Understanding the Cold Season Arctic Surface Warming Trend in Recent Decades

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Abstract Whether sea-ice loss or lapse-rate feedback dominates the Arctic amplification (AA) remains an open question. Analysis of data sets based upon observations reveals a 1.11 K per decade surface warming trend in the Arctic (70°–90°N) during 1979–2020 cold season (October–February) that is five times higher than the corresponding global mean. Based on surface energy budget analysis, we show that the largest contribution (∼82%) to this cold season warming trend is attributed to changes in clear-sky downward longwave radiation. In contrast to that in Arctic summer and over tropics, a reduction in lower-tropospheric stability explains the reduction of the downward longwave radiation associated with atmospheric nonuniform temperature and corresponding moisture changes. Our analyses also suggest that Arctic lower-tropospheric stability should be considered in conjunction with sea-ice decline during the preceding warm season to explain AA.

Plain Language Summary Observations and climate models have consistently shown a stronger surface warming in the Arctic than the global mean, a.k.a. Arctic amplification (AA), which has a strong asymmetry between the cold season and warm season. Previous studies suggested that key contributors to AA are the positive surface-albedo feedback and lapse-rate feedback. However, the lapse-rate feedback itself depends on temperature profiles and sea-ice loss. Whether sea-ice loss or lapse-rate feedback dominates AA remains an open question. Here, by analyzing the latest generation of observationally-based reanalysis data (1979–2020), we present a unique role of lower-tropospheric temperature inversion changes in representing the contribution of vertically inhomogeneous atmospheric temperature and associated moisture changes to clear-sky downward longwave radiation during the cold season. This unique role is not found either in the tropics or during Arctic summertime. We further link the inversion during the cold season to sea-ice loss during the preceding warm season. These results reinforce previous findings not only that lapse-rate feedback and sea-ice loss play a key role in AA but also that lapse-rate feedback in the cold season is likely a consequence of sea-ice albedo feedback during the preceding warm season.

1. Introduction

Arctic, the northern polar region of Earth, is especially sensitive to climate change and now experiencing an accelerated warming that is much faster than the global warming trend (Ballinger et al., 2020; Cohen et al., 2020; Post et al., 2019; Wuebbles et al., 2017). Since a pioneering study in 1980 (Manabe & Stouffer, 1980) popularized this phenomenon by naming it Arctic amplification (AA), many studies have investigated the mechanisms driving AA from both local and remote perspectives (Cohen et al., 2018; Screen et al., 2012; Stuecker et al., 2018). A local perspective invoking surface albedo feedback (Screen & Simmonds, 2010a; Taylor et al., 2013), lapse rate feedback (Pithan & Mauritsen, 2014; R. D. Zhang et al., 2018), radiative forcing due to greenhouse gases (GHGs) and cloud cover changes (Gong et al., 2017; Kay & Gettelman, 2009; Liu & Key, 2014; Vavrus, 2004) to explain AA. Other studies use a remote perspective involving changes in the atmosphere and ocean heat transports from extra-polar latitudes to the Arctic (Beer et al., 2020; Graversen et al., 2008; Holland & Bitz, 2003; Hwang et al., 2011; Park et al., 2015; Yoshimori et al., 2017) to explore AA.

