Role of Oceanic Memory Effects in the Barents Sea in the Seasonal Linkage Between the Winter and Summer Arctic Oscillation

Ryosuke Hoshi1,2 and Hiroshi G. Takahashi1

1Department of Geography, Tokyo Metropolitan University, Hachioji, Japan, 2Japan Meteorological Agency, Tokyo, Japan

Abstract We investigated the mechanism of inter-seasonal linkage between the winter Arctic Oscillation (AO) and the subsequent summer AO. Our focus was the collapse of the linkage in the late 1980s, associated with substantial sea ice loss in the Arctic. Because atmospheric circulation itself does not have long-lasting memory effects from winter to summer, the seasonal persistency of sea ice and sea surface temperature (SST) anomalies are likely to contribute to the inter-seasonal linkage. In the early significant-linkage period prior to 1990, the winter AO signal was maintained in the sea ice and SST anomalies over the Barents Sea during inter-seasonal periods. The surface anomalies then influenced the summer atmospheric circulation by exciting the Rossby wave propagation. However, in the recent poor-linkage period after 1990, sea ice loss in the Barents Sea and Northern Hemisphere was significant. In addition, the seasonal persistency of sea ice and SST anomalies over the Barents Sea were less significant, and the winter AO signal was not maintained beyond a given season. Thus, the oceanic memory effects of the Barents Sea were reduced and might not have been associated with the recent inter-seasonal linkage.

1. Introduction

Arctic Oscillation (AO) is a dominant statistical mode of atmospheric variability in the Northern Hemisphere. It is characterized by a north-south seesaw pattern of atmospheric mass between the Arctic and middle latitudes (Thompson & Wallace, 1998, 2000). The variability of the AO is strongest in winter, and it has a substantial impact on the winter climate of the Northern Hemisphere (Thompson & Wallace, 2001; Wang et al., 2005).

Several studies have investigated the influences of the winter AO on the hemispheric weather and climate in winter, while a few studies (e.g., Ogi et al., 2005; Otomi et al., 2013) have proposed that the anomalous positive summer AO condition can be related to unusual hot weather on a planetary scale. The spatial structure of the summer AO is displaced poleward and has a smaller meridional scale than that of the winter AO (Ogi, Yamazaki, & Tachibana, 2004). To extract seasonal differences in the spatial pattern of the AO, Ogi, Yamazaki, and Tachibana (2004) performed an empirical orthogonal function (EOF) analysis for individual calendar months. The axes of the polar and subtropical jets are located along the 70°N–75°N and 40°N–45°N latitudinal bands in a large positive phase of the summer AO, which is often referred to as the double-jet stream structure (Ogi, Yamazaki, & Tachibana, 2004; Ogi et al., 2005). In contrast, when the summer AO index is small or negative, the polar jet is absent, and the double-jet stream structure is unclear. Maeda et al. (2000) and Tachibana et al. (2010) indicated that the double-jet stream structure is related to the occurrence of atmospheric blocking, which contributes to long-lasting unusual weather anomalies such as heat waves and droughts.

In general, unusual weather such as cold air outbreaks associated with blocking tend to occur in the negative phase of the winter AO (Thompson & Wallace, 2001). On the other hand, Tachibana et al. (2010) demonstrated that anomalous blocking tends to occur and causes unusual hot weather in the positive phase of the summer AO. During the anomalous positive summer AO phases, such as those in the summers of 2003 and 2010, Europe, Russia, and Canada experienced extremely hot weather with a double-jet stream structure and blocking anticyclones in northeastern Europe and the northern Sea of Okhotsk, which are parts of a planetary-scale wave train (Feudale & Shukla, 2011; Grumm, 2011; Ogi et al., 2005; Otomi et al., 2013). Extremely hot weather with an anomalous positive summer AO phase causes enormous crop losses, droughts, forest fires,
and mortalities (Feudale & Shukla, 2011). Thus, it is important to investigate long-term forecasting of such anomalous summers.

