The Seesaw of Seasonal Precipitation Variability Between North China and the Southwest United States: A Response to Arctic Amplification

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Abstract  Surface albedo over the ice-covered polar ocean is decreased in a highly idealized numerical experiment, which leads to stronger (weaker) warming at high (low) latitudes, and forms Arctic amplification. With stronger warming at both high and low latitudes, the meridional temperature gradient decreases (increases) at midlatitudes (subtropical regions). A cyclonic wind anomaly appears and results in a weak Kuroshio current and warm interdecadal Pacific Oscillation phase. In response to Arctic-amplified-warming, the subtropical high weakens, leading to summer monsoon anomaly on its southside and increased westward transport of water vapor, resulting in a seesaw pattern: dry (wet) conditions in summer in North China (southwest United States, SWUS). In winter, the East Asian trough (North American high) decreases and induces decreased (increased) cold air transport to North China (SWUS) in response to Arctic-amplified-warming, which makes rain unlikely (likely) and exhibits a seesaw pattern. Moisture convergence always appears in a warmer SWUS winter, and increases precipitation there, which strengthens (counteracts) the precipitation anomaly led by the North American high in a warm (cold) interdecadal Pacific Oscillation phase. It may even shut down the seesaw pattern in winter in the cold interdecadal Pacific Oscillation phase.

Plain Language Summary  Polar warming led by smaller surface albedo decreases oceanic circulation and oceanic heat transport while enhancing atmospheric circulation and atmospheric heat transport, resulting in global warming, which could also be achieved by increased carbon dioxide (CO2) forcing. However, there are many differences between the midlow latitude climate response to polar albedo forcing and that to increased CO2 forcing. (1) In polar albedo forcing-induced global warming, North China becomes drier, and the southwest United States becomes wetter, which is opposite to the results in the increased CO2-induced global warming. Both results suggest a seesaw pattern of precipitation variability between North China and the southwest United States under climate change. (2) The seesaw pattern also appears in different seasons. In summer, the seesaw pattern is maintained by subtropical high. The summer monsoon determines the water vapor distribution. In winter, the seesaw pattern is maintained by wave-flow interaction, which maintains synchronous variations between the East Asian trough and North American high. (3) Moisture convergence in winter southwest United States (under global warming) enhances local precipitation. As a result, it strengthens the seesaw pattern in this study, while it may shut down the seesaw pattern in increased CO2-induced global warming.

1. Introduction

The global surface temperature has increased over the last two centuries in response to forcing resulting from anthropogenic greenhouse gas (GHG) emissions (mainly increases the carbon dioxide [CO2] concentration), with a characteristic pattern of Arctic-amplified-warming, or “Arctic amplification” (Screen & Simmonds, 2010). Previous studies suggested that Arctic amplification is a result of local forcing and feedback (Stuecker et al., 2018). Increased CO2 forcing is considered as the trigger for global warming in recent centuries (Dai et al., 2019). However, Dai (2021) argues that direct effect of increased CO2 forcing may decrease Arctic amplification by inducing stronger warming at low latitudes. According to estimates with radiative kernel method (Dai, 2021; Goosse et al., 2018; Graversen et al., 2014; Huang et al., 2017), positive feedback of surface albedo dominates the local forcing and feedback in polar regions (Dai, 2021), which induces large amounts of solar radiation absorption at the sea surface (in polar ocean). Surface temperature...
feedback (Planck feedback) is negative in polar regions and is smaller than that in the tropics (Pithan & Mauritsen, 2014). The lapse rate in polar regions leads to a weak positive feedback due to stable stratification (Goosse et al., 2018). Due to surface warming, vertical convection, and water vapor increases, more clouds appear. Cloud feedback can be positive or negative due to different cloud types and heights (Huang et al., 2017). Water vapor feedback in the polar region can be neglected due to a small magnitude of moisture (Dai et al., 2017).