It is well known that the positive surface albedo feedback (i.e., temperature increase → snow and ice retreat → surface albedo decrease → solar absorption increase → more temperature increase) plays an important...
role in AA (Hall, 2004; R. D. Zhang et al., 2019), but AA can also be found in idealized climate model experiments without sea ice-albedo feedback (Alexeev et al., 2005; Kim et al., 2018). This indicates that other processes involve modulation of the outgoing longwave radiation (OLR) emitted to space at the top of the atmosphere (TOA) in response to surface or tropospheric changes (Pithan & Mauritsen, 2014). This negative temperature feedback can be artificially decomposed into a Planck feedback (assuming the warming is vertically uniform) and a lapse rate feedback (describing the changes in OLR associated with deviations from a vertically uniform warming; Goosse et al., 2018). In contrast to the top-heavy warming structure found in the tropics, warming of the surface and lower troposphere is larger than that of the middle and upper troposphere at high latitudes (Manabe & Wetherald, 1975). This bottom-heavy warming structure is dictated by the surface temperature change and results in a positive lapse rate feedback on surface temperature (Boeke et al., 2020). Based on the TOA energy budget analysis, Pithan and Mauritsen (2014) concluded that the local positive lapse rate feedback is the largest contributor to AA relative to the tropics. Unlike the tropics that are approximately in radiative-convective equilibrium, however, the troposphere of high latitudes is closer to a radiative-advective equilibrium with poleward heat transport being balanced by radiative cooling (Cronin & Jansen, 2016). This suggests that the lapse rate feedback over the Arctic may not be entirely explained by a single local process affecting stable stratification but instead may result from both local and remote processes including advection that are responsible for a vertically nonuniform warming (Boeke et al., 2020). By decomposing the high-latitude lapse rate feedback into upper and lower components, Feldl et al. (2020) show that local sea ice loss is the driver of the lower troposphere lapse rate feedback, while the upper component is driven by poleward atmospheric energy transport. As the planet warms globally, the poleward atmosphere and ocean heat transport to the Arctic both increase and this can also contribute to AA (Bitz et al., 2006; Cai, 2005; Hwang & Frierson, 2010; Nummelin et al., 2017; Singh et al., 2017; X. D. Zhang et al., 2013). Thus, it remains unclear what is the main driving mechanism of AA due to the complexity of the underlying processes.

To better understand the Arctic surface warming trend in recent decades, we here present results from the fifth generation of European Center for Medium-Range Weather Forecasts (ECMWF) observationally based reanalysis, ERA5 (see the Section 2; Hersbach et al., 2020). ERA5 with enhanced temporal and spatial resolution has been substantially improved in representing temperature, humidity, and winds in the Arctic (Graham, Hudson, et al., 2019), especially showing an excellent performance of the Arctic energy budget closure (Mayer et al., 2019). We evaluate the trends of surface air temperature, OLR, and column-integrated atmospheric water vapor over the Arctic from ERA5 against observations (Figures S1–S3), which further strengthens confidence in using ERA5 (see Supporting Information S1 for more details). Based on ERA5 surface temperature (1979–2020), we calculate AA as a function of season and latitude. As shown in Figure 1a, although AA is a distinct feature in almost all months over high northern latitudes during the past four decades, the strong AA mostly occurred in the boreal cold season (October to following February, hereafter ONDJF) and north of 70°N. The strongest warming in the latitudinal band of 70°–83°N is intimately associated with sea ice loss (Screen & Simmonds, 2010b). The trend of area-averaged surface skin temperature over the Arctic (70°–90°N) is more than five times higher than the global mean during ONDJF from 1979 to 2020 (Figure 1b). To better understand this dramatic trend, we perform an Arctic surface energy budget analysis to decompose the surface temperature trend.
2. Materials and Methods

ERA5 is the fifth-generation ECMWF atmospheric reanalysis of the global weather and climate (Hersbach et al., 2020). ERA5 is based on the new data assimilation system, ECMWF Integrated Forecasting System Cy41r2, which has been in operation since 2016 and includes an improved four-dimensional variational scheme (4D-Var). Various observational data sets, such as weather station, buoy, radiosonde, and satellite data, have been used in ERA5. Relative to the fourth-generation ECMWF reanalysis (ERA-Interim; Dee et al., 2011), ERA5 has a higher spatial resolution (~31 km horizontal resolution and 137 model levels from surface to 0.01 hPa) and a higher output frequency (hourly analysis fields). Here, we use the monthly averaged ERA5 gridded data from January 1979 to February 2021, which are provided by the Copernicus Climate Change Service (C3S) at 0.25° × 0.25° (longitude × latitude) horizontal resolution and 37 pressure levels.

