmöller diagram shows that the Atlantic upper-ocean density suddenly decreases at the northern high-latitudes at the beginning of the weak period, which corresponds to the weakening of the AMOC (Figures 2a and S2a). The density reduction is solely pronounced in the deep convection region of the AMOC, the Labrador Sea (Figure 2b), where even small perturbations can induce large AMOC responses (e.g., Stocker & Wright, 1991). Salinity plays a dominant role in driving the density decrease. Although cold temperature has the potential to increase the density (Figure 2g-h), it cannot overcome surface freshening (Figure 2d-e). Note that these regions have been proposed to be diluted by freshwater fluxes from Greenland (Gillard et al., 2016), which partially explains the aforementioned similarity with the results of the Greenland freshwater hosing experiment (Li et al., 2023).

The local stratification is eventually terminated by Atlantic-wide cooling. Although local freshening continues to weaken the AMOC (Figures 2d-f), thermally-driven density anomalies emerge from low-latitudes and propagate northward through the upper ocean, gradually overcoming the stratification at the deep convection region (Figures 2a-c and S2).

Temperature anomalies evolve along the Atlantic water pathway (Figure S3), implying the importance of climatological upper-ocean circulation in regulating SPNA convection (e.g., Piecuch et al., 2017). Within a few decades, the whole North Atlantic upper ocean, as well as the SPNA, is eventually de-stratified by thermal contraction. It is followed by a strengthening of the AMOC and a cessation of the weak periods. This time scale is comparable to previous results, showing that it takes a few decades for tropical salinity to spread throughout the North Atlantic (see Figure S12 in Hu & Fedorov, 2019), and we note that the near-surface (0-5 m) density evolution is almost consistent with that in the upper ocean (Figure S4). Taken together, the density profiles clearly indicate that some rapid response to the Antarctic meltwater input abruptly dilutes the deep convection regions near Greenland, leading to an unrecognized weakening of the AMOC.

Atmospheric processes are intuitively the most likely candidate for abrupt salinity reduction, not only because of their fast time scale but also because of the importance of large-scale atmospheric circulation in regulating salinity (e.g., Durack et al., 2012). Thermal forcing imposed in the SO could influence tropical climate through the lower troposphere, leading to changes in the tropical hydrological cycle within a few years (Kim et al., 2022). As the tropical convection is modulated, the resulting Gill-type response could rapidly perturb the extratropical climate via the wave train propagating poleward and eastward (Hoskins & Karoly, 1981). Hence, we examine the precipitation and 300hPa stream function response at the onset of the weak period (Figure 3a). Suppressed convection is detected in the western Pacific warm pool region, which excites upper-level cyclonic flow and the wave energy continues to propagate northeastward across North America. The associated wave activity flux (WAF) suggests that the wave propagation eventually gives rise to the anticyclonic circulation over Greenland, which projects onto a negative NAO pattern. The teleconnection is quite similar to that shown in previous studies which suggest a strong positive correlation between western Pacific convection and NAO response (e.g., Geng et al., 2023; Huntingford et al., 2014; Scaife et al., 2017). Note that the wave train is consistently shown in the boreal winter (DJF), when both the Rossby wave teleconnection and the SPNA deep convection are dominant (Figure S5). The NAO is known to modulate the AMOC intensity through the surface buoyancy response in the deep convection regions, a relationship corroborated by many climate models, including the employed model (e.g., Delworth & Zeng, 2016; Kim et al., 2023; Medhaug et al., 2012). In alignment with this, the presence of an anticyclone over

Greenland corresponds to a weakening of the surface westerly winds over the SPNA, resulting in a significant reduction of regional evaporation (Figure 3b). While regional precipitation shows a slight decrease with the anticyclone (not shown), the downward water flux at the surface–precipitation minus evaporation (P-E)–shows a significant increase mainly due to the evaporation reduction. This increase explains the salinity drop at the onset of weak periods, inducing feeble convection at the Labrador Sea (Figure 2a,e).

We further analyzed the 1000-year control experiment to consolidate the tropical-induced salinity decrease over the SPNA. To establish a link between the high-latitude circulation response to tropical convection, we initially remove the time-mean value from the 1000-year precipitation data, thereby representing interannual precipitation variability. Linear regression analysis is then performed for each year, regressing the interannual variability pattern against the meltwater-induced precipitation pattern (depicted by the dashed box in Figure 3a). As a result, the regression coefficient quantifies the spatial similarity of the precipitation variability to the target precipitation pattern, the meltwater-induced response. Therefore, we repeat linear regressions that the pattern regression coefficient is regressed with respect to the interannual variability of the 300 hPa stream function, surface wind speed, and evaporation at each grid, which allows us to extract the atmospheric variability when the tropical convection resembles the meltwater-induced response (Figures 3c-d).

