Roles of Surface Albedo, Surface Temperature and Carbon Dioxide in the Seasonal Variation of Arctic Amplification

Haijin Dai1,2

1Key Laboratory of Physical Oceanography, Ministry of Education, Ocean University of China, Qingdao, Shandong, China, 2College of Meteorology and Oceanography, National University of Defense Technology, Changsha, Hunan, China

Abstract A decrease in surface albedo over ice-covered ocean leads to global warming and Arctic-amplified warming. Numerical results indicate seasonal variation in Arctic amplification (AA) is a result of local forcing and feedbacks in the Arctic. A decrease in surface albedo leads to a positive feedback, which dominates the local forcing and feedback mechanism. Ocean heat storage in the subsurface acts as a heat forcing to delay the influence of surface albedo feedback. In summer (autumn), heat storage increases (discharges) and contributes to a negative (positive) heat forcing, which decreases (increases) the positive local forcing and feedback and triggers the occurrence of the minimum (maximum) AA. In addition, increased CO2 forcing largely decreases the outgoing longwave radiation at the surface and increases surface temperatures, especially at low latitudes and in the Arctic winters, which decreases the AA magnitude and seasonal variation, although the AA remains nearly the same during winter.

Plain Language Summary The surface albedo over ice-covered ocean is decreasing, leading to global warming and amplified Arctic warming, which confirms the importance of sea-ice loss to the Arctic amplification (AA; both formation and seasonal variation). We also study the role of increased CO2 in the seasonal variation in AA by comparing experiments with and without increased CO2, and reveal the direct contribution of CO2 forcing, which is usually neglected.

1. Introduction

In response to the radiative forcing due to anthropogenic greenhouse gas emissions, the global surface temperature (GST) increased in the last few centuries. Furthermore, the increase in the observed Arctic surface temperature increased 1.2°C per decade in the past few decades, which is twice the increase in the observed GST, and this phenomenon is called Arctic amplification (AA; Screen & Simmonds, 2010). The AA influences both weather and climate at mid-low latitudes, such as the occurrence of extremely cold events (Kug et al., 2015; Luo et al., 2019), heat waves (Francis & Vavrus, 2012), variation in westerly jets, and summer precipitation (Coumou et al., 2018; Oudar et al., 2017), although Dai (Dai & Song, 2020) argued that the AA contributes little to mid-latitude climate. Still, we may better predict future global weather and climate (in low and high latitudes) by fully understanding what causes the AA.
Geophysical Research Letters
Mautisten, 2014), vertical lapse rate variation (Goosse et al., 2018), water vapor variation, and cloud variation (Sejas et al., 2014). Increased CO₂ forcing impedes longwave radiation escaping the Earth system, enhances surface temperature and is considered the original reason for global warming (Yang et al., 2018). Surface albedo variation (mainly due to sea-ice loss) allows more solar radiation to be absorbed at the surface of the polar ocean, which in turn enhances surface temperature and decreases sea ice (positive feedback; H. Park et al., 2018). Surface warming leads to weaker negative feedback (the Planck feedback) in the Arctic than in the tropics (Pithan & Mautisten, 2014). The vertical lapse rate leads to a negative feedback in the tropics (a warmer surface increases vertical convection, which brings more heat upward, causing the upper layers [surface] to warm up [cool down]), while it often leads to a positive feedback in the polar region due to the stable stratification there (Goosse et al., 2018). Due to surface warming, water vapor increases and amplifies the greenhouse gas effect and enhances the surface warming (positive feedback), although the water vapor feedback is much smaller in the Arctic than in the tropics (H. Dai et al., 2017). Clouds influence the radiative flux in both visible (shortwave) and infrared (longwave) bands, although the feedback is uncertain due to different heights and types of clouds (Huang et al., 2017). As a result of competition among the forcing and feedbacks mentioned above, the AA occurs. To quantitatively discuss the contributions of local forcing and feedbacks to the AA, the radiative kernel method has been employed (Goosse et al., 2018, Laîné et al., 2016, ST18, Taylor et al., 2013). According to Graversen et al. (2014), the surface albedo feedback accounts for 40% of the AA and is the most important factor in AA genesis. As the key role in decreasing the Arctic surface albedo, sea-ice melting changes ice-covered ocean into ice-free ocean. If the sea ice melts only at the surface, there is a water pool at the (ice) surface, which is the so-called melt pond. In this case, although the ocean is still covered by sea ice, the surface albedo may decrease by more than 50% (Hunke et al., 2013).

