Extratropical Ocean Warming and Winter Arctic Sea Ice Cover since the 1990s

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ABSTRACT

Despite the fact that the Arctic Oscillation (AO) has reached a more neutral state and a global-warming hiatus has occurred in winter since the late 1990s, the Arctic sea ice cover (ASIC) still shows a pronounced decrease. This study reveals a close connection (R = 0.5) between the extratropical sea surface temperature (ET-SST) and ASIC in winter from 1994 to 2013. In response to one positive (negative) unit of deviation in the ET-SST pattern, the ASIC decreases (increases) in the Barents–Kara Seas and Hudson Bay (the Baffin Bay and Bering Sea) by 100–400 km². This relationship might be maintained because of the impact of warming extratropical oceans on the polar vortex. Positive SST anomalies in the midlatitudes of the North Pacific and Atlantic (around 40°N) strengthen the equatorward planetary wave propagation, whereas negative SST anomalies in the high latitudes weaken the upward planetary wave propagation from the lower troposphere to the stratosphere. The former indicates a strengthening of the poleward meridional eddy momentum flux, and the latter implies a weakening of the poleward eddy heat flux, which favors an intensified upper-level polar night jet and a colder polar vortex, implying a stronger-than-normal polar vortex. Consequently, an anomalous cyclone emerges over the eastern Arctic, limiting or encouraging the ASIC by modulating the mean meridional heat flux. A possible reason for the long-term changes in the relationship between the ET-SST and ASIC is also discussed.

1. Introduction

Sea ice covers most of the Arctic Ocean during boreal winter and is a sensitive indicator of climate change. Satellite observations (from 1979 to the present) have indicated that the entire ice cover (seasonal and perennial ice) has declined over the past 30 years (Johannessen et al. 1995, 2004), and that this decline has accelerated during recent years (Comiso et al. 2008; Comiso 2012). Thus, Arctic sea ice cover (ASIC) trends and variability have become major foci of recent climate research (Li and Wang 2013; Li et al. 2014). Early studies have suggested that changes in atmospheric circulation (Proshutinsky and Johnson 1997), oceanic heat transport from the Atlantic (Polyakov et al. 2010) or Pacific Ocean (Woodgate et al. 2010), local radiative processes (Francis et al. 2005), and ice transport through the Fram Strait (Smelshrud et al. 2011; Langehaug et al. 2013) can lead to interannual changes in ASIC. The reduction of ASIC before the late 1990s has been attributed to the strongly positive Arctic Oscillation (AO) phase (Rigor et al. 2000, 2002). However, the recent, more neutral AO conditions (neither extreme negative nor positive phases, on average) may no longer be the key driver of sea ice trends (Lindsay and Zhang 2005).

As ASIC can be driven in part by poleward heat and moisture fluxes (Ukita et al. 2007; Yoo et al. 2014), it is
plausible that remote forcing such as enhanced warming over the Northern Hemisphere extratropical oceans may contribute to Arctic warming and sea ice reduction. Recently, an atmospheric modeling experiment forced by prescribed tropical SST (with a cooling trend in the eastern tropical Pacific) simulated surface warming over the Arctic (Ding et al. 2014). Nevertheless, more significant warming conditions are emerging in the subtropical North Pacific and North Atlantic, concurrent with cooling conditions in the central and eastern tropical Pacific (Wu et al. 2012). In this study, we explore the potential relationship between variations in the extratropical SST (ET-SST) and ASIC.

2. Data and methods

The four datasets employed in this study include the ERA-Interim data from 1979 to 2014 (resolution of 1.5° × 1.5°; Simmons et al. 2006); the National Centers for Environmental Prediction reanalysis, version 1 (NCEP1), from 1948 to 2014 (resolution of 2.5° × 2.5°; Kalnay et al. 1996); the National Oceanic and Atmospheric Administration (NOAA) SST data from 1854 to 2014 (resolution of 2.0° × 2.0°; Smith et al. 2008); and the Met Office Hadley Centre Sea Ice and SST dataset, version 1 (HadISST1), from 1870 to 2014 (resolution of 1.0° × 1.0°; Rayner et al. 2003).