The extracted precipitation pattern does not fit perfectly with that forced by the Antarctic meltwater (contour in Figure 3c), as the former mostly reflects zonal redistribution of diabatic heating, while the latter represents a combination of meridional shift and zonal redistribution. However, the western Pacific diabatic cooling induces a similar wave train that propagates northeastward to the SPNA, leading to anticyclonic circulation over Greenland (Figure 3c). In line with the change at the onset of the weak period, we also observe a subsequent decrease in both surface wind speed and evaporation (Figure 3d), although these changes are smaller than the forced response associated with the weaker anticyclone. The difference in background climates, one representing the present climate and the other the early stage of global warming, may contribute to the different amplitude in the circulation responses over Greenland. Nevertheless, strong correlations between forced and regression patterns, particularly 0.82 for evaporation and 0.71 for surface wind speed (dashed box in Figure 3b), highlight a spatial similarity over the Labrador and Irminger Seas. The spatial resemblance underscores the crucial role of atmospheric teleconnection in regulating the climate of the North Atlantic.

4. Summary and Discussion

In this study, we investigate the impact of Antarctic meltwater on the AMOC with particular interest in its temporal evolution. Previous studies have shown that Antarctic meltwater induces global cooling, and both the atmosphere and the ocean circulation adjust to it, such as the northward shift of the ITCZ and the strengthening of the Deacon cell (Bronselaer et al., 2018; Park & Latif, 2019). However, our study suggests that while the AMOC strengthening is stably detected after 30 years of simulation, it is unexpectedly weakened during the initial period by the Antarctic meltwater discharge, which makes a non-monotonic AMOC response to global cooling (Figure 4). The non-monotonic response is attributed to the differing timescales of atmospheric and oceanic teleconnections. The SO cooling could affect the tropical climate within a few years through near-surface propagation, resulting in convection redistribution (Kang et al., 2023; Kim et al., 2022). Then, the suppressed convection immediately triggers an upper-level cyclonic flow, which initiates Rossby wave propagation to the extratropics. Before the Atlantic gyre system brings tropical cooling into the SPNA convection region that is followed by the AMOC strengthening, the wave-induced anticyclone over Greenland, which is a polarity of the negative NAO pattern, weakens the surface westerlies, evaporation, and the strength of the AMOC. Therefore, the non-monotonic response of the AMOC, consistently shown as a tug-of-war in the SPNA upper-ocean density, is the result of competing influences between atmospheric teleconnection, which induces rapid and regional freshening, and oceanic propagation, which results in slow but strong thermal stratification.

We do not want to overemphasize our findings via the single model: the non-monotonicity examined would be model-dependent. For example, the AMOC is known as notorious spread among current climate models (e.g., Gong et al., 2022). Thus, even if the atmospheric teleconnection rapidly adjusts the AMOC, the magnitude of the adjustment would be highly variable across models (e.g., Kim et al., 2023). In addition, although the tropical response is similar to that shown in the studies imposing the Antarctic meltwater (Bronselaer et al., 2018; Oh et al., 2020), some models project a similar weakened convection in response to the Antarctic sea-ice loss, which accompanies with surface warming (e.g., Ayres et al., 2022; England et al., 2020c). Hence, the sensitivity of the tropical response to either forcing structures or model configurations would have a potential to modulate the non-monotonicity. All of these factors underscore the importance of the model intercomparison to the realistic Antarctic meltwater impacts. As the next generation of Coupled Model Intercomparison Project (CMIP7) is planned to include an interactive meltwater (e.g., Swart et al., 2023), further studies are warranted that carefully examine the meltwater-induced tropical response and consequent AMOC response.

Despite the plausible amount of meltwater under global warming (DeConto & Pollard, 2016), employing an abrupt and time-invariant injection may exaggerate the temporal evolution of the meltwater impacts. In the context of global warming, a gradual increase in meltwater discharge consistently induces relative cooling in the Southern Ocean, contributing to AMOC weakening by the atmosphere and strengthening by the ocean. However, the time scale of each adjustment would not be differentiated, resulting in blurred atmospheric impacts. It's important to note that wind variability substantially influences the Southern Ocean temperature (Roach et al., 2023), and constructive interference with the meltwater flux still holds the potential for non-monotonic AMOC responses. Overall, however, the non-monotonic evolution is less anticipated with gradual meltwater alone.