Previous studies mainly discussed the influences of annual local forcing and feedbacks to the AA. However, surface albedo feedback (additional solar radiation absorption) mainly occurs in spring and summer, while the AA reaches its maximum in autumn and winter (DA19). (a) It is necessary to examine how surface albedo feedback influences the AA in different seasons and eventually dominates the annual local forcing and feedbacks. (b) Are there other factors that have nonnegligible contributions to AA formation in different seasons? For example, the energy uptake (release) by the ocean results in a minimum (maximum) in AA in summer (autumn and winter; Laîné et al., 2016). (c) Since CO₂ forcing adds to maximum polar warming in fall and winter (Sejas et al., 2014), why does the AA hardly appear without sea-ice loss in CO₂ perturbation experiment (DA19)? To address the above mentioned three issues, we organize this paper as follows. A coupled model and perturbation experiments are introduced in Section 2. We analyze the mechanisms of AA formation in different seasons using the results from sensitivity experiments in Section 3. Finally, a summary and discussion are given in Section 4.

2. Model and Experiments
The model used in this study is the Community Earth System Model, version 1.0 (CESM 1.0), of the National Center for Atmospheric Research. The same model was used in our previous studies on global energy balance (H. 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 simulations of Earth’s past, present, and future climate. The model grid employed in this study is T31_gx3v7. The atmosphere model is the Community Atmosphere Model (CAM5; S. Park et al., 2014), which has 26 vertical levels, with a finite volume of 3.75 × 3.75° in the horizontal directions. The land model is the Community Land Model (Lawrence et al., 2012), which has the same horizontal resolution as the CAM5. The ocean model is the Parallel Ocean Program, version 2 (POP2; Smith et al., 2010), which has 60 vertical levels. The horizontal grid has a uniform spacing of 3.6° in the zonal direction. In the meridional direction, the grid is nonuniformly spaced: it is 0.6° near the equator, gradually increasing to a maximum of 3.4° at 35°N/S and then decreasing poleward. The ice model is the Community Ice CodE, version 4 (Hunke & Lipscomb, 2008), which has the same horizontal grid as the POP2. More details can be found in Yang and Dai (2015) and H. Dai et al. (2017).

Experiments analyzed in this study include a 2000-year control run (CTRL), a 500-year surface albedo perturbation run, and a 500-year CO₂-surface albedo (COA) perturbation run. The CTRL starts from the rest with standard configurations (using CO₂ concentration of 285 ppm). The model climate as a whole reaches
a quasi equilibrium state after 1,000 years of integration (Yang et al., 2015). To test the direct (radiative) contribution of CO$_2$ to AA formation, we first fix the CO$_2$ concentration as CTRL value in the surface albedo perturbation experiment (0.1 A), while the surface albedo (for the shortwave radiation) over the ice-covered ocean is set to nearly the same as that over the ice-free ocean. 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 remain unchanged. The surface albedo variation initiates the global warming in 0.1 A, instead of increased CO$_2$. By decreasing the surface albedo artificially, the solar radiation absorption increases (Figure 1a, blue line), which leads to Arctic warming (Figure 1b, thick red line), sea-ice melting (Figure 1a, cyan column),