In general, atmospheric circulation does not have long-lasting memory effects that persist from winter to summer. Thus, few studies have discussed the link between the winter AO and the subsequent summer AO. In contrast, the ocean has a larger heat capacity than the atmosphere; therefore, sea ice and sea surface temperature (SST) anomalies can persist for longer than, for example, anomalies in air temperature. Thus, anomalies in air temperature impact sea ice and SST anomalies, which can then be fed back to the atmosphere in later seasons. Rigor et al. (2002) demonstrated that sea ice concentration in the Arctic Ocean is affected by surface winds associated with the AO in winter, and this sea ice anomaly remains until the subsequent spring and summer. The amount of open water and thin ice associated with fluctuations in the heat flux from the ocean has a strong effect on the surface air temperature. Thus, they showed that sea ice concentration and surface air temperature in spring and summer were related to the AO in the previous winter.

Ogi et al. (2003) and Ogi, Yamazaki, and Tachibana (2004) examined the inter-seasonal linkage between the winter AO and the subsequent summer AO, which was analyzed from the NCEP-NCAR reanalysis data set (Kalnay et al., 1996) for the second half of the 20th century. They focused on the possibility that sea ice and SST anomalies play a role in the connection between winter AO and the atmospheric circulation in the subsequent summer in the Northern Hemisphere by conveying climatic memories from winter to summer. They showed that the winter AO and the subsequent summer AO have lagged positive correlations; when the winter AO is in a positive phase, the summer AO also tends to be in a positive phase, and when the winter AO is in a negative phase, the summer AO also tends to be in a negative phase. This is referred to as an inter-seasonal linkage. In contrast, the summer AO and the subsequent winter AO did not have significantly lagged correlations (Ogi, Yamazaki, & Tachibana, 2004).

Recently, Yamazaki et al. (2019) reexamined the seasonal linkage between the winter AO/North Atlantic Oscillation (NAO) and the subsequent summer AO using a longer reanalysis data set. The winter AO/NAO and the subsequent summer AO had lagged positive correlations lasting from the mid-1960s to the 1980s. However, the seasonal linkage collapsed circa 1990. In the last two to three decades, the correlation was weak or negative. Yamazaki et al. (2019) discussed the possible causes of seasonal linkage collapse, such as stratospheric ozone, solar activity, North Atlantic SST, Arctic Sea ice, Atlantic Multidecadal Oscillation (AMO), and chaotic internal variability in the climate system. However, the detailed mechanism and main cause of seasonal linkage collapse remain unclear.

In the last two decades, the surface air temperature increase in the Arctic has been more than double the global average (Chylek et al., 2009; Meredith et al., 2019; Serreze et al., 2009). Ice loss in the Arctic Ocean is a prominent indicator of global warming. In recent years, sea ice in the Arctic Ocean has rapidly decreased and has become much younger and thinner, with the ice melt season starting earlier (Comiso et al., 2017; Stroeve & Notz, 2018). When sea ice melts, more open water spreads and reduces the albedo, which consequently increases solar absorption and further melts the sea ice. Thus, the shift to thinner seasonal sea ice accelerates sea ice loss (Nicolaus et al., 2012). Additionally, sea ice loss in the Arctic has the potential to influence mid-latitude weather (Meredith et al., 2019). For example, some studies have indicated that sea ice loss could cause a cold anomaly in the mid-latitudes such as Eurasia in winter (Honda et al., 2009; Inoue et al., 2012; Outten & Esau, 2012).

The primary goal of this study was to determine the mechanism of inter-seasonal linkage between the winter AO and subsequent summer AO. This study focused on the long-term changes in the seasonal evolution of sea ice. Because atmospheric circulation does not have inter-seasonal memory effects, the seasonal persistency of sea ice and SST anomalies are likely to contribute to the inter-seasonal linkage. Focusing on the Arctic region, where the climate is changing rapidly, the atmosphere–ocean–ice interaction was examined in detail, and the influence of this rapid ice loss on the inter-seasonal linkage was revealed. Section 2 describes the data and methods used in this study. Section 3 presents the results, which detail the long-term changes in inter-seasonal linkage. The role of sea ice and SST in long-term changes in inter-seasonal linkage is discussed in Sections 4 and 5. Finally, the conclusions are presented in Section 6.
2. Data

We used the Japanese 55-year Reanalysis data set (JRA-55; Kobayashi et al., 2015) from 1958 to 2017 as the meteorological data, while Yamazaki et al. (2019) used the NCEP-NCAR reanalysis data set. Thompson and Wallace (2000) first defined the AO. They performed a single EOF analysis of the zonally averaged geopotential height field from 1,000 to 50 hPa in the Northern Hemisphere and indicated that the AO existed throughout the year. However, the AO defined by Thompson and Wallace (2000) mostly reflects winter influence because atmospheric variability is largest during winter.