In other words, the non-monotonicity is more likely to occur when sudden and disastrous changes happen, which aligns with the growing concern about various tipping elements (McKay et al., 2022) and their interaction, referred to as tipping cascading (Wunderling et al., 2023). Our current understanding of these cascading impacts is still limited to conceptual and idealized models. For example, conceptual models that consider the oceanic transport alone propose the abrupt meltwater discharge from the West Antarctic Ice Sheet (WAIS) as a potential stabilizing factor for the AMOC (Sinet et al., 2023). However, our findings point out a new dimension that the WAIS could also provoke AMOC destabilization through the atmospheric pole-to-pole teleconnection. Note that there are alternative pathways that could link SPNA and tropical climate such as Indo-Pacific Ocean (e.g., Hu & Fedorov, 2020; Orihuela-Pinto et al., 2023). Considering the intricate nature of climate systems, further studies are warranted to carefully investigate the meltwater impacts on both gradual and abrupt climate change by employing a more realistic configuration. The primary purpose of this study, however, is to highlight the potential of non-monotonic AMOC response by the interplay between atmosphere and ocean pathways.

Acknowledgments

This research was supported by Korea Institute of Marine Science & Technology Promotion (KIMST) funded by the Ministry of Oceans and Fisheries (RS-2023-00256677; PM23020) and by the National Research Foundation of Korea (NRF) grant funded by the Korean government (NRF-2022R1A3B1077622)

Data Availability Statement

The processed data of the simulations used for this study is available in Shin (2023).

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Figure list

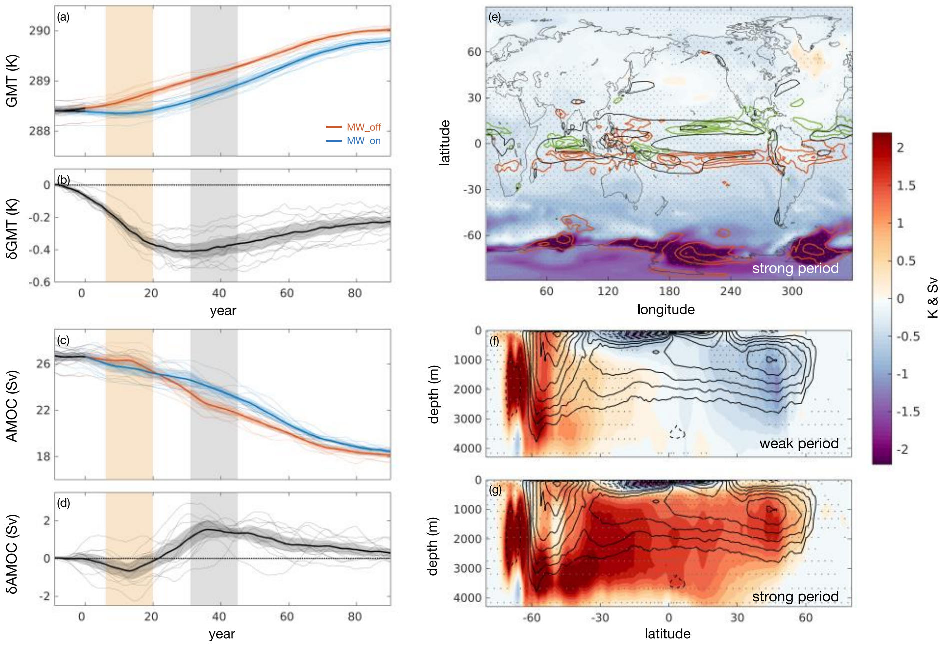


Figure 1. Time series of (a) global mean surface temperature. Red is the ensemble with CO_2 doubling only (MW_off), while blue is the ensemble with additional Antarctic meltwater discharge (MW_on). (b) Meltwater-induced response in global mean surface temperature (i.e., blue-red). (c,d) Same time series as (a,b), but for the AMOC index, measuring the maximum meridional stream function at 45°N. Every time series is smoothed by 21-yr running average. Each ensemble members are shown as a thin line, with its mean as a thick line. Shading indicates the statistical range at the 95% confidence level. Gray vertical shading indicates a strong period (year 31-45) regarding δAMOC, while orange shading indicates a weak period (year 6-20). (e) Meltwater-induced surface temperature (shading) and precipitation (interval = 0.13 mm day^{-1}; positive in green and negative in brown) anomalies at the strong period. Black solid contour is climatological precipitation (interval = 5 mm day^{-1}). (f) Response of the MOC to the meltwater (shading) at the weak period and (g) at the strong period. Climatological MOC is shown in solid-dashed contour (interval = 3 Sv; positive in solid and negative in dashed). Red shading and solid contour indicate clockwise circulation and vice versa. Dotted area indicates statistical significance at the 95% confidence level.