Figure 1. Climate variations between the 0.1 A (COA) experiment and CTRL in different months are shown. The shortwave radiation variation in 0.1 A (COA) experiment is shown as the blue (red) line in (a). Units: W/m$^2$. The sea-ice loss in the 0.1 A (COA) experiment is shown as the cyan (pink) column in (a). Units: 10$^5$ km$^2$. The global surface temperature variation in 0.1 A (COA) experiment is shown as a thin blue (red) line in (b), while the Arctic (66°–90°N) surface temperature variation is shown as a thick blue (red) line in (b). Units: °C. The variations in upward longwave radiation flux (thick line) and latent heat flux plus sensible heat flux (thin line) at the surface between 0.1 A experiment/COA experiment (blue/red line) and CTRL in different months are shown in (d). The AA index in 0.1 A (COA) experiment is shown as thick blue (red) line in (e), while the AA index excluding the temperature variation south of 30°S is shown as thin blue line in (e). Heat storage temperature in the Arctic ocean in the 0.1 A and COA experiments are shown in (c) and (f), respectively. Heat storage temperature ($\Delta T$) is defined as $\Delta T_{n+1} = T_{n+1} - T_{n+1}$, where $\Delta T$ is the temperature increase between the sensitivity experiment and CTRL, and $n$ is the month; if $n = 1$, $n-1 = 12$. Heat storage estimate is shown in Text S3. COA, CO$_2$-surface albedo; CTRL, control run; AA, Arctic amplification.
and surface ocean freshening (Figure S1a). As a result, it is more difficult for the surface water to sink (Figure S1b), which weakens the Atlantic meridional overturning circulation (Figure S1c, blue line) and oceanic heat transport (Figure S1d, blue line), resulting in global warming (Figure 1b thin red line). For comparison, we designed a COA perturbation experiment: we increased the CO₂ concentration by 1% yr⁻¹ for 70 years and held constant at the doubled level from the year 1571 to 2000, while we decreased the surface albedo (as in 0.1 A). Because the potential influence of shortwave absorption induced by sea ice loss are equally employed (due to artificially decreasing the surface albedo) in both 0.1 A and COA experiments, the differences between 0.1 A and COA can be considered as the direct (radiative) impact of increased CO₂ forcing. Both 0.1 A and COA are “parallel” to CTRL during the years 1501–2000, with the same initial conditions at the end of year 1500, and reach quasi equilibrium after the 500-year integration. The monthly mean outputs are used for analysis. The climate changes in 0.1 A or COA are obtained by subtracting the corresponding fields in CTRL. In this study, we focus on the equilibrium responses using the monthly averaged fields over the last 200-year integration, unless stated otherwise.

3. Formation and Variation of Seasonal Arctic Amplification

3.1. Seasonal Arctic Amplification and Heat Budget

In the 0.1 A experiment, the albedo of the ice-covered ocean is set nearly the same as the albedo of the ice-free ocean, and solar radiation absorption (net shortwave radiation) largely increases at the Arctic (66°–90°N) surface (Figure 1a, blue line), reaching 29.9 and 27.9 W/m² in spring (March-April-May) and summer (June-July-August), respectively. In response, Arctic surface temperature increases (Figure 1b, thick red line), which results in sea-ice cover in the Arctic decreasing more than 6,300,000 km² during summer (Figure 1a, cyan column). Surface temperature increase is determined by sea surface temperature (SST) variation, which is a result of heat flux variation, remote forcing (i.e., meridional heat transport [MHT]) variation, and heat storage (Figure 1c) in the subsurface ocean (H. Dai et al., 2017; Yang & Zhang, 2008); more details are given in Text S2. Since variation in MHT (sum of atmospheric meridional heat transport [AHT] and oceanic meridional heat transport [OHT]) contributes little (0.01 W/m²; Figure S1d) to the heat flux variation in the Arctic, the seasonal variation in AST is mainly a result of the local heat budget in the Arctic. Increased solar radiation absorption is dominant in spring and summer. The heat, which is stored in the subsurface ocean in summer, is released to the surface during autumn (September-October-November) and winter (December-January-February) through vertical diffusion and vertical advection. The released heat (upward longwave radiation; Figure 1d, thick blue line) reaches the maximum (65.8 W/m²) in autumn. Both sensible and latent heat flux (Figure 1d, thin lines) always transport heat from the ocean to the atmosphere due to a warmer surface in the Arctic. As a result, AST variation reaches its maximum (12°C) in autumn and minimum (2.3°C) in summer. On the other hand, since shortwave radiation absorption occurs primarily in the polar regions, the GST variation is much smaller (Figure 1b, thick red line) with small seasonal variation. Following DA19, the AA is defined as the rate between AST variation and GST variation. From the results of the 0.1 A experiment, the seasonal variation in AA is mainly determined by the AST variation (which is a result of local heat budget).