In section 3, we analyze the period from 1994 to 2013, which includes the most recent decades of warm ET-SST (see Fig. 5a). We speculate that decades of warm ET-SST can potentially drive Arctic winter conditions, and this is discussed in section 4. We focus on the winter season [December–February (DJF)], or the dark season, because the albedo effect is practically absent (Graversen et al. 2008). Moreover, the extratropical warming effect is most pronounced during winter because stationary/transient eddies, which are the dominant mechanism for meridional heat transport (Trenberth et al. 2002), are stronger during winter than summer (Kosaka and Xie 2013). The Eliassen–Palm (E–P) flux (Andrews et al. 1987) is used to measure the planetary wave (wavenumber 1–3) activity propagation. The E–P flux is defined as follows:

\[ F(\phi) = -r_0 \cos \phi \frac{\partial u}{\partial \phi} \]

and

\[ F(p) = r_0 \cos \phi \left( \frac{\partial v}{\partial \phi} \right) \].

Here \( r_0 \) is the radius of Earth, \( \phi \) is the latitude, \( \theta \) is the potential temperature, \( u \) and \( v \) are the zonal and meridional wind velocities, respectively, \( p \) is the pressure, and \( \theta_p = \frac{\partial \theta}{\partial p} \). The prime symbols denote zonal deviations, and the overbars denote zonal averages. Accordingly, an equatorward E–P flux vector corresponds to a poleward meridional eddy momentum flux and an upward E–P flux is proportional to a poleward eddy heat flux. In this study, all data and indices were detrended before the analysis (e.g., atmosphere climate variables, SST, and sea ice).

3. Observed ET-SST–ASIC relationship

The leading empirical orthogonal function (EOF) mode for the ASIC anomalies (ASIC EOF-1) during 1994–2013 DJF (Fig. 1a) exhibits a seesaw pattern, with
one polarity in the Barents–Kara Seas and the opposite polarity in Baffin Bay and the Bering Sea. The corresponding principal component time series (ASIC PC-1) of the ASIC EOF-1 is shown in Fig. 1c (black dashed line). The correlation coefficient $R$ between the ASIC PC-1 and sea ice cover averaged over the Barents–Kara Seas (SIC-BaKa, in the domain 70°–80°N, 30°–80°E) is $-0.86$, and that between the ASIC PC-1 and sea ice cover averaged over the Bering Sea (SIC-Bering, in the domain 56.5°–66.5°N, 160°E–150°W) is approximately 0.65 (Fig. 1d). This means that positive PC-1 values correspond to reduced sea ice cover in the Barents–Kara Seas and to increased sea ice cover in the Bering Sea. The ASIC PC-1 has been used as a representative index to express ASIC variability in climate research (Deser et al. 2000; Deser and Teng 2008; Ukita et al. 2007).

In the current study, we mainly focus on the SST anomalies (SSTAs) over extratropical latitudes. The leading EOF mode for the ET-SSTAs (ET-SST EOF-1) during 1994–2013 DJF (Fig. 1b) is a horseshoe mode in the North Pacific and a zonally oriented pattern in the North Atlantic. The largest amplitudes occur in the central North Pacific and western–central North Atlantic, surrounded by anomalies with opposite signs. The principal component time series (ET-SST PC-1) of the ET-SST EOF-1 shows interannual variability (Fig. 1c, blue solid line). The linear correlation coefficient between the ET-SST and ASIC PC-1s (with trend) is 0.73. When the linear trends are removed (detrended), the correlation coefficient is 0.5 (above the 95% confidence level). This implies a potential link between the ET-SST and ASIC.

To further reveal the relationship between the ET-SST and ASIC, we present a regression between the ASIC anomalies and the ET-SST PC-1 during 1994–2013 DJF (Fig. 2b). Concurrent with the variability in the SSTAs displayed in Fig. 1b, there are significant negative sea ice anomalies in the Barents–Kara Seas and
positive ones in the Baffin Bay and Bering Sea. In response to one positive (negative) unit of deviation in the ET-SST PC-1, sea ice decreases (increases) in the Barents–Kara Seas (Baffin Bay, Bering Sea, and Sea of Okhotsk) by 100–400 km$^2$. It should be noted that the regressed map of ASIC (Fig. 2b) resembles the leading mode of the EOF for ASIC (Fig. 1a). The pattern reveals ASIC variability with respect to both its distribution and magnitude, reflecting a potential link with the annual variability in the ET-SST PC-1.