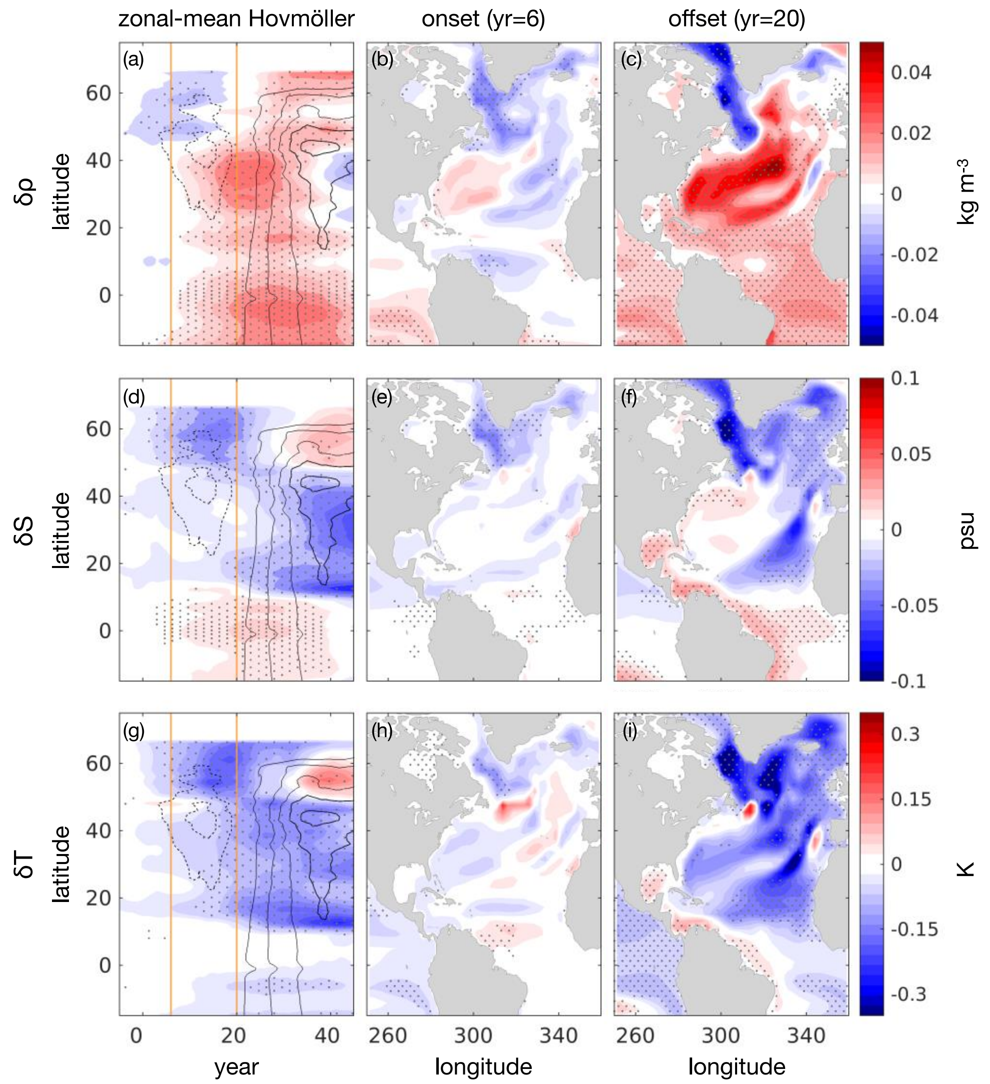


Figure 2. (left) Atlantic zonal-mean Hovmöller diagram of upper-level (0-500m) (a) density, (g) salinity, and (g) temperature. The meridional stream function at 1000 m is shown as a solid-dashed contour (interval = 0.3 Sv; positive in solid and negative in dashed). The onset and offset of the weak period are shown as orange lines. All variables are 21-year running averaged. (middle) Anomalous map for corresponding variable at onset and (right) offset. Dotted area indicates statistical significance at the 95% confidence level.