Ogi, Yamazaki, and Tachibana (2004) performed an EOF analysis of the zonally averaged geopotential height for individual calendar months from 1,000 to 200 hPa, north of 40°N, to accurately determine the seasonal variations of the AO. They named the time series of the new AO as the seasonally varying northern annular mode (SV NAM) index. Figure 1 shows the time series of the conventional AO index and SV NAM index for January and July. Some studies (e.g., Ogi et al., 2005; Ogi, Yamazaki, & Tachibana, 2004; Otomi et al., 2013) have shown that although the SV NAM index coincides with the conventional AO index in winter, both indices are different in summer. They also indicated that while the SV NAM can capture extremely hot weather conditions, such as during the summers of 2003 (Ogi et al., 2005) and 2010 (Otomi et al., 2013), the conventional AO defined by

![Figure 1](image-url)
Thompson and Wallace (2000) cannot capture such hot summers. We used the SV NAM index calculated from JRA-55 from 1958 to 2017, and all mentions of the AO index refer to the SV NAM index hereafter.

The monthly mean sea ice concentration and SST used in this study were retrieved from the Hadley Center Sea Ice and SST data set (HadISST; Rayner et al., 2003) on a 1° latitude-longitude grid from 1958 to 2017. The sea ice data were obtained from various data sources, including digitized sea ice charts from ship observations and passive microwave satellite retrievals. In the pre-satellite period, while the sea ice data near the ice edge were the most reliable, owing to many historical reports from ship observations, data for areas where the ship observation was difficult might have been comparatively uncertain. To omit low-accuracy data during the pre-satellite era, the interannual variations in seasonally averaged sea ice concentrations were verified. When the grid value was the same as that used in the previous year, we counted. If the count was more than half of the number of observations during the pre-satellite era, the grid was omitted. The sea ice concentration values were in tenths of the ice coverage on a grid.

In the data set, the SST data with sea ice cover of 100% at a grid were fixed at −1,000, and data greater than 90% at a grid were fixed at −1.8. Fixed values of SST, such as −1.8 and −1,000 in the polar region, may have relatively large uncertainty. Thus, fixed values of SST were regarded as missing data in the analysis to accurately elucidate the effect of SST on inter-seasonal linkage.

To examine the inter-seasonal linkage between the winter AO and the subsequent summer climate, lag-correlation coefficients were calculated. When the correlation coefficients were calculated, a high-pass filter was applied to remove the 11-year running mean of the decadal variability from the original time series, and the filtered components of year-to-year variations were used. This was performed because inter-seasonal linkage was the focus of this study. Statistical significance was determined using a two-sided test with a null hypothesis of non-correlation.

3. Long-Term Changes in the Inter-Seasonal Linkage

Yamazaki et al. (2019) showed that the inter-seasonal linkage between the winter AO/NAO and the subsequent summer AO changed before and after 1990. However, they did not demonstrate the transition of spatial patterns from the early good-linkage to the recent poor-linkage periods. We examined the transition of the north-south seesaw pattern between the winter AO and the atmospheric circulation in the subsequent summer and confirmed when the inter-seasonal linkage was broken, in terms of spatial pattern. To understand this long-term change, the lag-correlation coefficients between the AO index in winter (December–February) and the monthly zonal mean 500-hPa geopotential height were calculated on an interannual timescale for 20 years, using a sliding correlation (Figure 2).

In January and February, a meridional seesaw pattern in the zonal mean 500-hPa geopotential height of simultaneous positive correlation in the mid-latitudes and negative correlation in the high latitudes was dominant for all periods. The node was near 55°N, where the westerly winds were strong (or weak) when the AO index was positive (or negative) in winter (Thompson & Wallace, 2000), which represents a typical AO structure in northern winter. However, the meridional seesaw pattern in the zonal mean 500-hPa geopotential height almost disappeared during March–May for all periods, indicating that the winter AO signals disappeared during the northern spring. In the early period (Figures 2a and 2b), a weak positive correlation was only observed near 55°N–60°N in April. In terms of spatial distribution, weak positive correlations in spring were scattered over Alaska, Siberia, and northern Europe (not shown).