When increased CO₂ forcing is added to the decreased surface albedo experiment (COA experiment), the AST increases by as much as 3.9°C (Figure 1b, thick blue line) from the 0.1 A experiment (Figure 1b, thick red line). However, the solar radiation absorbed at the Arctic surface in COA (Figure 1a, red line) remains nearly the same as that in 0.1 A with larger sea-ice loss (Figure 1a, pink column); this is because the albedo of the ice-covered ocean is set nearly the same as the albedo of the ice-free ocean. On the other hand, the GST increases by as much as 1.8°C with increased CO₂ forcing (Figure 1b, thin blue line) because increased CO₂ leads to a positive feedback at low latitudes. The upward longwave radiation decreases due to the greenhouse effect of increased CO₂, which leads to warmer tropics. The warmer surface enhances moisture content (Figure S2a) and cloud coverage (Figure S2b), which in turn decreases the upward longwave radiation and warms the tropics further. Similar to the result in the 0.1 A experiment, there is small seasonal variation in the increase of GST in the COA experiment. The AA is still determined by local heat budget, though increased CO₂ forcing may weaken the magnitude of the AA (Figure 1e, thick red line) by inducing strong warming at lower latitudes (not shown) and increases heat storage in the subsurface Arctic ocean (Figure 1f).
Further analysis, however, suggests that the AA in autumn is much smaller than that in autumn in a world with decreased surface albedo (Figure 1e, thick blue line). This finding conflicts with the conclusion drawn by studies based on observations (DA19). The AA reaches its maximum in winter (DA19). According to D. M. Smith et al. (2019), the role of the Southern Ocean in maintaining the additional heat is underestimated in numerical models, which may result in overestimated the warming in the Southern Hemisphere. If we neglect the temperature variation south of 30°, the AA increases (Figure 1e, thin blue line) due to a smaller GST variation. However, significant differences between the AA in winter and that in autumn still exist in the 0.1 A experiment. Moreover, in the COA experiment, the AA decreases significantly except in winter (Figure 1e, thick red line). What is the role played by increased CO2 in the seasonal variation of the AA? What determines the local heat budget in the Arctic in different seasons? We answer these questions next.

### 3.2. Seasonal Arctic Amplification, Local Forcing and Feedbacks

To reveal more details about the partition of local forcing and feedbacks, and about the eventual formation of the AA, a so-called “radiative kernel technique” was employed to quantify these processes (Jonko et al., 2012; Shell et al., 2008; Soden et al., 2008; Zelinka & Hartmann, 2012). In this approach, the net change in the heat flux at the top of the atmosphere (TOA, ΔRTOA; or at the surface ΔRsurf) is the sum of external forcing (FTOA/Fsurf) and feedbacks (ΔTTOA/ΔTsurf) resulting from planetary albedo (λTOA ΔTsurf/λsurf ΔTsurf), surface temperature (ΔP Planck ΔTsurf/Δsurf ΔTsurf), vertical lapse rate (λTTOA ΔTsurf, Δsurf ΔTsurf), water vapor (ΔQ TOA ΔTsurf, Δsurf ΔTsurf), and total clouds (ΔC TOA ΔTsurf, Δsurf ΔTsurf), influenced by both low and high clouds). Latent (ΔLH) and sensible heat flux (ΔSH) should also be included at surface. These relationships are as follows.

\[
\Delta R_{TOA} = F_{TOA} + \lambda_{TOA} \Delta T_{surf} + \lambda_{P Planck} \Delta T_{surf} + \lambda_{T TOA} \Delta T_{surf} + \lambda_{C TOA} \Delta T_{surf} + \lambda_{LH} + \Delta SH \tag{1}
\]

\[
\Delta R_{surf} = F_{surf} + \lambda_{surf} \Delta T_{surf} + \lambda_{P Planck} \Delta T_{surf} + \lambda_{T surf} \Delta T_{surf} + \lambda_{C surf} \Delta T_{surf} + \lambda_{LH} + \Delta SH \tag{2}
\]

λTOA (λsurf) is the total feedback parameter at the TOA (surface), while λP Planck (λsurf), λT TOA (λsurf), λC TOA (λsurf) are the feedback parameters of albedo, surface temperature, vertical lapse rate, water vapor, and clouds at the TOA (surface), respectively. ΔTsurf is the GST variation (Huang & Zhang, 2014; Shell et al., 2008). More details can be found in Huang et al. (2017).