Arctic amplification also influences both the weather and climate at mid-low latitudes (Park et al., 2018; Smith et al., 2019). The intensity and position of the polar vortex changes in response to Arctic-amplified-warming, which transports more cold air to lower latitudes and produces extremely cold events (Kug et al., 2015; Luo et al., 2019). The polar vortex also determines the jet strength and jet position, which may further influence the precipitation and eddy activity at lower latitudes (Coumou et al., 2018; Oudar et al., 2017).

However, all these conclusions have been drawn from increased CO$_2$-induced global warming. In that case, climate anomalies in mid-low latitudes are induced by both increased CO$_2$ forcing (local forcing) and Arctic-amplified-warming (remote forcing). The former changes local radiation flux immediately; the latter changes the meridional circulation and heat transport, which induces climate responses at mid-low latitude. To isolate the influence of Arctic-amplified-warming, we artificially change the surface albedo in the polar region to achieve global warming and Arctic amplification. Without increased CO$_2$ forcing, the local radiation flux (in mid-low latitudes) remains nearly unchanged in the beginning. In this case, climate anomalies in the midlow latitudes are responses to Arctic-amplified-warming.

Since the mechanism of Arctic amplification formation varies in different seasons (Dai, 2021), seasonal variability in climate response (i.e., precipitation) should be carefully considered. In increased CO$_2$-induced global warming, the summers in North China become increasingly wet because of a stronger subtropical high due to the enhanced temperature gradient between the Indian Ocean and west Pacific (He & Zhou, 2015), while the summers in the southwest United States (SWUS) become increasingly dry because of increased atmospheric stability due to monsoon variation (Pascale et al., 2017). The seesaw pattern of annual precipitation variability is consistent with that during a cold phase of the interdecadal Pacific Oscillation (IPO) (Power et al., 1999). During the IPO cold phase, the positive sea surface temperature (SST) anomaly is stronger in the western North Pacific than that in eastern North Pacific, while the positive SST anomaly is weaker in western North Pacific than in the eastern North Pacific during the IPO warm phase.

In this study, we explore whether Arctic-amplified-warming will also maintain a seesaw pattern in annual precipitation variability as in increased CO$_2$-induced global warming. This study also evaluates whether the seesaw pattern is maintained in different seasons. To answer these questions, this study is organized as follows: The coupled model and perturbation experiments are introduced in Section 2. We analyze the results from the polar albedo forcing experiment in Section 3. A summary and discussion are provided in Section 4.