In summer (June, July, and August), the results were considerably different between the early and late periods. In the first half of the period (Figures 2a–2c), a meridional seesaw pattern of positive correlation in the mid-latitudes and negative correlation in the high latitudes appeared again in the subsequent summer. However, the seesaw pattern of the 500-hPa geopotential height anomalies in summer was located more to the north than that in winter, which could be associated with the northward displacement of the westerly winds with the seasonal march (Ogi, Yamazaki, & Tachibana, 2004). This result shows that the winter AO is linked to atmospheric circulation in the subsequent summer. Specifically, an atmospheric circulation anomaly with a positive AO phase in winter is likely to be followed by an atmospheric circulation anomaly with the same positive AO phase in the subsequent summer.
Figure 2. Lag-correlation coefficients between the winter (DJF) AO index and the monthly zonal mean 500-hPa geopotential height on an interannual timescale of 20 years, using a sliding correlation by 5 years. (a) 1963–1982, (b) 1968–1987, (c) 1973–1992, (d) 1978–1997, (e) 1983–2002, (f) 1988–2007, and (g) 1993–2012. The shadings indicate that the correlation coefficients exceed the confidence levels (90%, 95%, and 99%), as shown in the legend.
However, after 1978–1997 (the center of the period was in the late 1980s), the seesaw pattern did not reappear in the subsequent summer (Figures 2d–2g). This indicates that the inter-seasonal linkage between the winter AO and the subsequent summer AO was broken. In addition, in the most recent period (1993–2012), a seesaw pattern between negative correlation at 30°N–50°N and positive correlation at 75°N–90°N was observed in summer (Figure 2g), which seems opposite to the meridional seesaw pattern in the early period. The results reconfirmed the dramatic changes in the inter-seasonal linkage between the winter AO and atmospheric circulation in the subsequent summer around the late 1980s. This result is consistent with the findings of Yamazaki et al. (2019). In addition, we confirmed the long-term change in the inter-seasonal linkage between the winter AO and atmospheric circulation in the subsequent summer in terms of spatial distribution (see Figure A1 in the Appendix A), which also shows the same results.

4. Oceanic Memories Left by the Winter AO Signal

4.1. Seasonal Linkage Among the Winter AO, Sea Ice, and SST in Spring and Summer

Essential concerns regarding the inter-seasonal AO linkage, for instance, the cause of the break in the inter-seasonal linkage between the winter AO and the subsequent summer AO, remain unaddressed. This work focuses on the changes in oceanic memory effects (sea ice and SST) in the Arctic region because of the recent dramatic decrease in sea ice. Memory effects play a role in preserving the winter AO signal in the ocean and conveying climatic memories from winter to summer. To understand the process of the oceanic memory effects, the lag relationship between the AO index in winter and sea ice concentration in the Northern Hemisphere in the subsequent months (February–July) for the early (1963–1989) and recent (1990–2012) periods were analyzed (Figures 3 and 4).

Positive correlation signals were observed in the west of Greenland, while negative correlation signals were found in the Barents Sea in both early and recent periods. A positive correlation indicates that sea ice concentration increases in a target month after a positive phase of the winter AO and vice versa. The significant negative correlation in the Barents Sea indicates that sea ice is reduced when the AO index in winter is positive, which agrees with the results of Dickson et al. (2000). The Icelandic Low is enhanced during the positive winter AO phase. The enhanced Icelandic Low brings strong near-surface southwesterly winds and more transport of warm North Atlantic water into the Barents Sea (Armitage et al., 2018; Wu et al., 2004), which tends to warm the sea and reduce sea ice. Conversely, when the AO index in winter is positive, strong northerly winds blow to the west of Greenland (Wu et al., 2004), which increases the local sea ice. Comparing the results between the early and recent periods every month, the lag relationship between the winter AO and sea ice concentration in the subsequent months changed before and after the late 1980s, particularly in the Barents Sea. In the early period, the negative correlation in the Barents Sea appeared most clearly in April and persisted until May. After June, the negative signal was unclear. In addition, a positive correlation was evident west of Greenland until May.