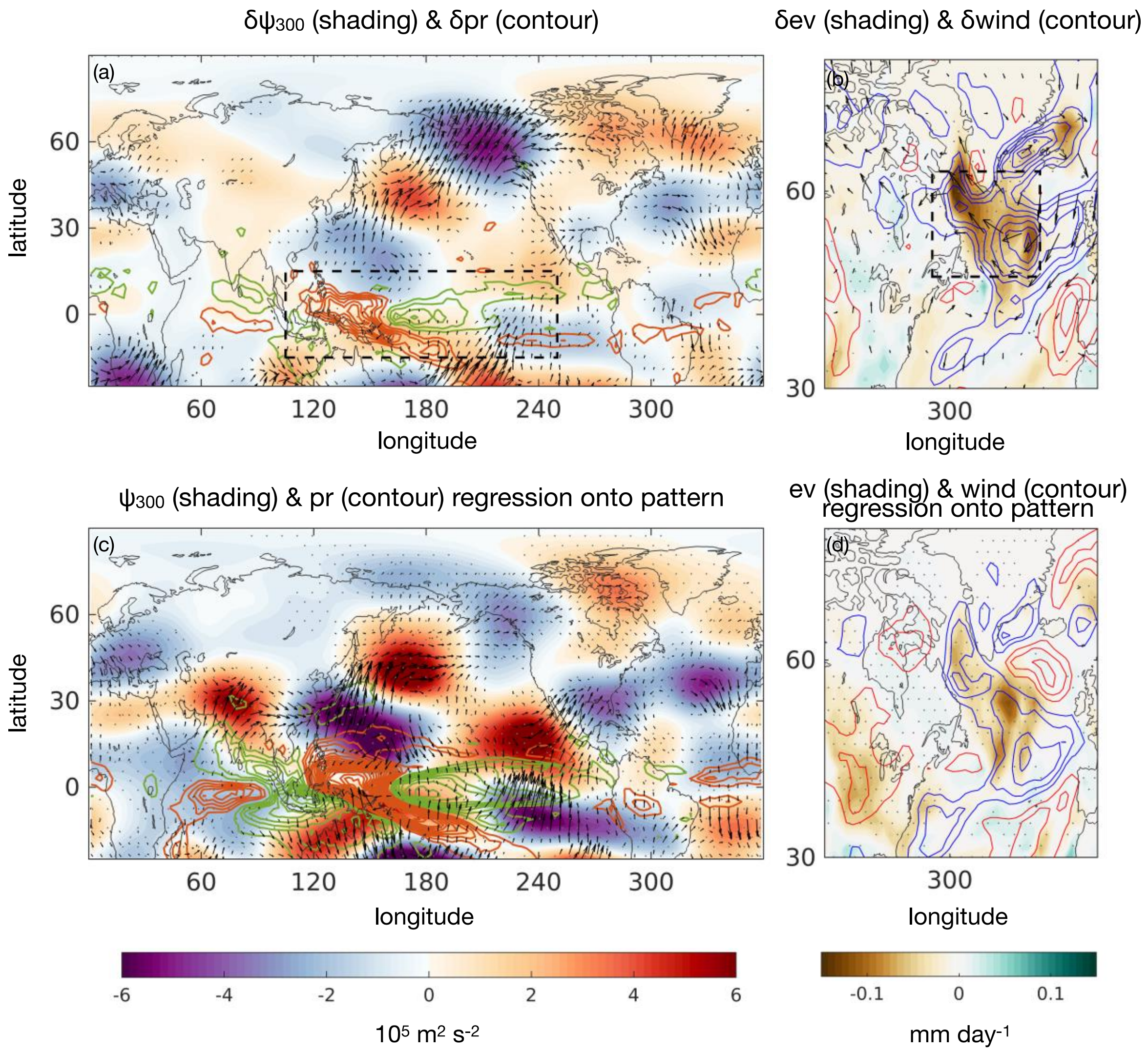


Figure 3. (a) Annual-mean stream function at 300 hPa (shading), precipitation (interval = 0.12 mm day^{-1}; positive in green and negative in brown), and wave activity flux (quiver) anomalies, and (b) evaporation (shading), surface wind speed (interval = 0.015 m s^{-1}; positive in red and negative in blue), and 10 m wind (quiver) anomalies to the Antarctic meltwater input at the onset of weak period (year 6 to 12). (c,d) Same maps as (a,b) but for the regression coefficient of each variable on the anomalous precipitation pattern. Dotted regions are statistically significant at the 95% confidence level. The dashed box in (a) indicates the anomalous precipitation pattern for regression analysis, and that in (b) indicates the Labrador and Irminger Seas.