The contribution of temperature (T) and moisture (Q) to clear-sky downward longwave radiation at the surface is calculated using the surface radiative-kernel method that is described by equation:

\[ \Delta R_s = K_x \Delta x, \]  

where \( \Delta x \) denotes the monthly anomalies of climate variable \( x \) (T or Q) relative to its monthly climatological mean at each grid and level, \( K_x \) is the monthly clear-sky longwave radiative kernel at the surface, and \( \Delta R_s \) is the associated clear-sky surface radiative flux anomalies. The radiative kernels used in this study are from Huang et al. (2017), which have a horizontal resolution of 2.5° × 2.5° and 24 vertical pressure levels. Given the coarser resolution of kernels, we first regrid ERA5 data to the same resolution as kernels and then calculate the mass-weighted vertical integral of \( \Delta K_x \). Unlike the Q kernel with 24 vertical levels, the T kernel has 25 vertical levels, including an additional surface layer that also contributes to surface downward longwave radiation.

The trend uncertainty (i.e., 95% confidence interval of the linear regression slope) and the statistical significance consider the effective degree of freedom to account for the lag-1 autocorrelation coefficient of the time series based on the effective sample size (Bretherton et al., 1999; Santer et al., 2000).

3. Results

3.1. Decomposing Surface Warming Trend

The first goal is to quantify the contribution of each energy source and sink to the trend of Arctic surface warming in the cold season (shown in Figure 1b). To accomplish this, a surface energy budget equation (Hartmann, 2016) is expressed as:

\[ G = \text{rsns} + \text{rlds} - \text{rlus} - \text{SH} - \text{LE}, \]  

where rsns is the net absorption of shortwave radiation, rlds is the downward longwave radiation, rlus is the upward longwave radiation, SH is the upward sensible heat flux, LE is the upward latent heat flux, and \( G \) is the net gain of heat fluxes underneath the surface through energy transport (i.e., ocean heat transport). We calculate \( G \) term at each grid for each month based on Equation 2. The rlds can be further split into corresponding clear-sky component (hereafter rldsclr) and cloud radiative effect (hereafter rlscld; Ramanathan et al., 1989). The hypothetical clear-sky radiation in ERA5 is diagnosed for exactly the same atmospheric conditions but with cloud removed from the radiative transfer calculations (Hersbach et al., 2020). The rlus can be expressed as \( \text{rlus} = \varepsilon \sigma T_s^4 \), where \( \varepsilon \) is the surface longwave emissivity, \( \sigma \) is the Stefan-Boltzmann constant, and \( T_s \) is the surface skin temperature. By assuming that \( \varepsilon = 1 \) (Clark et al., 2021), we can decompose the Ts trend into six terms by the following equation:

\[ \frac{\Delta T_s}{\Delta t} = \frac{\Delta}{\Delta t} \left[ \text{rsns} + \text{rldsclr} + \text{rlscld} - \text{SH} - \text{LE} - G \right] + \gamma, \]  

where the operator \( \Delta/\Delta t \) represents the linear trend of time series, \( T_s \) is its monthly mean climatology of 1979–2020, and \( \gamma \) denotes a residual term.