In contrast, in the recent period, the negative correlation signal in the Barents Sea appeared most clearly in March. However, it weakened in April and diminished in May. These results suggest that the winter AO signal persisted longer in sea ice anomalies over the Barents Sea until April and May in the early period, whereas in the recent period did not persist until May. In the west of Greenland, the positive signal in the recent period persisted until May, although the statistically significant areas in March–April were smaller than those in the early period. In addition, a negative correlation was observed in the East Siberian Sea in July.

The long-term changes in the lag relationship between the winter AO and SST in the Northern Hemisphere in the subsequent months (March–August) were examined (Figures 5 and 6). In the early period, a positive correlation signal was evident in the Barents Sea until August. A positive correlation indicates that SST becomes warmer in the target month after a positive phase of the winter AO and vice versa. Focusing on the positive signal of the Barents Sea in detail, the signal was centered in the western part of the Barents Sea until May; however, it moved east after June. In general, the summer SST in the Barents Sea is greatly affected by sea ice anomalies during the previous spring (Bushuk et al., 2015; Smedsrud et al., 2013). For example, when the spring sea ice concentration has a negative anomaly, the summer SST becomes warmer. Sea ice anomalies affected by atmospheric circulation in winter persist in the eastern part of the Barents Sea until late spring in the early period (Figures 3 and 4). Sea ice has a role in thermal insulation between the ocean and atmosphere; thus, sea ice anomalies alter the albedo and influence the amount of solar energy absorbed into the ocean during late spring, which then impacts the
Figure 3. Lag-correlation coefficients between the winter (DJF) AO index and sea ice concentration in the Northern Hemisphere in the subsequent months (February–April). Sea ice concentration in (a) February, (c) March, and (e) April from 1963 to 1989; (b) February, (d) March, and (f) April from 1990 to 2012. Shadings indicate that correlation coefficients exceed confidence levels (90%, 95%, and 99%). Contours represent the climatological sea ice concentration (purple line: 0, orange line: 0.5, and green line: 0.9).
Figure 4. As for Figure 3 but in May–July. Sea ice concentration in (a) May, (c) June, and (e) July from 1963 to 1989; (b) May, (d) June, and (f) July from 1990 to 2012.
summer SST anomaly. Thus, we speculate that sea ice anomalies convey climatic memory to SST anomalies, and the winter AO signal persists in SST anomalies over the Barents Sea until summer through the sea ice concentration variations (Figure 6).

In the recent period, a positive signal in the Barents Sea was evident in March, but almost disappeared in April. The signal reappeared slightly in the eastern part of the Barents Sea in June; however, it was not observed after July. Thus, this result suggests that the oceanic memory effects were weak in the recent period, and the winter AO signal did not persist in SST anomalies until summer. In addition, in both periods, a tripolar SST signal in the North Atlantic Ocean (positive correlation over the mid-latitudes and negative correlations over the tropics and high latitudes) was apparent during spring (Figure 5). This was also reported by Otomi et al. (2013). The negative correlation in the tropical North Atlantic remained evident until August during the recent period, whereas it almost disappeared in summer in the early period (Figure 6).

4.2. Long-Term Change of the Oceanic Memory Effects Over the Barents Sea

In the early good-linkage period, oceanic memory effects (sea ice and SST) of the Barents Sea for inter-seasonal linkage were prominent in the Arctic region. The extent of the Barents Sea is small in this area, covering only...
approximately 10% of the Arctic Ocean. However, two-way heat transport between the ocean and the atmosphere is vigorous. The Barents Sea has the most active ocean-atmosphere heat exchange in the Arctic Ocean and dominates the seasonal changes in the Arctic heat budget (Serreze et al., 2007; Smedsrud et al., 2013). In contrast, in the recent poor-linkage period, the oceanic memory effects of the Barents Sea weakened. The reason for the winter AO signal not persisting in sea ice and SST anomalies within the Barents Sea until the subsequent spring and summer in the recent period is unknown. Thus, this section focuses on the Barents Sea to understand the long-term changes in oceanic memory effects on inter-seasonal linkages.