Figure 2a displays the changes in the DJF air temperature at 2 m ($T_{2m}$) based on the linear regression with the ET-SST PC-1 during 1994–2013. The east–west contrast in the surface air temperature (SAT) over the Arctic region is clearly apparent. Significant negative SAT anomalies are centered over the Baffin Bay and Bering Sea. Moreover, there are significant positive SAT anomalies centered over the Barents–Kara Seas and adjacent regions. The spatial distribution of surface warming (cooling) with a magnitude of 0.5$^\circ$–2.5$^\circ$C is consistent with the regional negative (positive) sea ice anomaly.

How might ET-SSTAs potentially impact the variability of Arctic sea ice? Considering that the variability in stationary planetary wave activity is strongly related to surface thermal conditions and their impacts on atmospheric circulation (Andrews et al. 1987), we examine the wave–mean flow interactions associated with ET-SSTAs. Figures 3a and 3b illustrate the zonally averaged zonal wind (red contours), E–P flux cross sections (vectors), and the divergence (shaded) for high and low ET-SST PC-1s during 1994–2013 DJF. Clear upward propagating planetary waves (vectors) from the mid-latitudes of the lower troposphere, which bifurcate into two branches in the upper troposphere, are observed in both cases (Figs. 3a,b). One branch propagates straight into the stratosphere along the polar waveguide (Dickinson 1968), and the other propagates equatorward along the low-latitude waveguide (Huang and Gambo 1982). Compared with the situation related to the lower ET-SST PC-1 (Fig. 3b), the positive SSTAs in the midlatitudes of the North Pacific and Atlantic (around 40$^\circ$N) associated with the higher ET-SST PC-1 would lead to a relatively stronger low-latitude waveguide (Fig. 3c, 20$^\circ$–60$^\circ$N, surface–50 hPa: vectors). In contrast, the negative SSTAs in the high latitudes of the North Pacific might lead to a weaker polar waveguide (Fig. 3c, 50$^\circ$–80$^\circ$N, 500–10 hPa: vectors), which would contribute to E–P flux convergence anomalies in the upper troposphere around 30$^\circ$N and divergence anomalies around 60$^\circ$N (Fig. 3c: red contour). According to the assumptions of quasigeostrophic theory and linear perturbations [Andrews et al. 1987, their Eq. (3.5.5a)], the zonal-mean flow decelerates (accelerates) where the E–P flux converges (diverges). Consequently, the westerly zonal-mean flow, which is associated with the ET-SSTA distribution shown in Fig. 1b, decelerates in the troposphere around 30$^\circ$N (Fig. 3c: blue contours). In contrast, positive westerly anomalies are found in the region of the polar night jet (around 60$^\circ$N), which lead to the strengthening of the polar vortex. However, the
downward propagating E–P flux anomalies along the polar waveguide (around 60°N) might strengthen the polar vortex because it is proportional to the weakening of the poleward eddy heat transport (Peings and Magnúsdottrí 2014). Moreover, such an ET-SST pattern, with positive anomalies in midlatitudes (around 40°N) and opposite sign in the north, would strengthen the meridional temperature gradient in high latitudes, which might also directly favor an intensified polar night jet.

Because of the zonal average in the definition of the E–P flux, it is difficult to identify the relative contribution of different oceans (i.e., the Pacific and Atlantic). In other words, are the sources of the planetary wave mentioned above consistent with the spatial distribution of the ET-SSTAs? To answer this question, we further calculate the winter vertical and horizontal stationary wave activity flux (Plumb 1985). The top panels of Fig. 4 show the climatological vertical wave activity fluxes (contours) and their anomalies (shaded) regressed against the standardized ET-SST PC-1 during 1994–2013 DJF. Dotted regions indicate significant values based on the 95% confidence level from a two-tailed Student’s t test. Quasigeostrophic streamfunction (contours, 106 m2 s⁻²) and horizontal wave activity flux (vectors, scale: m² s⁻²) anomalies at (d) 500, (e) 300, and (f) 100 hPa regressed against the standardized ET-SST PC-1 during 1994–2013 DJF. Shaded regions indicate significant values based on the 95% confidence level from a two-tailed Student’s t test. For display, the zero line is ignored.