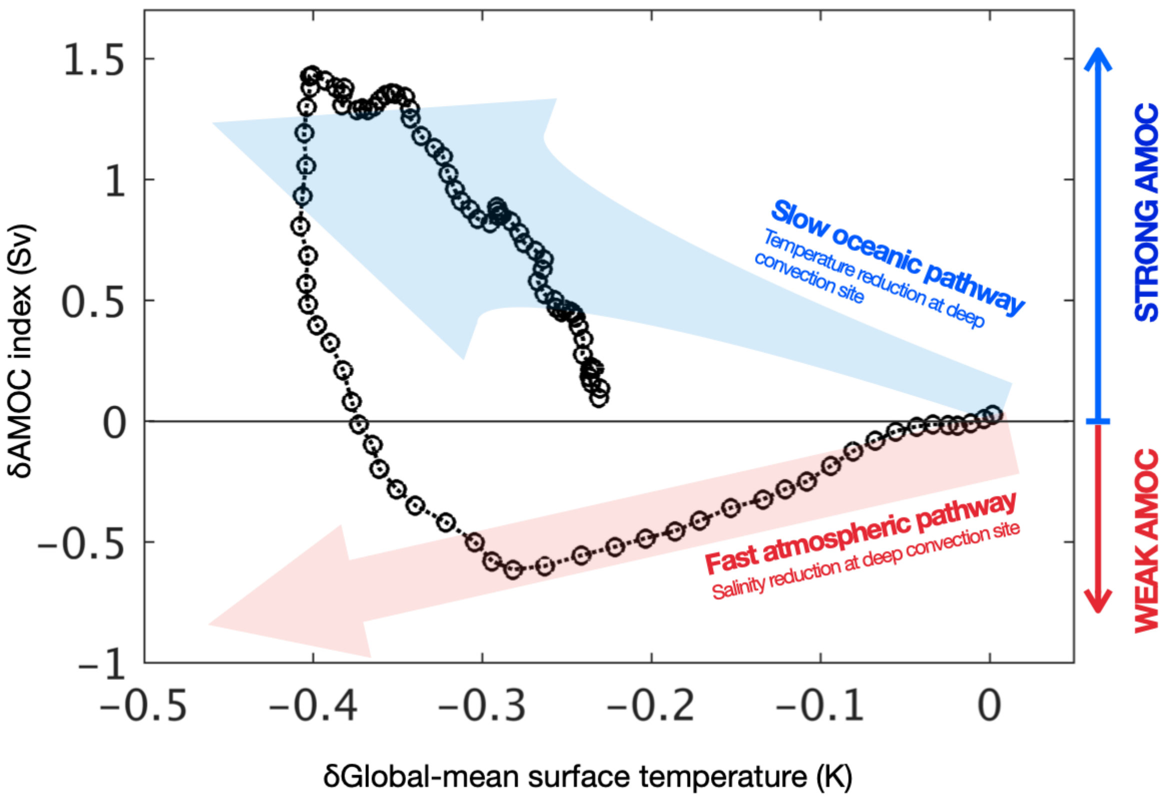


Figure 4. Meltwater-induced AMOC anomalies with respect to the global-mean surface cooling. Each circle indicates 21-year running-mean value. Although the global-mean temperature decreases by the meltwater hosing, the AMOC response is muted for the first few years corresponding to the time lag between Southern Ocean cooling and tropical response. Then, the AMOC response is not following the intuitive strengthening, but weakening which is driven by fast atmospheric pathway. As time goes by, slow but strong ocean pathway reaches the deep convection region, the AMOC exhibits abrupt transition under relatively consistent global-mean cooling.

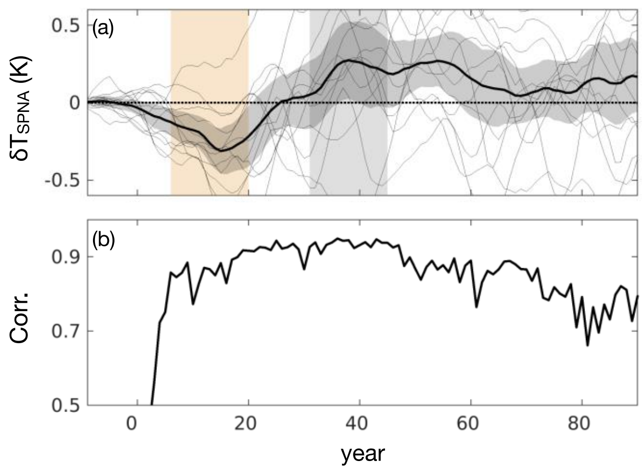


Figure S1. (a) Same time series as Figure 1b but for SPNA surface temperature (50°N~60°N, 300°E~320°E). (b) Time series of pattern correlation coefficient between δT_S at each year and that of strong period, shown in Fig. 1e.