In winter, sea ice anomalies in the Barents Sea, which are influenced by increased heat transport of warm North Atlantic water, dominate the recent decrease in sea ice in the Northern Hemisphere and play an important role in Arctic amplification and the Arctic heat budget (Smedsrud et al., 2013). Variations in sea ice in the Barents Sea are strongly associated with the Arctic climate system. Figures 7a and 7b show the climatology and interannual variance in sea ice concentration in the Arctic region in April. The Barents Sea is in a boundary region between open water and ice areas (sea ice concentration: 0.1–0.9). In addition, the interannual variance of sea ice concentration in the Barents Sea is the largest in the Arctic Ocean, which indicates that the oceanic conditions of the Barents Sea are the most variable in the Arctic Ocean. Sea ice in the Barents Sea reaches a maximum in late winter (March–April) and a minimum in late summer (August–September) owing to solar heating, and the seasonal variation is large. In addition, Figure 7c shows the difference in climatological sea ice in April between the early and recent periods. The sea ice loss in the Barents Sea in April was remarkable in the Arctic region.

Figure 6. As for Figure 5 but in June–August. SST in (a) June, (c) July, and (e) August from 1963 to 1989; (b) June, (d) July, and (f) August from 1990 to 2012.
Figures 8 and 9 show the climatology and interannual variance in sea ice concentration in the Barents Sea from March to July. Sea ice in the area where interannual variance was the largest in the early period decreased significantly in the recent period. In the early period, the interannual variance was largest in April and May. In contrast, in the recent period, the area with the largest interannual variance of sea ice concentration shifted to the north compared to that in the early period, associated with the northward displacement of the sea ice edge, and the values of the interannual variance in May have become smaller. This result indicates that in the recent period, the ice melt season started earlier, and the expansion of open water areas was remarkable in May compared to the early period.

In addition, we examined the period during which sea ice over the Barents Sea was remarkably reduced. Figure 10 shows the time-series of sea ice concentration in the Barents Sea for March, April, and May from
1958 to 2017. Focusing on the change in the 11-year running mean, sea ice in the Barents Sea decreased in recent years, and the average sea ice concentration was remarkably different before and after the late 1980s. The winter AO index changed to a strongly positive value in the late 1980s and tended to be positive until the early 1990s (Figure 1a). The positive AO phase caused a divergence in the motion of the sea ice in the Eastern Hemisphere and increased the sea ice flux out of the Arctic through the Fram Strait (Rigor et al., 2002). In the

Figure 8. Climatology (shadings) and interannual variance (contours) in sea ice concentration in the Barents Sea in March–May. (a) March, (c) April, and (e) May from 1963 to 1989; (b) March, (d) April, and (f) May from 1990 to 2012. Contour interval is 0.03.
late 1980s, larger ice outflow rates at the Fram Strait were observed with a highly positive AO phase. The area in the Arctic covered with old ice has decreased rapidly and has remained small since that time (Lindsay & Zhang, 2005; Rigor & Wallace, 2004).

To understand the change in the persistency of sea ice anomalies, the autocorrelation from January through December with sea ice concentration in the Barents Sea in February (Figure 11a) was calculated based on high-pass-filtered data. The correlation coefficients between February and the subsequent months in the early period were higher than those in the recent period until October. Thus, the persistence of sea ice anomalies in the Barents Sea during the early period was higher. Furthermore, we calculated the autocorrelation of SST in the Barents Sea in May (Figure 11b), and the persistence of SST anomalies was also higher in the early period than in the recent period. These results imply that in the early period, sea ice and SST anomalies in the Barents Sea tended to persist longer until the subsequent spring and summer.

5. Influence of Oceanic Memories on the Summer AO

After the winter AO signal is maintained in the ocean, oceanic memories may affect atmospheric circulation in the subsequent summer. To reveal the sea area that affects atmospheric circulation in summer, the lag-correlation coefficients between sea ice concentration in the Northern Hemisphere in late spring (April–May) and the AO index in the subsequent summer were calculated (Figure 12). In the early period, a negative correlation signal was observed in the Barents Sea. The negative correlation indicates that when the late spring sea ice concentration over the Barents Sea has a negative anomaly, the summer AO tends to be in a positive phase.

Figure 9. As for Figure 8 but in June–July. (a) June, (c) July from 1963 to 1989; (b) June, and (d) July from 1990 to 2012.
and vice versa. However, in the recent period, the negative signal in the Barents Sea almost disappeared, and a negative signal to the west of Greenland appeared.