In the 0.1 A experiment, decreased surface albedo leads to a positive feedback in shortwave radiation (Figure 2, blue bar) of as much as 65 W/m²/K, which dominates the net flux increase at both the TOA (Figures 2a and 2c) and the surface (Figures 2b and 2d) and results in Arctic warming in spring and summer. In response, the warmer surface leads to a negative feedback (Planck feedback) in longwave radiation of approximately −7 (~2) W/m²/K and −19.5 (~8) W/m²/K at the TOA (Figure 2a [Figure 2c, red bar] and the surface (Figure 2b [Figure 2d], red bar) in spring (summer), respectively. The warmer surface also enhances vertical convection, which transports large amounts of heat and water vapor upward. The former changes the vertical lapse rate and leads to a negative feedback in longwave radiation (Figure 2, cyan bar). The latter leads to negative feedback in both shortwave and longwave radiation. In contrast, the water vapor feedback is always small over the Arctic because of the small amount of moisture in the air. Due to the increased vertical convection and water vapor, high clouds increase, which leads to a negative feedback at the TOA (Figures 2a and 2c, purple bar) and the surface (Figures 2b and 2d, purple bar). In addition, increased CO2 seems to have little influence on the net flux during spring (Figures 2a and 2b, green bar) and summer (Figures 2c and 2d, green bar). Total local forcing and feedback (Figure 2, black bar), which is mainly determined by the albedo feedback, is always positive at both the TOA and surface during spring and summer.

In autumn and winter, the Planck feedback (Figure 3, red bar) induced by the warmer surface dominates the net flux variation at both the TOA and surface. The Planck feedback reaches −29.8 (~14.7) W/m²/K and −10.4 (~5.8) W/m²/K at the surface and TOA in autumn (winter), respectively. Meanwhile, the albedo feedback, moisture feedback, lapse rate feedback, and cloud feedback are negligible (Figure 3, blue, yellow, cyan, and purple bars, respectively). Increased CO2 forcing (Figures 3c and 3d, green bar) contributes 6.2 (6.0) W/m²/K to longwave radiation at the surface during autumn (winter). Finally, the net radiative flux variation (Figure 3, black bar), which is mainly determined by the Planck feedback, is always negative at both the TOA and surface during autumn and winter.
Recall the questions raised in Section 1, “How surface albedo (already be analyzed), CO₂ forcing and energy uptake influence AA in different seasons?” In the 0.1 A experiment, although the net radiative flux variation at the surface is stronger in summer (21 W/m²/K) than in spring (12.5 W/m²/K; Figure 2, black bar), the AST variation in spring is larger than that in summer. According to analyses in Section 3.1, the net radiative flux variation is not the only source of the AST variation. There is strong heat storage in the subsurface ocean due to vertical advection and vertical diffusion forms in the summertime Arctic (Figure 1e), which is negative heat forcing (−32 W/m²/K; method is given in Text S3) to the surface. As a result, the pseudo total feedback (defined as the sum of the net radiative flux variation and heat storage) becomes

Figure 2. Variations in the radiation flux, external forcing, and feedback at the TOA/surface: in spring (a)/(b), and in summer (c)/(d). The red, yellow, cyan, blue and purple bars represent the surface temperature, water vapor, lapse rate, planetary albedo, and cloud feedbacks, respectively. The green bar presents CO₂ forcing, the black bar is the net radiative flux variation, and the orange bar is pseudo total feedback, which is the sum of net radiative flux and heat storage. The left and right columns show the results of 0.1 A experiment and COA experiment, respectively. Units: W/m²/K. TOA, top of the atmosphere; COA, CO₂-surface albedo.

Figure 3. Same as Figure 2, except for results in autumn (a)/(b) and winter (c)/(d).
negative, resulting in a minimum AA in summer. In the 0.1 A experiment, the net radiative flux variation at the surface reaches −13.1 W/m²/K in autumn, which is much stronger than that in winter (−7.4 W/m²/K). However, the heat storage discharge in the subsurface ocean is positive heat forcing to the surface, which is 21.2 W/m²/K and 7.8 W/m²/K in autumn and winter, respectively. In October and November, the positive heat forcing reaches 27.6 W/m²/K, making the pseudo total feedback (12.3 W/m²/K) even larger than that in April (4.3 W/m²/K). As a result, the AST variation in the 0.1 A experiment reaches a maximum in autumn, has a secondary peak in spring, and attains a minimum in summer. With increased CO₂ forcing (COA experiment), the surface temperature increases, especially at low latitudes. Local forcing and feedbacks decrease due to a larger GST variation in spring and summer. The Planck feedback to longwave radiation at the surface remains dominant with respect to local forcing and feedbacks in autumn, although it decreases by 30%. On the contrary, increased CO₂ forcing appears to have little influence on the Planck feedback in winter. As a result, the total local forcing and feedback decreases to −5.7 (−4.9) W/m²/K, while the pseudo total feedback decreases (increases) to 6.0 (1.6) W/m²/K in autumn (winter). The difference in the pseudo total feedback between autumn and winter decreases from 8.0 W/m²/K to 4.0 W/m²/K. As the CO₂ concentration continues to increase, the Planck feedback in autumn decreases further due to a larger global warming. The pseudo total feedback may continue to decrease in autumn, although remaining positive, while it increases in winter due to more heat storage discharge and stronger increased CO₂ forcing. When the pseudo total feedback in autumn becomes similar to or even smaller than that in winter, the AA in winter is comparable to that in autumn and is consistent with recent observations.