Before discussing the results based on Equation 3, it is noted that the contributions from individual terms on the right-hand side of Equation 3 to the Ts trend on the left-hand side of Equation 3 do not necessarily
indicate a causal relationship. Vargas Zeppetello et al. (2019) noted that rlds and Ts changes are intimately connected because rlds is largely determined by temperatures in the boundary layer for given relative humidity, and those temperatures are tightly coupled to surface temperature through boundary layer turbulence. The trend of area-weighted average of Ts over the Arctic during ONDJF season is $1.11 \pm 0.35$ K decade$^{-1}$ (Figure 1b). The trends of individual terms in Equation 3 are plotted in Figure 2b (see Figure S4 for more details). While the trend of solar radiation (rsns) term is small ($0.02 \pm 0.01$ K decade$^{-1}$) due to the polar night in the cold season, the rlds has a dominant contribution to the Arctic surface warming trend. The trend of clear-sky downward longwave radiation (rldsclr) term and the surface longwave cloud forcing (rldscld) term are $0.91 \pm 0.25$ and $0.23 \pm 0.1$ K decade$^{-1}$, respectively. The latter is consistent with the increasing total cloud water path (Figure S5a). The total cloud water path dominated by the liquid cloud water path has a high correlation coefficient ($R = 0.89$) with the rldscld term in Equation 3 (Figure S5b). This is consistent with the increase of poleward moisture transport and a warmer and wetter Arctic revealed by reanalysis and satellite observations (Boisvert & Stroeve, 2015; X. D. Zhang et al., 2013). The negative climatological mean value of downward positive G (Figure 2a) indicates that G is an energy source to the surface in the cold season. The trend in G has a positive contribution to the surface warming (Figure 2b). This is usually explained in terms of seasonally delayed warming mechanism: ocean takes up more heat in summer and then releases more heat in fall and winter (Screen & Simmonds, 2010b; Serreze et al., 2007). The trend of G term is $0.19 \pm 0.17$ K decade$^{-1}$. Because of the much larger area of ocean than land over the Arctic ($70^\circ$–$90^\circ$N) and much smaller heat capacity of land than ocean, the G term in the cold season mainly results from ocean heat storage of solar energy during the preceding warm season and ocean heat transport from extra-polar regions. As the response terms to surface warming, the upward SH and LE fluxes are increasing and thus their trends are $-0.06 \pm 0.07$ and $-0.21 \pm 0.06$ K dec$^{-1}$, respectively. Note that the residual term in Equation 3 is extremely small (as indicated by the sum of explained Ts trends of $1.09 \pm 0.34$ in Figure 2b vs. $1.11 \pm 0.35$ in Figure 1b).

3.2. Attribution of Clear-Sky Downward Longwave Radiation Changes

The clear-sky surface downward longwave radiation (rldsclr) depends on the vertical profiles of temperature, water vapor, and other GHGs in the atmosphere (Fu et al., 1997). Recently, the radiative kernel technique has been widely used to quantify climate feedbacks (Soden et al., 2008; Zelinka et al., 2020; R. D. Zhang et al., 2020). The temperature and moisture kernels represent the response of surface or TOA radiation to, respectively, a 1 K warming at each vertical level including the surface and increment in water vapor caused by 1 K warming with the relative humidity unchanged (Huang et al., 2017). Here, we attribute the rldsclr trend to changes in temperature and moisture using the surface radiative kernels (see the Section 2). The trend of rldsclr over the Arctic during ONDJF season is $3.48$ W m$^{-2}$ decade$^{-1}$ (red line in Figure 3a). The contributions of changing atmospheric temperature and moisture to the rldsclr trend are, respectively, $2.24$ (sum of trends in Figures S6b and S6d) and $0.98$ (sum of trends in Figures S6c and S6e) W m$^{-2}$ decade$^{-1}$. The remaining $0.26$ W m$^{-2}$ decade$^{-1}$ in the rldsclr trend unexplained by temperature and moisture changes is largely caused by the increase in CO$_2$ concentrations. Based on observations at the Arctic North Slope of Alaska station operated by the US Department of Energy Atmospheric Radiation Measurement program (Stokes &
Figure 3. Attribution of clear-sky surface downward longwave radiation (rldsclr). Area-weighted average of ONDJF-mean anomalies (relative to 1979–2020) over the Arctic in (a) rldsclr (red circle) and its attribution to (b) Combination of atmospheric uniform temperature and corresponding moisture changes (black circle) and (c) Combination of atmospheric nonuniform temperature and corresponding moisture changes (black circle). Each black circle in panel (a) represents the sum of corresponding black circles in panels (b and c) for each year. Orange circles in panel (b) are based on the linear least squares fitting of black circles onto the lower-tropospheric inversion, defined as the difference between atmospheric temperature at 850 hPa and surface skin temperature (hereafter T850 – Ts). Green circles in panel (c) are based on the linear least squares fitting of black circles onto lower-tropospheric temperature inversion (difference between air temperature at 850 hPa and Ts). $R^2$ is the coefficient of determination of fitting using data with trends (and without trends in parentheses). In all panels, lines represent the corresponding trend of circles based on the linear least squares regression. All trends are statistically significant at the 99% confidence level. Units are in W m$^{-2}$ for all panels.