FIG. 4. Climatological vertical stationary wave activity fluxes (contours; intervals: 30, 15, and 2 × 10⁻³ m² s⁻², respectively) and their anomalies (shaded; 10⁻² m² s⁻²) at (a) 500, (b) 300, and (c) 100 hPa regressed against standardized ET-SST PC-1 during 1994–2013 DJF. Dotted regions indicate significant values based on the 95% confidence level from a two-tailed Student’s t test. Quasigeostrophic streamfunction (contours, 10⁶ m² s⁻¹) and horizontal wave activity flux (vectors, scale: m² s⁻²) anomalies at (d) 500, (e) 300, and (f) 100 hPa regressed against the standardized ET-SST PC-1 during 1994–2013 DJF. Shaded regions indicate significant values based on the 95% confidence level from a two-tailed Student’s t test. For display, the zero line is ignored.
The bottom panels of Fig. 4 show the quasigeostrophic streamfunction (contours) and horizontal wave activity flux (vectors) anomalies at 500, 300, and 100 hPa regressed against the standardized ET-SST PC-1 during winter from 1994 to 2013. The main features revealed in Fig. 4d are the positive streamfunction anomalies located in the extratropical North Pacific and Atlantic, negative streamfunction anomalies located in the Pacific–Arctic and Atlantic–Arctic sectors, and associated anomalous eastward stationary waves. The wave sources are mainly located in the North Pacific and partially in the eastern North Atlantic, which is consistent with the vertical wave activity fluxes. This pattern also extends upward to 300 hPa (Fig. 4e). In contrast, the negative streamfunction anomalies over the Pacific–Arctic sector amplify significantly at 100 hPa (Fig. 4f), suggesting a deepened polar vortex. In summary, the ET-SST can significantly impact the vertical and horizontal propagation of stationary waves. It should be noted that the anomalous vertical stationary wave activity over the North Pacific is relative stronger compared with that over the North Atlantic. Specifically, the anomalous upward stationary wave activity (Fig. 4a; 20°–60°N, 500 hPa) in the North Pacific is approximately 69% higher than that in the North Atlantic.

The strengthened polar vortex is more evident in Figs. 2d and 2e, which display the 300- and 100-hPa geopotential height and air temperature (Z300 and T300, and Z100 and T100) anomalies regressed against the standardized ET-SST PC-1, respectively. Corresponding to stronger (weaker) ET-SSTAs (Fig. 1b), the polar vortex becomes colder (warmer) (Figs. 2d,e, blue contours). This means that the polar vortex is deeper (shallower) and characterized by significant negative (positive) geopotential height anomalies over the Arctic region in the upper troposphere (Figs. 2d,e, black contours). Moreover, the sea level pressure (SLP) anomaly shows the cyclonic circulation anomaly over the North Atlantic (Fig. 2c, blue contours). These results suggest an intensified (weakened) polar vortex structure from surface to the upper troposphere. Consistent with the change in the atmospheric circulation, an obvious anomalous northward surface heat transport is found over the Barents–Kara Seas (Fig. 2c, vectors), which limits the sea ice growth there (Fig. 2b). In contrast, there is a southward surface heat transport around Greenland and over the western Bering Sea, which encourages sea ice growth there. Additionally, there is an eastward surface heat transport over the eastern Bering Sea (Fig. 2c, vectors), which might be attributed to the intensified polar night jet. Correspondingly, negative T2m anomalies emerge in the Bering Sea (Fig. 2a) and the sea ice shows positive anomalies (Fig. 2b). Therefore, the variability in the ET-SST could have potentially influenced the winter ASIC during 1994–2013 by modulating the corresponding atmospheric circulation.
4. Possible reason for the ET-SST–ASIC relationship change based on observations