2. Model and Experiments

The model used in this study is the Community Earth System Model (CESM) version 1.0 of the National Center for Atmospheric Research, which was used in our previous studies on global energy balance (Dai et al., 2017; Yang & Dai, 2015). Here, we briefly introduce the model setup. CESM 1.0 is a fully coupled global climate model that provides state-of-the-art simulations of Earth’s past, present, and future climate (Park et al., 2014; Wang et al., 2015). The model grid employed in this study is T31_gx3v7. The atmosphere model is the Community Atmosphere Model (CAM5; Park et al., 2014), which has 26 vertical levels, with a finite volume of 3.75° × 3.75° in the horizontal direction. The land model is the Community Land Model (CLM4; Lawrence et al., 2012), which has the same horizontal resolution as CAM5. The ocean model is the Parallel Ocean Program (POP2; Smith & Gent, 2010), which has 60 vertical levels. The horizontal grid is a nominal 3° grid with the North Pole displaced to land. The ice model is the Community Ice CodE (Hunke & Lipscomb, 2008), which has the same horizontal grid as POP2. More details can be found in our previous studies (Dai et al., 2017; Yang & Dai, 2015). Although planetary wave and polar vortex are well resolved in our study (Kim et al., 2014; Luo et al., 2019; Yin et al., 2019), we suggest that reviewers may employ a finer resolution in synoptic scale discussion.
The experiments analyzed in this study include a 2000-years control run (CTRL) and a 500-years surface albedo-perturbation run (0.1 A). The CTRL starts from a state of rest, as in Dai et al. (2017). The CO₂ concentration in CTRL is set to 285 ppm (no increased CO₂ forcing). The model climate reaches a quasi-equilibrium after 1000 years of integration (Yang & Dai, 2015). In the surface albedo-perturbation run, the CO₂ concentration is also set to 285 ppm. However, the surface albedo (to shortwave radiation) over the globally ice-covered ocean is set to nearly the same as that over the ice-free ocean (by setting the sea ice albedo to 10% of its original value; land ice albedo is unchanged). The surface albedo in the other parts of the coupled climate model is not artificially changed. The dynamics and thermodynamics of sea ice and snow are not artificially changed. The inspiration for this idealized albedo-perturbation experiment comes from melt ponds (Hunke et al., 2013), which appear as a result of ice melt only at the surface of sea ice during global warming and largely decrease the surface albedo of ice-covered ocean. We assume the melt ponds concentration is underestimated in the CTRL, and thus we artificially decrease the surface albedo in the albedo-perturbation experiment. The albedo-perturbation experiment starts from year 1500 of the CTRL and reaches quasi-equilibrium after 150 years of integration. The monthly mean outputs are used for analysis. The climate mean is the seasonally/annually averaged field over the last 200-years integration, unless stated otherwise. Based on radiative kernel technique (Dai, 2021; Huang et al., 2017), at the top of atmosphere (TOA), global mean instantaneous radiative forcing induced by surface albedo in 0.1 A is estimated as 2.6–3.1 W/m², which is only 85% of radiative forcing induced by CO₂ (3.1–3.6 W/m²) in double CO₂ scenario (2CO; Text S1). However, CO₂ radiative forcing in 2CO spread globally, while surface albedo radiative forcing in 0.1 A mainly spread over the polar regions. On the other hand, the net global mean TOA radiative imbalance during the last year of CTRL, 0.1 A and 2CO are only 0.21, 0.24, and 0.46 W/m², respectively, which suggests that the model is in radiative balance in all experiments.

3. Seasonal Response to Arctic Amplification

Although there is no increased CO₂ forcing in either CTRL or 0.1 A experiments, Arctic amplification is still achieved in the 0.1 A experiment. Important reasons are revealed in Section 3.1, with zonal mean and annual analysis. On the other hand, Arctic amplification largely influences the climate in midlow latitudes. An interesting one is the precipitation anomaly in North China and the SWUS, which is discussed in Section 3.2 with seasonal analysis.

3.1. Arctic Amplification due to Surface Albedo Perturbation

Decreased surface albedo artificially induces additional solar radiation to be absorbed in the polar region (nearly 13 and 12 W/m² in both Arctic [66°N–90°N] and Antarctic [66°S–90°S], respectively), which leads to polar ocean warming and sea ice melt. In the North Atlantic Ocean, the ice edge, which is defined as the region where annual mean sea ice covers 15% of the ocean area (Figure 1a, green line and red line), retreats northward to Greenland, leaving...
half of the Labrador Sea exposed as ice-free sea in the 0.1 A experiment. Fresh water from ice melt freshens salty water in the upper layers, which decreases the density by as much as 0.6 kg/m^3 in the Labrador Sea (Figure 1a, shading) and impedes deep-water formation (precipitation, evaporation, and river runoff play less important roles in decreasing deep-water formation in this case). As a result, the Atlantic Meridional Overturning Circulation (AMOC) nearly decreased by 20% in the 0.1 A experiment (Figure 1b, blue line). It required nearly 28 years for the AMOC to reach equilibrium in 0.1 A. Although the Subtropical Cell remains nearly unchanged (not shown), less energy (nearly 0.34 PW, 1 PW = 10^{12} W, Figure 1c, blue line) is transported poleward by ocean circulation with decreasing AMOC. On the other hand, the oceanic heat transport (OHT) carried by mesoscale or submesoscale activity remains nearly unchanged due to parameterization (not shown). In the midlow latitudes, local radiative forcing is nearly unchanged in the beginning (Figure S1), while poleward ocean heat transport (Figure 1c, blue line) decreases. It leads to more heat is conserved locally and results in a warmer surface (Figure 1c, green line). The heated sea surface enhances vertical convection (upwelling) in the atmosphere at lower latitudes, which strengthens the Hadley cell (HC) by as much as 11% (Figure 1b, red line) and enhances poleward atmospheric heat transport (AHT, Figure 1c, red line). The AHT and OHT varied in the region from 0°N–20°S with nearly the same magnitude but the opposite signs, compensating each other well (the so called Bjerknes Compensation, Bjerknes, 1964).