Schwartz, 1994), Feldman et al. (2015) found that the least squares trend of 2000–2010 full time series in clear-sky CO$_2$ surface radiative forcing is 0.2 W m$^{-2}$ decade$^{-1}$, with a seasonal range of 0.1 W m$^{-2}$.

Following Vargas Zeppetello et al. (2019), we can further decompose the atmospheric temperature contribution into the contribution of vertically uniform warming matching the surface temperature change (2.7 W m$^{-2}$ decade$^{-1}$, Figure 3b) and the deviation from the vertically uniform profile (−0.46 W m$^{-2}$ decade$^{-1}$, Figure 3d). We also do a similar decomposition of the atmospheric moisture contribution into that due to vertically uniform warming with unchanged relative humidity profiles (1.56 W m$^{-2}$ decade$^{-1}$, Figure 3c) and that largely related to the nonuniform temperature change with unchanged relative humidity profiles (i.e., lapse rate-induced water vapor contribution to rldsclr, −0.58 W m$^{-2}$ decade$^{-2}$, Figure 3e). Therefore, the combination of vertically uniform atmospheric temperature and related moisture changes results in 4.25 W m$^{-2}$ decade$^{-1}$ (black line in Figure 3b), which is very close to the trend of surface upward longwave radiation (4.18 W m$^{-2}$ decade$^{-2}$) and can be fully reconstructed by surface temperature changes (orange line in Figure 3b).

While we cannot attribute a cause-effect relationship between the increase in downward longwave radiation and surface warming in the Arctic cold season as in previous studies (Boeke & Taylor, 2018; Gong et al., 2017; Lee et al., 2017; Lu & Cai, 2009) due to the tight coupling between them (Vargas Zeppetello et al., 2019), we are interested in exploring mechanisms that lead to −30% (−1.04 W m$^{-2}$ decade$^{-1}$, black line in Figure 3c) contribution of nonuniform changes in atmospheric temperature and associated moisture to the rldsclr trend. This negative contribution indicates that less downward longwave radiation is emitted to the surface relative to the assumption of vertically uniform temperature change profile as the surface warming. In the positive lapse rate feedback identified over the Arctic from a TOA perspective (Pithan & Mauritsen, 2014), the OLR to space decreases relative to a vertically uniform warming profile. Thus, the nonuniform atmospheric profiles in the Arctic play the same role from both the TOA- and surface-based perspective by emitting less longwave radiation from the atmosphere to space and to surface. However, we cannot derive the feedback processes based on the surface energy budget where the surface temperature is not related to the radiative forcing. Another difference is that the nonuniform temperature changes increase the radiative energy in the earth-atmosphere system from the TOA-based perspective but decrease it at the surface from the surface-based perspective.

The lower-tropospheric stability determines the sign (positive or negative) of the TOA-based lapse rate feedback (Boeke et al., 2020). The TOA-based lower troposphere lapse rate feedback has a good correlation with temperature inversion across climate models over the Arctic (Feldl et al., 2020). In line with this inspiration, we regress the contribution of nonuniform changes in atmospheric temperature and associated moisture onto the lower-tropospheric inversion, defined as the difference between atmospheric temperature at 850 hPa and surface skin temperature (hereafter T850 – Ts). The trend of regression-based time series is −0.94 W m$^{-2}$ decade$^{-1}$ (green line in Figure 3c). The coefficient of determination ($R^2$), which measures the proportion of variance explained by the predictors, is 0.96 based on time series with trends. Given cautions regarding the interpretation of high correlation between two variables with significant trends, we further
calculate $R^2$ using detrended data ($R^2 = 0.93$ in parentheses of Figure 3c). To confirm this special feature over the Arctic during the cold season, we perform the same analysis over the tropics (30°S–30°N) during the same ONDJF season (Figure S7) and over the Arctic but for the summertime (June–August, Figure S8). The contribution of nonuniform changes in atmospheric temperature and associated moisture is only $-10\%$ and $-6\%$ to rldscrl trends in the Arctic summer and over tropics, respectively. Furthermore, there is no good correlation between $T_{850}$ and $T_s$ and the contribution of nonuniform atmospheric profiles, as shown in Figures S7c and S8c.