By examining time series from only the most recent two decades, a close connection between the ET-SST and ASIC is identified (Fig. 2c). However, from a long-term perspective, the relationship between the ET-SST and ASIC is not stable (see Fig. 5b). Using long-term SST observational data, we present evidence that changes in the ET-SST PC-1 generally occur in conjunction with low-frequency variability: decades of warm and cold SST alternate from 1870 to 2013 (Fig. 5a). We speculate that decades of warm ET-SST can drive Arctic winter conditions. Figure 5b illustrates a wavelet coherence analysis (Grinsted et al. 2004) of the ASIC and ET-SST PC-1s. Large sections with a statistically significant in-phase behavior exist between the two time series at a period of <10 yr from 1948 to 1970s and a period of 4–7 yr after 1994.

Based on the identified change in the ET-SST–ASIC relationship and the low-frequency variability in ET-SST, we select another two periods, 1948–75 and 1976–93 (decades of warm and cold ET-SST, respectively), for further analysis. The spatial distribution of the T2m, SLP and 850-hPa horizontal heat flux (HHF850), and Z100 and T100 anomalies related to the ET-SST PC-1 during 1948–75 and 1976–93 DJF is illustrated in Fig. 6. The SAT anomalies during 1948–75, associated with the positive ET-SST phase, closely resemble those during 1994–2013, which display an east–west contrast in the SAT over the Arctic region. Significant negative (positive) SAT anomalies are centered over the Baffin Bay and Alaska Peninsula (the Barents–Kara Seas and Sea of Okhotsk) (Fig. 6a). The spatially variable characteristics of SLP and Z100 and T100 suggest a strengthened polar vortex from the surface to upper troposphere; Z100 decreases in the Northern Hemisphere polar region (Fig. 6e, black contours), coinciding with the tropospheric cooling (Fig. 6e, blue contours). Meanwhile, the SLP anomaly is negative over the Arctic region, and its center is located in Iceland (Fig. 6e, blue contours). However, the atmospheric circulation and temperature anomalies over the Arctic region from 1976 to 1993 are very weak and statistically insignificant (Fig. 6, bottom panels).

To identify the possible reason for the change in the ET-SST–ASIC relationship, we investigate the
regression between the quasigeostrophic streamfunction (contours) and horizontal wave activity flux (vectors) anomalies at 300 hPa regressed against the standardized ET-SST PC-1 during 1948–75 and 1976–93 DJF. Decadal composite of quasigeostrophic streamfunction (contours, $10^6 \text{ m}^2 \text{s}^{-1}$) and horizontal wave activity flux (vectors, scale: $\text{m}^2 \text{s}^{-2}$) at 300 hPa (c) between 1948–75 and 1976–93 and (d) between 1994–2013 and 1976–93 DJF. Shaded regions indicate significant values in the 300-hPa quasigeostrophic streamfunction based on the 95% confidence level from a two-tailed Student’s $t$ test. For display, the zero line is ignored.

5. Results from numerical simulations

To test our hypothesis, we conduct simulations using the National Center for Atmospheric Research Community Atmospheric Model, version 3 (CAM3; Collins et al. ...
The simulation configuration has a horizontal resolution of approximately 2.8° and 26 vertical levels extending up to 3.5 hPa. Two experimental designs are employed. For the first experimental design, we perform 10 simulations in which CAM3 is forced by seasonally varying climatological-mean SST and sea ice (1979–2010), except for the Northern Hemisphere extratropical oceans (20°–70°N), for which the model is forced by the observed 35-yr (from January 1979 to January 2013, 35 × 12 months) Northern Hemisphere ET-SST. For the second experimental design, we perform 20 control simulations in which CAM3 is forced by seasonally varying climatological mean SST and sea ice (1979–2010), and 20 sensitivity simulations in which the model is forced by perturbed ET-SST and NP (North Pacific)-SST; all other external variables remain fixed. The perturbed 3-month (from December to February) ET-SST and NP-SST, which is obtained from the EOF analysis in Fig. 1b, is added to the climatological-mean values. The prescribed SST in both experimental designs is from the NOAA weekly optimum interpolation SST analysis (Hurrell et al. 2008).