To better understand the distribution of surface warming at different latitudes, a zonal heat budget is analyzed. At high latitudes ([90°S–60°S] and [60°N–90°N]), strong local forcing due to additional solar radiation absorption overcomes the decreased poleward energy transport (Dai, 2021) and induces a strong surface warming (reach the maximum of 11°C close North Pole, Figure 1d green line). At low latitudes (10°S–20°N), the rather long time for radiation to achieve equilibrium (Figure S1) suggests that energy balance in midlow latitudes is broken mainly due to the decreased poleward energy transport, which is much smaller than the local radiative forcing at high latitudes, thus resulting in weaker surface warming (nearly 0°C in 20°N–40°N, Figure 1d green line). As a result, Arctic-amplified-warming appears. To better understand the distribution of surface warming at midlow latitudes, a meridional gradient of the heat transport anomaly (\(d(\Delta NET / dy)\)) is employed in the following discussion: Where \(\Delta NET = \Delta AHT + \Delta OHT\) is the net heat transport (NET) anomaly (Figure 1c, black line), \(\Delta AHT\) (Figure 1c, red line) is the AHT anomaly, and \(\Delta OHT\) (Figure 1c, blue line) is the OHT anomaly. The local net energy budget (\(\Delta S\)) at lower latitudes is written as \(\Delta S = F + (dNET / dy)f / [f]\), \(F\) is the net radiative forcing, which is nearly unchanged at lower latitudes at the beginning of our experiment (Figure S1), and \(f\) is the Coriolis parameter. In an equilibrium state, we have \(\Delta S = 0\). However, when the heat transport changes due to large-scale circulation adjustment, the local energy balance is broken, which leads to warming or cooling. If the gradient of heat transport increases (decreases), there is additional heat convergence (divergence) due to the large-scale circulation (Trenberth & Caron, 2001; Trenberth & Solomon, 1994). However, poleward heat transport is positive (negative) in the Northern (Southern) Hemisphere, which means that the gradient of heat transport increases due to a positive \(d(\Delta NET) / dy\) in the Northern Hemisphere or a negative \(d(\Delta NET) / dy\) in the Southern Hemisphere. Thus, a negative gradient of \(\Delta OHT\) in (40°S–10°S) (Figure 1c, blue line) leads to heat convergence by large-scale circulation and results in a larger surface warming (Figure 1c, green line), while a negative gradient of \(\Delta AHT\) in (20°N–40°N) leads to heat divergence due to large-scale circulation and results in a nearly unchanged surface temperature. As a result, the surface warming in (40°S–10°S) is larger than that in (20°N–40°N) (Figure 1c, green line).

In the midlow latitudes of the Northern Hemisphere, \(d(\Delta NET) / dy\) is nearly unchanged. However, surface warming in the tropics may trigger positive feedback because of water vapor and low cloud formation (Dai et al., 2017), which results in stronger (weaker) warming in the tropics (midlatitudes).

Here, we briefly summarize the mechanism of Arctic amplification in 0.1 A experiment (Figure 2). Decreasing the surface albedo over the polar ocean absorbed more solar radiation, which led to polar warming. Sea ice melts and freshens the surface water, making it more difficult to sink. As a result, the AMOC weakens and the poleward OHT decreases, making it warmer at lower latitude. However, warming at the lower latitude is much weaker than polar warming, since the decreased NET is much smaller than the solar radiation absorption in the polar ocean (Dai, 2021), which forms Arctic amplification. On the other hand, warming in the tropics enhances the Hadley Cell. However, the Hadley Cell strengthens mainly in the north branch
and expands southward, weakening the south branch, which means that the AHT increases in the midlow latitudes of the Northern Hemisphere, and decreases in the Southern Hemisphere. As a result, increased AHT and decreased OHT make the weakest warming in (20°N–40°N), while decreasing both in the OHT and AHT makes a stronger warming in (40°S–10°S).