The correlation coefficients between SST in the Northern Hemisphere in late spring and summer and the AO index in summer were calculated (Figure 13). In the early period, a positive correlation signal in the Barents Sea and a negative correlation signal south of Greenland were found from late spring to summer. In addition, positive correlation signals appeared in the Sea of Okhotsk and the North Sea only in summer. A positive correlation indicates that when SST is anomalously warmer, the summer AO tends to be in a positive phase. Conversely, in the negative correlation area, SST is colder in the positive AO phase. However, in the recent period, the significant correlations observed in the early period disappeared. Negative correlations were observed in the North Sea and the eastern part of the North Atlantic Ocean at mid-latitudes only in late spring. Thus, in the results of both

Figure 10. The time series of sea ice concentration in the Barents Sea from 1958 to 2017 (black dashed line). The red solid line denotes the 11-year running mean. (a) March, (b) April, and (c) May.
Auto–Correlation Coefficients (Barents Sea)

(a) SIC in Feb         (b) SST in May

Figure 11. (a) Sea ice concentration (SIC) autocorrelation coefficients from January through December with February within the Barents Sea. The red line denotes the early period (1963–1989), and the blue line denotes the recent period (1990–2012). (b) Same as (a) but for sea surface temperature (SST) in the Barents Sea in May. The dashed lines denote a 95% confidence level in each period.

(a) AM (1963–1989)         (b) AM (1990–2012)

Figure 12. Lag-correlation coefficients between sea ice concentration in the Northern Hemisphere in late spring (AM) and the summer (JJA) AO index for (a) 1963–1989 and (b) 1990–2012. Shadings indicate that the correlation coefficients exceed the confidence levels (90%, 95%, and 99%). Contours are the climatology in sea ice concentration in late spring (purple line: 0, orange line: 0.5, and green line: 0.9).
sea ice and SST (Figures 12 and 13), there were significant signals in the Barents Sea in the early period, which disappeared in the recent period.

To confirm the change in the impact of oceanic memories over the Barents Sea on atmospheric circulation, in terms of spatial pattern, the simultaneous correlation coefficients between SST in the Barents Sea and 500-hPa geopotential height in the Northern Hemisphere were calculated on a monthly basis (May–August) (Figure 14). In the early period, patchy correlations were observed in May, whereas the negative correlation signal in the Arctic Ocean was surrounded by positive correlation signals over the south of the Barents Sea, the Sea of Okhotsk, and Canada in June and July, which was similar to the summer AO pattern. In August, the negative signals were divided into eastern Siberia and Greenland. Thus, the north-south seesaw pattern was broken. These results suggest that the warm signal in the Barents Sea contributes to the formation of the positive AO condition in June–July and vice versa. Conversely, in the recent period, the seesaw pattern did not appear in May–August, which implies that the connection between the surface anomalies in the Barents Sea and the summer AO disappeared.

In addition, Figure 15 shows the simultaneous correlation coefficients between SST in the Barents Sea and the 300-hPa zonal wind in the Northern Hemisphere on a monthly basis (May–August). In the early period, positive correlation signals were observed around the Arctic Ocean in June and July. The positive correlation indicates that when SST in the Barents Sea is anomalously warmer, the westerly winds surrounding the Arctic Ocean are strong. This feature is consistent with the double-jet stream structure of the positive summer AO. Ogi, Yamazaki, and Tachibana (2004) showed that in the positive summer AO phase, temperature gradients around the Arctic Ocean are large, and the Arctic frontal zone is active; however, these features are absent in the negative summer AO phase. Thus, the large temperature gradients enhance the baroclinicity and relate to strong westerly winds due to the thermal wind relationship. The SST anomalies in the Barents Sea have the potential to impact temperature gradients because the heat transport between the ocean and atmosphere is vigorous there. Compared to Figures 14c and 14e, the negative signal of the 500-hPa geopotential height in the Arctic in June–July in the early period was surrounded by positive westerly wind signals. In the positive summer AO phase, Arctic cold air is trapped by strong westerly winds around the Arctic Ocean and accumulates in the Arctic, which can cause a negative 500-hPa geopotential height anomaly in the Arctic. Thus,
Figure 14. Simultaneous correlation coefficients between sea surface temperature (SST) in the Barents Sea and 500-hPa geopotential height in the Northern Hemisphere for the early (1963–1989) and recent (1990–2012) periods. (a) May, (c) June, (e) July, and (g) August from 1963 to 1989; (b) May, (d) June, (f) July, and (h) August from 1990 to 2012. Shadings indicate that correlation coefficients exceed confidence levels (90%, 95%, and 99%).
Figure 15. Simultaneous correlation coefficients between sea surface temperature (SST) in the Barents Sea and 300-hPa zonal wind in the Northern Hemisphere for the early (1963–1989) and recent (1990–2012) periods. (a) May, (c) June, (e) July, and (g) August from 1963 to 1989; (b) May, (d) June, (f) July, and (h) August from 1990 to 2012. Shadings indicate that correlation coefficients exceed confidence levels (90%, 95%, and 99%).
we consider that the relationship between SST in the Barents Sea and the polar jet was important for the variations in the summer AO in the early period. In contrast, in the recent period, the significant correlation signals surrounding the Arctic Ocean did not appear in May–August, that is, the relationship between SST in the Barents Sea and the polar jet was unclear. Outten and Esau (2012) indicated that recent Arctic warming with a decrease in sea ice caused a weakening of the meridional temperature gradients. This may be related to the change in the relationship between SST in the Barents Sea and the polar jet; thus, further investigation is needed.