The model analysis parallels the observational analysis to facilitate comparison between the two. Figure 8 illustrates the ensemble-mean 1994–2013 and 1979–93 SAT, SLP and HHF850, and Z300 and T300 anomalies related to the ET-SST PC-1 in the DJF mean fields in CAM3, forced by the observed Northern Hemisphere ET-SST (20°–70°N) (the first experimental design); over the Arctic region they closely resemble the observed anomalies for each period, as shown in Figs. 2 and 6 (bottom panels) respectively. During 1994–2013, the east–west contrast in SAT, the cyclonic circulation anomaly over the North Atlantic, and the intensified upper-tropospheric polar vortex (Fig. 8, top panels) are an integral part of the response to the imposed ET-SSTAs. In contrast, during 1979–93, the simulated anomalies over the Arctic region are statistically insignificant (Fig. 8, bottom panels). The model simulation confirms speculations that winter ET-SST could influence atmospheric circulation, which further causes variations in the simultaneous ASIC during 1994–2013, but winter ET-SST has no significant effect on the simultaneous ASIC during 1979–93.
experiments in the DJF mean fields in CAM3, forced by perturbed ET-SST (20°–70°N), respectively (the second experimental design). Similar to the above results, over the Arctic region they closely resemble the observed anomalies during 1994–2013, as shown in Fig. 2. Moreover, regions showing positive (negative) streamfunction anomalies are found in the midlatitudes of the North Pacific and Atlantic (Pacific–Arctic and Atlantic–Arctic sectors), which are accompanied by anomalous westward stationary waves that originate from the North Pacific (Figs. 9d and 10d). Although the regional details differ somewhat between the responses of the modeled SAT (Fig. 9a) and observed SAT (Fig. 2a), the simulations show below-normal winter SAT in Scandinavia. The simulations also confirm that the winter ET-SST (mostly NP-SST) is linked to simultaneous variations in the ASIC through their impacts on the cyclonic circulation anomaly over the North Atlantic and the intensified upper-tropospheric polar vortex.

6. Conclusions and discussion

In this study, we find close covariability between the ET-SST and ASIC during winter from 1994 to 2013, with a significant correlation coefficient of 0.5 between their principal component time series of the first leading EOF mode. The ASIC shows negative (positive) anomalies in the Barents–Kara Seas and Hudson Bay (Baffin Bay and Bering Seas), which are associated with positive SSTAs in the midlatitudes of the North

**Fig. 9.** Differences in patterns between the perturbed and controlled experiments in the DJF mean fields in CAM3, forced by perturbed ET-SST (20°–70°N), which is obtained from the EOF analysis in Fig. 1b and added to climatological-mean values (the second experimental design). (a) SAT (°C). Dotted regions indicate significant values based on the 95% confidence level from a two-tailed Student’s t test. (b) SLP (blue contours, hPa) and HHF850 (vectors, scale: m K s⁻¹), (c) Z300 (black contours, gpm) and T300 (blue contours, °C), and (d) quasigeostrophic streamfunction (contours, 10⁶ m² s⁻¹) and horizontal wave activity flux (vectors, scale: m² s⁻²) at 300 hPa. Shaded regions indicate significant values in HHF850, Z300, and 300-hPa quasigeostrophic streamfunction in (b)–(d), respectively, based on the 95% confidence level from a two-tailed Student’s t test. For display, the zero line is ignored.
Pacific and Atlantic (around 40°N) and negative SSTAs to the south and north of these regions. The link between ET-SST and ASIC might be maintained through its impact on the polar vortex. It is revealed that the upper-tropospheric polar temperatures are lower and the upper-tropospheric polar vortex is strengthened during the high ET-SST PC-1.