### 3.2. Seesaw Pattern of Seasonal Precipitation Variability in Response to Arctic Amplification

In this subsection, we focus on climate change in the Northern Hemisphere. As discussed in the previous subsection, surface warming in the 0.1 A experiment (Figure 3a) reaches the maximum at high latitudes due to additional radiative forcing, which is stronger than the meridional heat transport anomaly (Dai, 2021). On the other hand, surface warming may easily trigger positive feedback in the tropics and result in stronger (weaker) warming in the tropics (midlatitudes). As a result, the meridional temperature gradient, decreases at midlatitudes and increases in the subtropical regions, resulting in a cyclonic wind anomaly over the North Pacific (Figure 3b, vector) via thermal wind relationship. Due to the wind stress curl anomaly, the Kuroshio current and Kuroshio extension (Figure 3b, shading) decrease by as much as 7.3 Sv (1 Sv = 10^6 m^3/s) via the Sverdrup relation (Sakamoto et al., 2005). Poleward-transport heat carried by the Kuroshio current decreases. As a result, the Northwest Pacific warms slightly, and a cold anomaly is even found near Japan in the albedo-perturbation experiment (0.1 A). On the other hand, the decreased eastern boundary current (i.e., California current) warms the east Pacific by transporting less cold water southward. Eventually, a warm IPO phase appears in response to Arctic-amplified-warming (0.1 A). Consistent with the conclusion drawn by Yang et al. (2019), the warm IPO phase induces an anomalous cyclone over the North Pacific, which brings more (less) moisture and rainfall to the SWUS (North China), and a seesaw pattern appears in the albedo-perturbation experiment. The seesaw pattern is in the opposite phase as that in the increased CO$_2$-induced global warming (North China gets wetter and SWUS gets drier, Yang et al., 2019), in which a cold IPO phase is found. However, when we consider seasonal precipitation variability, the seesaw pattern does not always appear, it may shut down in some scenarios. Thus, we attempted to determine why the seesaw pattern was maintained or shut down.

In summer, precipitation in North China is highly dependent on the East Asian monsoon, which transports water vapor northward and mainly follows the Southerly on the westside of the subtropical high (SH). In the albedo-perturbation experiment (0.1 A), the summer SST anomaly (Figure S3a) shares a similar pattern as the annual SST anomaly. Due to the same mechanism discussed in the previous paragraph, the negative meridional temperature gradient decreases (increases) in the region of (40°N–50°N) (20°N–40°N) in the Pacific (Figure S3a), which leads to the easterly (westerly) anomaly in the upper layer over the North
As a result, the cyclonic wind anomaly dominates over the central Pacific, which counteracts the anticyclonic wind field of the SH, makes the SH decrease (Figure 4a, shading), and retreats southward, leading to moisture divergence (Figure 4b, shading) and dry summers in North China (Figure 4c, shading). On the other hand, the westerly anomaly on the south side of the SH brings moisture to the SWUS (Figure 4b, vector), and moisture convergence occurs (Figure 4b, shading) due to stream convergence near the Rocky Mountains. As a result, the SWUS experiences a wet summer (Figure 4d, shading). The dry conditions in North China and the wet conditions in the SWUS result in a seesaw pattern of summer precipitation variability, which is in the opposite phase as in the increased CO$_2$-induced global warming: a decreased meridional SST gradient (not shown) at midhigh latitudes in the North Pacific results in an anomalous anticyclone over the North Pacific, which leads to wet conditions in North China (He & Zhou, 2015) and dry conditions in the SWUS (Pascale et al., 2017). It seems that the seesaw pattern is well maintained by the SH and monsoons in summer.