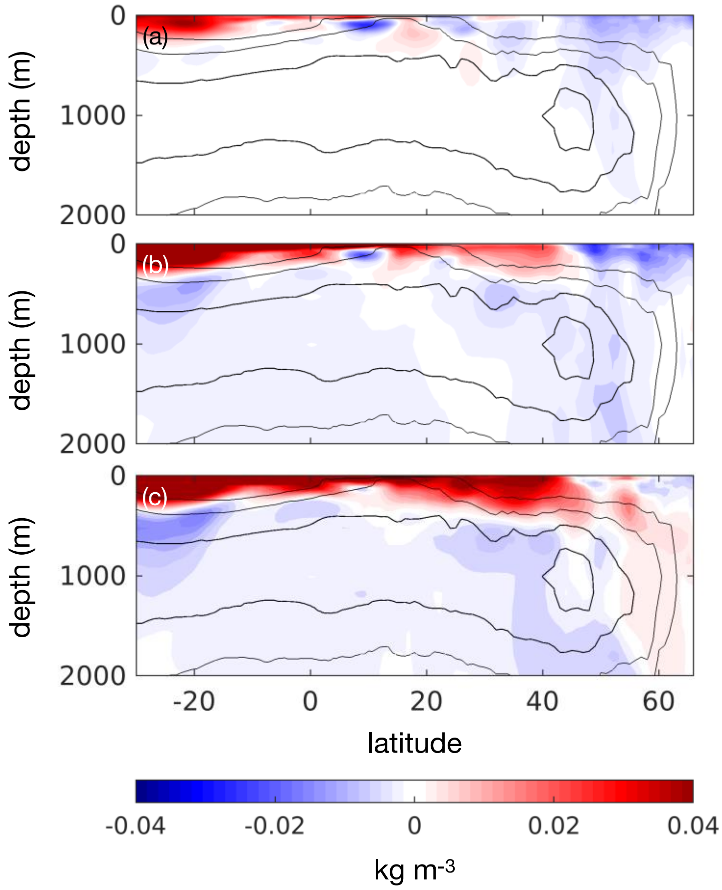


Figure S2. (a) Atlantic depth-latitude profile of the anomalous density response to the Antarctic meltwater at the onset, (b) middle, (c) offset of the weak period. Climatological mean meridional stream function is shown in solid contour (interval = 4 Sv).

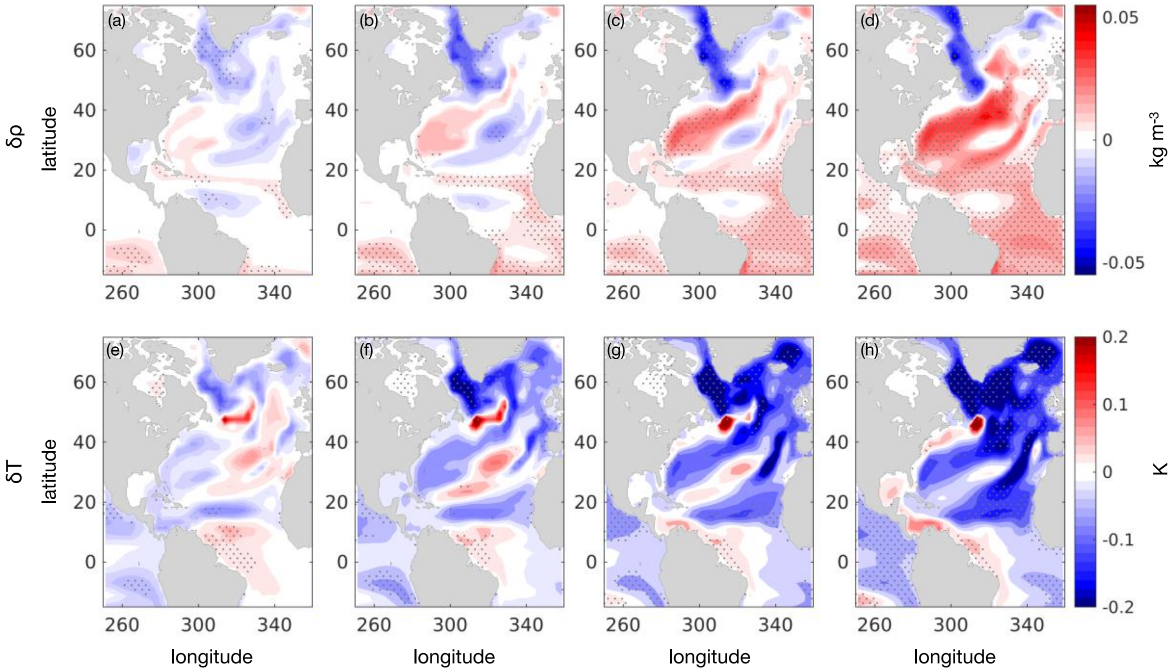
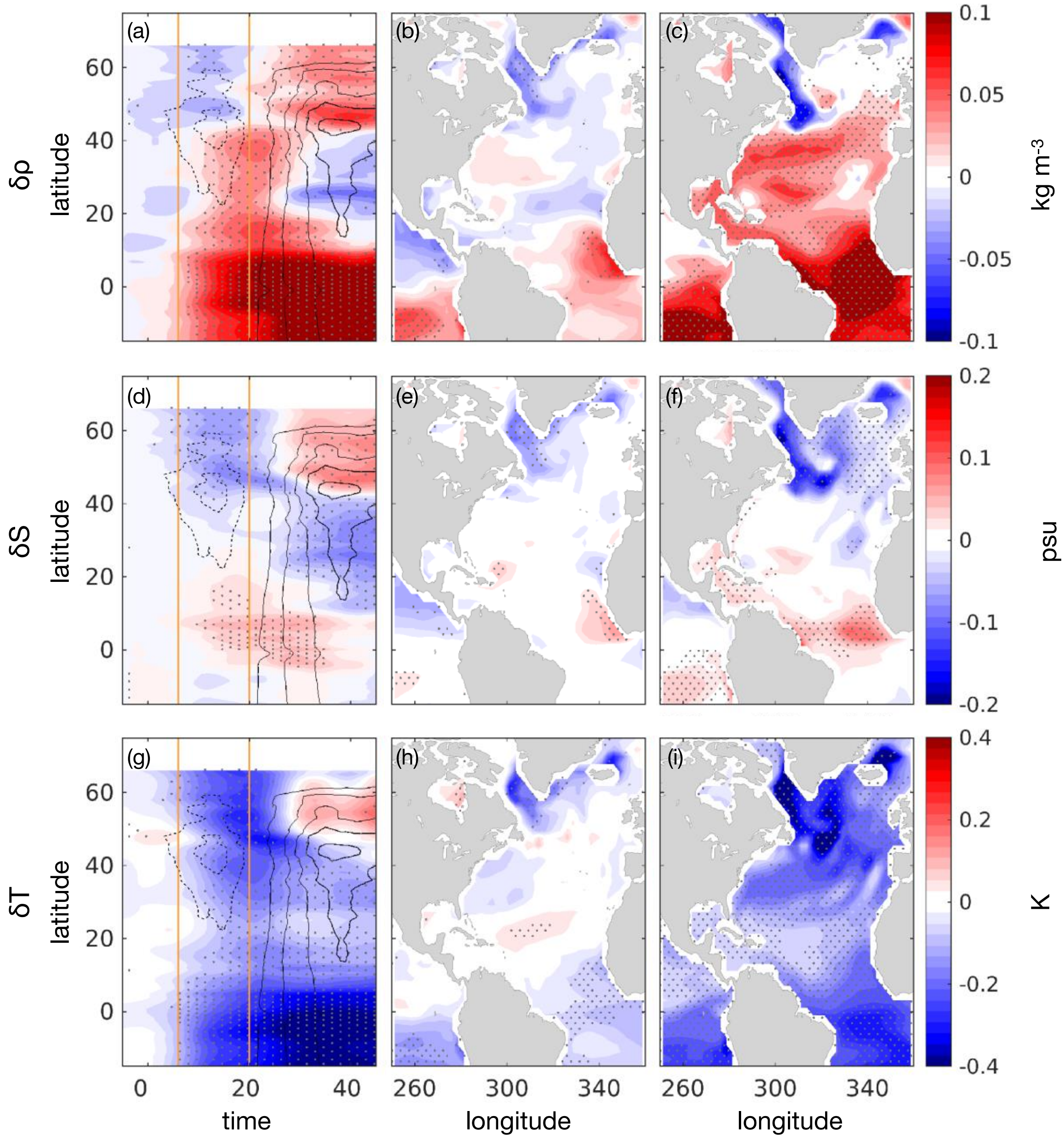