Compared with the situation related to the lower ET-SST PC-1, the positive SSTAs in the midlatitudes of the North Pacific and Atlantic (around 40°N) associated with the higher ET-SST PC-1 cause an anomalous upward and equatorward E–P flux (Fig. 3c; 20°–60°N, surface–50 hPa; vectors). Therefore, the polar night jet strengthens because an equatorward-pointing E–P flux vector corresponds to a poleward meridional eddy momentum flux. Moreover, the negative SSTAs in the high latitudes of the North Pacific and Atlantic might lead to an anomalous downward E–P flux (Fig. 3c; 50°–80°N, 500–10 hPa; vectors), which would favor a colder polar vortex because the upward E–P flux is proportional to the poleward eddy heat flux. This means that the high ET-SST PC-1 period is characterized by weakened upward propagating planetary waves from the midlatitudes of the lower troposphere (around 60°N) to the stratosphere, which cause a weakening of the poleward eddy heat transport and induces an anomalous divergence of E–P flux (around 60°N from the lower troposphere to the stratosphere). The former leads to a colder polar vortex, and the latter favors an intensified polar night jet, both of which indicate a stronger-than-normal polar vortex. Consequently, an anomalous surface cyclonic circulation pattern emerges over the eastern Arctic, which limits or encourages the Arctic sea ice growth in the Barents–Kara Seas and Hudson Bay (Baffin Bay and Bering Seas) by modulating the mean meridional heat flux.

The long-term change in the relationship between ET-SST and ASIC appears to be related to the low-frequency ET-SST variability. We speculate that warm decades of ET-SST can drive the winter ASIC. Through the analysis of the decadal composite of the quasigeostrophic streamfunction and horizontal wave activity flux at 300 hPa...
between decades of warm (1948–75 and 1994–2013) and cold (1976–93) ET-SST, we reveal positive, negative, and positive streamfunction anomalies located in the extratropical North Pacific, western Canada, and western North Atlantic, respectively, which are accompanied by a decadal Rossby wave train that may originate from the North Pacific and extends poleward and eastward. This decadal Rossby response may contribute to the strengthened connection between the ET-SST and ASIC during 1948–75 and 1994–2013. Unfortunately, the period from 1994 to 2013 (20 yr) may be not long enough to draw robust conclusions, and more observational data are needed to test our mechanism in the future.

In the current study, we also conduct numerical experiments and two experimental designs are employed to test the mechanism proposed based on the observational data. For the first experimental design, the atmospheric response to the observed Northern Hemisphere ET-SST during 1994–2013 and 1979–93 DJF closely resembles the observed anomalies during each period. For example, during the period from 1994 to 2013, the east–west contrast in SAT, the cyclonic circulation anomaly over the North Atlantic, and the intensified upper-tropospheric polar vortex occur in the Arctic. In contrast, the simulated anomalies over the Arctic region during 1979–93 are statistically insignificant. For the second experimental design, the atmospheric response to both perturbed ET-SST and NP-SST, which is obtained from the EOF analysis in Fig. 1b and added to the climatological-mean values, includes positive (negative) streamfunction anomalies located in the midlatitudes of the North Pacific and Atlantic (Pacific–Arctic and Atlantic–Arctic sectors) and anomalous westward stationary waves that originate from the North Pacific. The consistency between the simulations and observations supports the hypothesis that the winter ET-SST (mostly NP-SST) is linked to the simultaneous variations in ASIC through their impacts on atmospheric circulation (including the cyclonic circulation anomaly over the North Atlantic and the intensified upper-tropospheric polar vortex).

Additionally, the Arctic Ocean is tightly connected to the global ocean system via water mass exchange with the Pacific and Atlantic Oceans through several main oceanic gateways: the Bering Strait, Canadian Arctic Archipelago, Davis Strait, Fram Strait, and the entrance to the Barents Sea (Beszczynska-Möller et al. 2011). The oceanic heat transport from the Pacific and Atlantic inflow might be an important trigger for the recent sea ice decline (e.g., Woodgate et al. 2010; Smedsrud et al. 2013). For example, Årthun et al. (2012) concluded on the basis of observations and model simulations that the interannual variability and long-term trend of the sea ice in the Barents Sea are associated with the variability in ocean heat transport due to the Atlantic inflow. In addition, other studies have shown that Arctic near-surface warming is mainly caused by the reduction in Arctic sea ice (e.g., Screen and Simmonds 2010). It is difficult to quantify the relative role of the ocean and atmosphere in forcing the change of winter ASIC. This study suggests that the ET-SST included atmospheric warming is also a potential driver of the winter ASIC.

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