In winter, precipitation in North China is highly dependent on the East Asian trough, which brings cold air from high latitudes, while the cold air induced by the North American high may influence precipitation in the SWUS. The East Asian trough and North American high are two patterns of quasi-stationary planetary wave—which is formed from topographical forcing and surface thermal forcing (He et al., 2014; Matsuno, 1971; Tung & Lindzen, 1979)—and exhibit over East Asia and North America, respectively. In winter with high (planetary) wave activity, the polar vortex becomes more stable and colder, which weakens the East Asian trough and North American high (Chen et al., 2005). In contrast, polar vortex becomes more unstable and warmer in winter with low (planetary) wave activity, which strengthens the East Asian trough and North American high. In the albedo-perturbation experiment (0.1 A), North China becomes drier (Figure 5a) and the SWUS becomes wetter (Figure 5b) in winter. The mechanism can be explained as follows. The winter SST anomaly (Figure S3b) shares a similar pattern as the annual SST anomaly. As discussed in

Figure 3. SST anomalies (shading and contour, unit: °C) in the 0.1 A experiment are shown in (a). The warm colors and solid line indicate a positive anomaly, while the cold colors and dashed line indicate a negative anomaly. The surface wind stress anomaly (vector, unit: dyne/cm$^2$) and barotropic stream function anomaly (shading, unit: Sv, 1 Sv = 10$^6$ m$^3$/s) are shown in (b). SST, sea surface temperature.
the previous two paragraphs, stronger surface warming in the tropics (due to positive feedback) and weaker surface warming at midlatitudes result in an enhanced meridional temperature gradient (negative) in the region of (20°N–40°N), which leads to a strengthened westerly jet in the upper troposphere (Figure 5c, vector) due to the thermal wind relationship. On the other hand, enhanced HC, which is led by meridional temperature gradient anomaly at low latitudes, also enhances the westerly jet. Stronger zonal flow (westerly jet; $U$ in Equation 1; Figure 5c, contour) increases the frequency of quasi-stationary planetary wave ($\sigma$ in Equation 1, as well as the group velocity [$C_g$]). As a result, a larger group velocity decreases the planetary
wave amplitude ($A$ in Equation 2) and the planetary wave activity ($\mathcal{E}$ in Equation 3; Peloskly, 1987). Note, Equations 1–3 are presented in Appendix A. To better demonstrate how wave-flow interact with each other, Eliassen-Palm (EP) flux $\langle u' v' \rangle - \langle u' \rangle \langle v' \rangle$ and its divergence should be shown in meridional cross section (Charney & Drazin, 1961), where $u$ and $v$ are horizontal velocity, $N$ is Brunt-Väisälä frequency, $b$ is the buoyance, $\overline{\theta}$ is zonal mean, and $\theta = \overline{\theta} - \overline{\phi}$ is disturbance. However, all variables used in EP flux analysis should be daily data, which are not included in our output.

The quasi-stationary planetary wave anomaly also influences the zonal flow in the strength and location (Dickinson, 1969). Eventually, (the decreased amplitude of quasi-stationary planetary) wave-(the enhanced