3.3. Diagnosis of Lower-Tropospheric Temperature Inversion Changes

Lower-tropospheric temperature inversion has long been recognized as a pervasive feature of Arctic climate, especially over sea ice in winter (Medeiros et al., 2011; Y. Zhang et al., 2011). The Arctic bottom-heavy warming structure leads to a less OLR increase, relative to a uniform warming, by emitting less radiation from the atmosphere to space, which results in additional surface warming as compared to a uniform warming (Pithan & Mauritsen, 2013). Generally, surface cooling and warm-air advection over a cooler surface are considered as two mechanisms to form lower-tropospheric inversions (Bradley et al., 1992). Surface upward latent heat fluxes (LE) deliver energy from the surface and then release that latent heat to directly warm the atmosphere through condensation. Here we link $T_{850}$ and $T_s$ to LE over the Arctic in the same ONDJF season using the regression analysis with and without trend (Figures 4a and 4d). The surface upward sensible heat flux (SH) shows a weak trend and a weak correlation with $T_{850}$ and $T_s$ (Figure S9). There are strong correlations between $T_{850}$ and LE ($R = -0.89$ before detrending and $R = -0.76$ after detrending). The trend of regression-based time series (black line in Figure 4a) can explain 96% of the total trend in $T_{850}$ and $T_s$ (green line in Figure 4a). The standard deviation of regression-based detrended time series (black line in Figure 4d) can explain 76% of the interannual variability of detrended $T_{850}$ and $T_s$ (green line in Figure 4d).

Arctic sea ice cover is an important indicator of climate change and its reduction has been proposed to play a key role in Arctic warming (Dai et al., 2019; Deser et al., 2010; Kumar et al., 2010; Screen et al., 2012; Screen & Simmonds, 2011a). As a barrier between cold atmosphere and warm ocean in the cold season, sea ice plays an insulating role to directly prevent heat release from the surface to the lower troposphere. Arctic sea ice decline exposes more sea surface to air and allows more heat release, which is referred to as the insolation mechanism to explain the role of sea ice loss (Screen & Simmonds, 2010b). Thus, we further link LE to sea ice concentration (hereafter SIC) over the Arctic in the same ONDJF season. The correlation coefficient $R$ between LE and SIC is $-0.88$ before detrending (Figure 4b) and $-0.65$ after detrending (Figure 4e). The trend of regression-based time series (black line in Figure 4b) can explain 98% of the total trend in LE (red line in Figure 4b). As has been discussed previously, another mechanism of sea ice is the delayed seasonal effect on the sea ice-albedo feedback (solar energy storage in summer and subsequent heat release in winter through the seasonal ocean memory). Our results do show good correlations ($R = 0.95$) between SIC during ONDJF season and SIC during the preceding warm season (March to September, hereafter MAMJJAS) in Figures 4c and 4f. These correlation and regression analyses help us to understand the relationship between sea-ice albedo feedback in the warm season and Arctic warming in the subsequent cold season through local coupled atmosphere-ice-ocean interactions (Boeke & Taylor, 2018; Burt et al., 2016; Feldl et al., 2020; Stuecker et al., 2018).