4. Conclusion and Discussion

To understand how surface albedo, surface temperature, and CO₂ influence the seasonal variation in AA, we use a coupled climate model with simulations that elucidate these forcing effects. The results show that the seasonal variation in AA is mainly a result of local forcing and feedbacks in the Arctic. A decrease in surface albedo can largely enhance solar radiation absorption in the Arctic. This phenomenon should be viewed as the real heat source for amplified Arctic warming. Moreover, CO₂ prevents heat from escaping the Earth and enhances surface temperatures, especially at low latitudes. The Planck feedback is a good indicator for the magnitude of AA.

In this study, we discuss the seasonal variation in AA to a greater extent in several ways.

First, although polar warming and sea-ice loss in recent centuries were initiated by increased CO₂ forcing, polar ocean warmed up under many conditions (i.e., the Milankovitch theory, Berger, 1988; He et al., 2013; Sigman & Boyle, 2000). So, we discuss the influence of surface albedo and increased CO₂ individually in this study. To separate the influence of surface albedo from increased CO₂ forcing, global warming is produced by artificially decreasing the albedo of the ice-covered ocean. In this way, both surface albedo and solar radiation absorption over the polar ocean remain nearly unchanged even with larger sea-ice loss. Thus, when increased CO₂ forcing is applied in the surface albedo perturbation experiment, new variation should be contribution from the direct (radiative) impact of CO₂ forcing. As a result, we separate the influences of surface albedo and CO₂ forcing to AA successfully.

Second, all the sensitivity experiments in this study are highly idealized, which induce stronger polar warming and sea ice loss than those in the recent observations. Although the Arctic surface albedo can hardly reach such a small magnitude in the real world and we only employ the CESM in this study, many features in our polar forcing experiment (0.1 A) are similar as those in the polar forcing experiment carried out using other models (ST18). (1) AHT (OHT) is northward (southward) anomaly south (north) of 40°N (equator) in 0.1 A (Figure S1d), which shares similar distribution as that in Figure 3 of ST18. (2) Surface temperature increase is maximum in the Arctic, with the second peak in the Antarctic, and the minimum is in the subtropic region (Figure 4 in ST18; our result is not shown though). All the results above confirm the robustness of the model feature-AA is dominant by local forcing and feedbacks.

Third, the positive feedback of surface albedo dominates the local forcing and feedbacks throughout the year (at the surface and TOA), and the heat storage in the subsurface ocean plays a role in delaying the influence of this positive feedback. Subsurface ocean heat storage forms in summer and discharges in autumn and winter. Thus, the pseudo total feedback, which is the sum of the heat storage and net radiative flux
variation, can well explain the seasonal AST variation and AA, which reaches the maximum in autumn, has a secondary peak in spring, and attains the minimum in summer. The direct (radiative) impact of CO2 forcing enhances the GST variation and decreases the local forcing and feedbacks in spring, summer, and autumn. However, with more heat storage discharged from the subsurface ocean (due to a warming climate) and stronger CO2 forcing, a larger pseudo total feedback (positive) is achieved in winter.

When we study the climate change in recent centuries, the surface albedo feedback becomes an indirect influence of increased CO2 forcing. As a result, two opposite impacts of increased CO2 forcing on AA occur: the direct (indirect) impact of increased CO2 forcing reduces (increases) the difference between GST and AST. The result in Figures S4 and S5 suggests that the indirect impact dominates the local forcing and feedbacks in the Arctic. Another key factor for the AA is the heat storage in the subsurface ocean, and how heat storage forms/discharges (through vertical advection and vertical diffusion) in different seasons will be reported in future.

Data Availability Statement
All the data and codes for making the figures in this paper can be found at https://www.researchgate.net/publication/347667326_figure_for_Roles_of_Surface_Albedo_Surface_Temperature_and_Carbon_Dioxide_in_the_Seasonal_Variation_of_Arctic_Amplification.