Figure 5. The winter (December-January-February) precipitation anomaly (shading, unit: mm) in North China (marked with a black rectangle) are shown in (a), while the variations in winter precipitation (shading, unit: mm) in the southwest United States (marked with a black rectangle) are shown in (b). The westerly jet anomaly (vector, unit: m/s) and zonal velocity disturbance ($u' = u - \overline{u}, \overline{u}$ is zonal mean of zonal velocity, shading, unit: m/s) in 0.1 $A$ experiment are shown in (c). The moisture flux (vector, unit: kg/m/s) anomaly and moisture divergence (shading, unit: mm/day) anomaly are shown in (d).
mean) flow interaction leads to a weakened synoptic system (East Asian trough and North American high). The weakened East Asian trough (the wind fields at the 500 hPa and 200 hPa layers are similar) induces lessened cold air transport to North China, which makes rain difficult and leads to a dry winter (Figure 5a, shading). On the other hand, the weaker North American high retreats westward and allows more cold air to be transported along the edge on its eastside and southside (Figure 5c, vector), making rain in the SWUS easier and leading to a wet winter in the SWUS. The seesaw pattern is maintained by the quasi-stationary planetary wave, which synchronously adjusts the East Asian trough and North American high and makes the opposite contribution to precipitation between North China and the SWUS. Similar to the pattern of the annual SST anomaly, the winter SST anomaly is stronger in the high latitudes (tropics) than in the midlatitudes due to heat release (Dai, 2021) from the subsurface ocean (due to positive feedback), which leads to negative meridional temperature gradient decreases (increases) in the midlatitudes (subtropical regions). As a result, an anomalous cyclone appears over the North Pacific due to the thermal wind relationship. Wind field anomalies lead to moisture convergence anomalies occurring over the SWUS in winter (Figure 5d, shading), which provides promising conditions for rain in our experiment. If moisture convergence also occurs in a cold IPO phase (i.e., increased CO₂-induced global warming, not shown), the seesaw pattern may be shut down. In a cold IPO phase, the quasi-stationary wave amplitude increases, and the East Asian trough induces more cold air to North China, resulting in a wet winter. Although an enhanced North American high decreases the cold air leading to the SWUS, the increased moisture convergence may also enhance precipitation. As a result, precipitation is determined by the competition between decreased cold air and increased moisture convergence. If moisture convergence succeeds, the North American high will lose primary control on precipitation, which would shut down the seesaw pattern.

A summary of how the seesaw of precipitation variability between North China and the SWUS forms in summer and winter is shown in Figure 6. In summer, the SST in the high latitudes (tropics) increases more than that in the midlatitudes due to additional shortwave absorption (positive feedback), which leads to a meridional temperature gradient decrease (increase) and anomalous cyclone over the Pacific. As a result, subtropical high decreases and transports more (less) moisture to SWUS (North China), which leads to more (less) rainfall there. In winter, stronger meridional temperature gradient (similar reason as in summer, though high latitude warms due to heat release from subsurface ocean) leads to anomalous cyclone and weaker planetary wave activity. Wind anomaly enhances moisture and rainfall in SWUS. Planetary wave anomaly appears as weaker East Asian trough and North American high. The former (later) induces less (more) cold air to North China (SWUS), and leads to less (more) rainfall there. In conclusion, a seesaw pattern of precipitation variability appears both in summer and winter under the condition of Arctic amplification.
Furthermore, the seesaw pattern sometimes appears in spring and/or autumn, while it disappears in other scenarios. Either spring or autumn is a transition period between trough/ridge-controlled precipitation (winter) and SH/monsoon-controlled precipitation (summer), and the situation becomes more complicated. We would like to study these patterns in our future work.

4. Conclusions and Discussions

This work delved into “how Arctic-amplified-warming influences the climate at low latitudes” using a coupled climate model. The results show that polar warming may induce a temperature increase at low latitudes, although the temperature anomalies are much smaller than those in the increased CO\textsubscript{2} forcing scenario. Arctic-amplified-warming also decreases the strength of the Kuroshio current and leads to reduced warming (even cold conditions) in the western Pacific. The precipitation decreases (increases) in North China (SWUS) in response to Arctic-amplified-warming and appears to exhibit a seesaw pattern.

In this study, compared to previous studies, we discussed “the seasonal response to polar-amplified warming” to a greater extent in several ways.

First, sea ice albedo is artificially changed in this study, and Arctic-amplified-warming forms due to the additional absorption of solar radiation. In this way, we exclude the influence of increased CO\textsubscript{2} forcing, which means the climate change at lower latitudes is mainly the response to polar warming instead of local increased CO\textsubscript{2} forcing. Arctic amplification in the surface albedo-perturbation experiment also confirms the conclusion drawn by an earlier study (Dai et al., 2019) that sea ice loss (mainly the surface albedo decrease) serves as the main reason for Arctic amplification (more discussion is offered by Dai, 2021).