Figure S3. Anomalous map for upper-level (0-500m) density and temperature (a,e) of the onset of the weak period and (b-d,f-h) every 4 year from thereafter.



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604 **Figure S4.** Same as Figure 2 but for surface layer.

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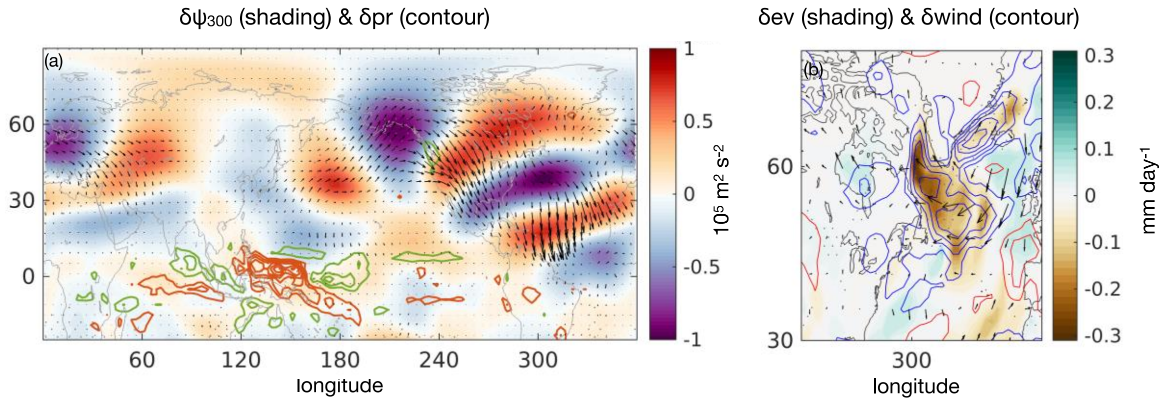

Figure S5. Same as Figure 3a and b but for the boreal winter (DJF).