References
Berger, A. (1988). Milankovitch theory and climate. Reviews of Geophysics, 26(4), 624–657. https://doi.org/10.1029/RG026i004p00624
Couchou, D., Di Capua, G., Vavrus, S., Wang, L., & Wang, S. (2018). The influence of arctic amplification on mid-latitude summer circulation. Nature Communications, 9(1), 2595. https://doi.org/10.1038/s41467-018-05256-8
Dai, A., Luo, D., Song, M., & Liu, J. (2019). Arctic amplification is caused by sea-ice loss under increasing CO2. Nature Communications, 10(1), 121. https://doi.org/10.1038/s41467-018-07954-9
Dai, A., & Song, M. (2020). Little influence of Arctic amplification on mid-latitude climate. Nature Climate Change, 10(3), 231–237. http://doi.org/10.1038/s41558-020-0694-3
Huang, Y., & Tan, X. (2017). On the pattern of CO2 radiative forcing and poleward energy transport. Reviews of Geophysics, 55(6), 2015–2050. https://doi.org/10.1002/2015RG000521
Huang, Y., & Zhang, M. (2014). The implication of radiative forcing and feedback for meridional energy transport. Geophysical Research Letters, 41(5), 1665–1672. https://doi.org/10.1002/2013GL059079
Hunke, E. C., Heber, D. A., & Leconte, O. (2013). Level ice melt ponds in the Los Alamos sea ice model, CICE. Ocean Modelling, 71, 26–42. https://doi.org/10.1016/j.ocemod.2012.11.008
Hunke, E. C., & Lipscomb, W. H. (2008). CICE: The Los Alamos Sea Ice Model, documentation and software user’s manual, version 4.0 (Tech. Rep. LA-CC-06-012). Los Alamos, Los Alamos National Laboratory.
Jonko, A. K., Shell, K. M., Sanderson, B. M., & Danabasoglu, G. (2012). Climate feedbacks in CCSM4 under changing CO2 forcing, Part I: Adapting the linear radiative kernel technique to feedback calculations for a broad range of forcings. Journal of Climate, 25(15), 5260–5272. http://doi.org/10.1175/JCLI-D-12-00479.1
Kug, J. S., Jeong, J. H., & Jang, Y. S. (2015). Two distinct influences of Arctic warming on cold winters over North America and East Asia. Nature Geoscience, 8(10), 759–762. https://doi.org/10.1038/NGEO2517
Laïné, A., Yoshimori, M., & Abe-Ouchi, A. (2016). Surface Arctic amplification factors in CMIP5 models: Land and oceanic surfaces and seasonality. Journal of Climate, 29(9), 3297–3316. https://doi.org/10.1175/JCLI-D-15-0497.1
Lawrence, D. M., Oleson, K. W., Flanner, M. G., Fletcher, C. G., Lawrence, P. J., Levis, S., et al. (2012). The CCSM4 land simulation, 1850–2005: Assessment of surface climate and new capabilities. Journal of Climate, 25(7), 2240–2266. https://doi.org/10.1175/JCLI-D-11-00103.1
Luo, D., Chen, X., Overland, J. E., Simmonds, I., Wu, Y., & Zhang, P. (2019). Weakened potential vorticity barrier linked to recent warmer Arctic sea ice loss and midlatitude cold extremes. Journal of Climate, 32(14), 4235–4261. https://doi.org/10.1175/JCLI-D-18-0449.1
Oudar, T., Sanchez-Gomez, E., Chauvin, F., Cattiaux, J., Terray, L., & Cassou, C. (2017). Respective roles of direct GHG radiative forcing and induced Arctic sea ice loss on the Northern Hemisphere atmospheric circulation. Climate Dynamics, 49(11/12), 3693–3713. https://doi.org/10.1007/s00382-017-3541-0
Park, S., Bretherton, C. S., & Rasch, P. J. (2014). Integrating cloud processes in the community atmosphere model, version 5. Journal of Climate, 27(18), 6821–6856. https://doi.org/10.1175/JCLI-D-14-00087.1
Park, H., Kim, S., Seo, K., Stewart, A. L., Kim, S., & Son, S. (2018). The impact of Arctic sea ice loss on mid-Holocene climate. *Nature Communications*, 9(1), 4571. https://doi.org/10.1038/s41467-018-07068-2