Second, the seesaw pattern of summer precipitation variability between North China and the SWUS should always appear, since these conditions are determined by the monsoon led by the southerly and easterly jets on the west side and south side of the SH. When the SH weakens, the southerly (easterly) jet decreases and transports less (more) water vapor to North China (SWUS), leading to a dry (wet) summer in North China (SWUS). In winter, the quasi-stationary planetary wave synchronously adjusts the East Asian trough and North American high, with opposite influences on precipitation in North China and the SWUS, which maintains the seesaw pattern. However, moisture convergence also influences precipitation, which may cause the precipitation in the SWUS to no longer be mainly controlled by the North American high, leading to the absence of a seesaw pattern. Previous studies (Yang et al., 2019) suggested that the Atlantic Multidecadal Oscillation (AMO) also influences the seesaw pattern, though with the opposite impact to the IPO. In our study, the AMO phase is always similar to the IPO phase, which means that the IPO dominates the formation of the seesaw pattern in our experiments, while the AMO partly decreases its impact.

Third, we offered an important way to determine whether the climate change in midlow latitudes is mainly a response to Arctic-amplified-warming (remote forcing) or to increased CO\textsubscript{2} forcing (local forcing). If the climate change in 0.1 A experiment exhibits the phases similar to those in increased CO\textsubscript{2}-induced global warming, such as weakened ocean circulation and OHT; and strengthened atmospheric circulation and AHT, the climate change is determined by both the local forcing and the remote forcing. In contrast, the climate in response to Arctic amplification in 0.1 A experiment exhibits opposite phases to those in increased CO\textsubscript{2}-induced global warming, such as the decrease in the Kuroshio current, a colder west Pacific, a drier North China, and a wetter SWUS. We suggest that the variability anomaly in the real world is mainly a response to local forcing. Furthermore, although we did not discuss it in this paper because of topic preference, but we also found that HC did not expand poleward in 0.1 A experiment, though it was strengthened due to warmer tropics, which suggests that the poleward expansion of HC is a result of local forcing.

Another issue is related to Arctic amplification formation in the 0.1 A experiment, which is briefly discussed in this study because of topic preference. We confirmed that Arctic amplification formation in different seasons is determined by sea ice loss-induced surface albedo decreases and heat storage in the subsurface ocean (Dai, 2021). However, how the heat storage charges and discharges in different seasons remain an unsolved question, which will be studied in our following work.
Appendix A: Important Formulas for Wave-Flow Interaction Discussion

Three important formulas are presented for wave-flow interaction discussion:

The dispersion relation of Rossby wave (3.18.7) in Pedlosky, 1987:

$$\sigma = \frac{k}{K^2 + F} (\bar{U}k^2 - \beta)$$

(1)

where $\bar{U}$ is the zonal flow velocity, $\sigma$ and $K$ are the wave frequency and wavenumber of Rossby waves, while $k$ is the wavenumber in the zonal direction, $F$ is the Coriolis parameter, $\beta = dF/ dy$ is the meridional gradient of the Coriolis parameter.

The envelope of disturbance $A$ (3.19.15) in Pedlosky, 1987:

$$A = B(\bar{r} - C_{r}f)$$

(2)

where $B$ is a constant, $\bar{r}$ is a vector in the $x$-$y$ plane, $C_{r} = \nabla_{x} \sigma$ is the group velocity.

The energy of Rossby wave (3.21.8a) in Pedlosky, 1987:

$$\bar{E} = \left( K^2 + F \right) \bar{A}^2 / 4$$

(3)

Data Availability Statement

All data and code for the figures in this study can be found at https://www.researchgate.net/publication/342435076_data_and_figure_for_response_to_polar_amplification?_sg=z5te8EosOIWJK_uG329-MW/jpx_dLyjy8eN9cSoSC5wB_1BervFXSy_ZgeNhDxGy9o925hHZg-FAYY8E_Mm5b9LuQO-wKDXQ4fLTb7 bdQanJUYPeuDnt1JOOM9V7taXkNI13r_mHKK8Sr7JHHN-BRNs7fyO1krgtKn0glEHlggXEgyloc2xE-D3cyIFSQ.

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