4. Summary

We use a surface energy budget analysis based on the fifth-generation ECMWF reanalysis (ERA5) to quantify the contribution of various polar mechanisms controlling energy sources and sinks over the Arctic (70°–90°N). These processes lead to a 1.11 K per decade surface warming trend during 1979–2020 cold season (October–February) that is over five times faster than the corresponding global warming trend. Quantitatively, the largest contributor to the Arctic surface warming trend appears connected to a trend in clear-sky downward longwave radiation at the surface that is sensitive to the lower-tropospheric temperature inversion. Trends in inversion strength influence the downward flux of longwave radiation through the vertically inhomogeneous atmospheric temperature and associated moisture changes. These connections are not found in the tropics or during Arctic summertime. By performing correlation and regression analyses among inversion, surface latent heat flux, SIC during the cold season, and SIC in the preceding warm season
(March to September), we also demonstrate that the surface albedo feedback in the warm season might play an important role in reducing the stable stratification of the lower troposphere in the following cold season through seasonal ocean heat uptake and release. The relationships derived in our study between the lower-tropospheric inversion and other fields help to clarify the mechanisms affecting Arctic amplification and could be used to better understand and interpret the large spread of responses to anthropogenic forcing in the Arctic simulated by climate models (Bintanja et al., 2012; Boe et al., 2009), and therefore might also be used in emergent constraint analysis to narrow uncertainty in model projection of Arctic changes (Hall et al., 2019; Lu, 2020; Thackeray & Hall, 2019). Given the projected Arctic sea ice decline and transition to an emerging new Arctic by climate models (Landrum & Holland, 2020; Wang & Overland, 2012), it is

Figure 4. Lower-tropospheric inversion in the cold season linked to sea-ice loss in the preceding warm season. (a) Area-weighted average of ONDJF-mean anomalies in surface upward latent heat (LE in red dot) and inversion (green dot) over the Arctic relative to 1979–2020. Black circles are reconstructed inversion anomalies based on the linear least squares fitting of green dots onto red dots. Panel (b) as in panel (a) but for LE and sea ice concentration (SIC in light blue dot). Panel (c) as in panel (a) but for SIC in ONDJF and SIC in MAMJJAS (deep blue dot). Panels (d–f) as in panels (a–c), but for detrended anomalies. Lines in panels (a–c) represent the linear trend marked in colors at the bottom (in per decade). The standard deviations (σ) in panels (d–f) represent the interannual variability of detrended time series. All of the trends and correlation coefficients R between solid dots in panel (a–c) and R between solid lines in panel (d–f) are statistically significant at the 99% confidence level.
essential to accelerate efforts to better resolve the debate about mechanisms of Arctic amplification (Serreze & Francis, 2006).

Data Availability Statement
All data used in this study are publicly available. The ERA5 monthly data on pressure levels from 1979 to present are available at https://cds.climate.copernicus.eu/cdsapp#!/dataset/reanalysis-era5-pressure-levels-monthly-means?tab=form. The ERA5 monthly data on single levels from 1979 to present are available at https://cds.climate.copernicus.eu/cdsapp#!/dataset/reanalysis-era5-single-levels-monthly-means?tab=form. The surface radiative kernel data are available at https://portal.nersc.gov/project/m1199/papers/2021GL094878/. The Berkeley Earth Surface Temperatures (BEST) monthly land and ocean surface air temperature gridded (1° × 1° latitude-longitude grid) data (1850–recent) are available at http://berkeleyearth.lbl.gov/auto/Global/Gridded/file name: Land_and_Ocean_LatLong1.nc. The Clouds and the Earth’s Radiant Energy System (CERES) Energy Balanced and Filled (EBAF) TOA Edition-4.1 monthly longwave flux data (1° × 1° grid) are available at https://ceres-data.larc.nasa.gov/ord-tool/jsp/EBAFTOA41Selection.jsp. The Atmospheric Infrared Sounder (AIRS) monthly integrated water vapor data (1° × 1° grid) are available at https://acdisc.gesdisc.eosdis.nasa.gov/data/Aqua_AIRS_Level3/AIRS3STM.006/.

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