Pithan, F., & Mauritsen, T. (2014). Arctic amplification dominated by temperature feedbacks in contemporary climate models. *Nature Geoscience*, 7(3), 181–184. https://doi.org/10.1038/ngeo2071

Screen, J. A., & Simmonds, I. (2010). The central role of diminishing sea ice in recent Arctic temperature amplification. *Nature*, 464(7293), 1334–1337. https://doi.org/10.1038/nature09051

Sejas, S. A., Cai, M., Hu, A., Meehl, G. A., Washington, W., & Taylor, P. C. (2014). Individual feedback contributions to the seasonality of surface warming. *Journal of Climate*, 27(14), 5653–5669. https://doi.org/10.1175/JCLI-D-13-00658.1

Shell, K. M., Kiehl, J. T., & Shields, C. A. (2008). Using the radiative kernel technique to calculate climate feedbacks in NCAR's Community Atmospheric Model. *Journal of Climate*, 21(10), 2269–2282. https://doi.org/10.1175/2007jcli2044.1

Sigman, D. M., & Boyle, E. A. (2000). Glacial/interglacial variations in atmospheric carbon dioxide. *Nature*, 407(6806), 859–869. https://doi.org/10.1038/35038000

Smith, R. D., & Gent, P. (2010). *The Parallel Ocean Program (POP) reference manual* (Tech. Rep. LAUR-10-01853). Los Alamos, Los Alamos National Laboratory.

Smith, D. M., Screen, J. A., Deser, C., Cohen, J., Fyfe, J. C., Garcia-Serrano, J., et al. (2019). The polar amplification model intercomparison project (PAMIP) contribution to CMIP6: Investigating the causes and consequences of polar amplification. *Geoscientific Model Development Discussions*, 12(3), 1139–1164. https://doi.org/10.5194/gmd-2018-8

Soden, B. J., Held, I. M., Colman, R., Shell, K. M., Kiehl, J. T., & Shields, C. A. (2008). Quantifying climate feedbacks using radiative kernels. *Journal of Climate*, 21(14), 3504–3520. https://doi.org/10.1175/2007JCLI2110.1

Stuecker, M. F., Bitz, C. M., Armour, K. C., Proistosescu, C., Kang, S. M., Xie, S. P., et al. (2018). Polar amplification dominated by local forcing and feedbacks. *Nature Climate Change*, 8(12), 1076–1081. https://doi.org/10.1038/s41558-018-0339-y

Taylor, P. C., Cai, M., Hu, A., Meehl, J., Wahshington, W., & Zhang, G. J. (2013). A decomposition of feedback contributions to polar warming amplification. *Journal of Climate*, 26(18), 7023–7043. https://doi.org/10.1175/JCLI-D-12-00696.1

Yang, H., & Dai, H. (2015). Effect of wind forcing on the meridional heat transport in a coupled model: Equilibrium response. *Climate Dynamics*, 45(5), 1451–1470. https://doi.org/10.1007/s00382-014-2393-0

Yang, H., Li, Q., Wang, K., Sun, Y., & Sun, D. (2015). Decomposing the Meridional Heat Transport in the Climate System. *Climate Dynamics*, 44(9–10), 2751–2768. https://doi.org/10.1007/s00382-014-2380-5

Yang, H., & Zhang, Q. (2008). Anatomizing the ocean’s role in ENSO changes under global warming. *Journal of Climate*, 21(24), 6539–6555. https://doi.org/10.1175/2008JCLI2324.1

Yang, Q., Zhao, Y., Wen, Q., Yao, J., & Yang, H. (2018). Understanding Bjerknes compensation in meridional heat transports and the role of freshwater in a warming climate. *Journal of Climate*, 31(12), 4791–4806. https://doi.org/10.1175/jclim-d-17-0587.1

Zelinka, M. D., & Hartmann, D. L. (2012). Climate feedbacks and their implications for poleward energy flux changes in a warming climate. *Journal of Climate*, 25(2), 608–624. https://doi.org/10.1175/JCLI-D-11-00096.1

Zheng, J., Zhang, Q., Li, Q., Zhang, Q., & Cai, M. (2019). Contribution of sea ice albedo and insulation effects to Arctic amplification in the EC-Earth Pliocene simulation. *Climate of the Past*, 15(1), 291–305. https://doi.org/10.5194/cp-15-291-2019