Ogi, Tachibana, and Yamazaki (2004) and Tachibana et al. (2004) analyzed the wave activity flux (Takaya & Nakamura, 2001) to reveal the source region of the Rossby waves, which can contribute to the formation of the Okhotsk High. As a result, a large emanation of the wave activity flux at 300 hPa was observed around the southern Barents Sea, and the Rossby wave propagated eastward from the Barents Sea toward the Sea of Okhotsk region. They suggested that warm signals around the Barents Sea excite the Rossby wave, and the Barents Sea can be a forcing area for anomalous summer atmospheric circulation. Ogi, Yamazaki, and Tachibana (2004) also suggested that in the positive summer AO phase, strong westerly winds around the Arctic Ocean act as a waveguide for the Rossby wave, which contributes to the formation of wave trains. In the early period, the positive winter AO signal persisted in SST anomalies over the Barents Sea until the subsequent summer through the negative sea ice concentration anomalies (Figures 3–6), and it is likely that the positive SST anomalies in June and July generated Rossby waves and contributed to the formation of the positive summer AO pattern. Therefore, the Barents Sea played an important role in the connection between the winter AO and the subsequent summer AO in the early period; however, the Barents Sea may have recently lost this role.

6. Summary and Discussion

This study investigated the long-term changes in the inter-seasonal linkage from the winter AO to the subsequent summer AO in the mid- and high latitudes in the Northern Hemisphere using JRA-55, sea ice concentration, and SST observations. The winter AO and the summer AO had lagged positive correlations prior to the late 1980s; however, the inter-seasonal linkage collapsed after the late 1980s. The results suggest that the change in oceanic memory effects over the Barents Sea is closely related to the collapse of the linkage.

In the early good-linkage period, the winter AO affected the sea ice variability within the Barents Sea and sea ice anomalies persisted until late spring. This sea ice anomaly influenced the amount of solar energy absorbed into the ocean in late spring, which then impacted the summer SST anomaly. For example, when the winter AO is positive, the negative sea ice anomaly contributes to more absorption of the solar energy during late spring, which then causes a positive summer SST anomaly; thus, the winter AO signal persisted in SST anomalies over the Barents Sea until summer through the sea ice concentration variations. The oceanic memories then influenced atmospheric circulation in summer. Therefore, the Barents Sea can convey climatic memories from winter to summer. The oceanic memory effects of the Barents Sea may have played an important role in the inter-seasonal linkage.

In contrast, in the recent poor-linkage period, the winter AO signal did not persist throughout the seasons in the Barents Sea because the persistence of sea ice and SST anomalies were less significant. Oceanic memory effects were weakened. In the late 1980s, sea ice loss was remarkable in the Barents Sea (Figure 10). Lindsay and Zhang (2005) hypothesized that the flushing of old, thick ice through the Fram Strait in the late 1980s and the early 1990s was a tipping point when the Arctic ice-ocean system entered a new regime of thinning ice and increasing summer open water due to ice-albedo feedback. The ice melt season started earlier, and the expansion of open water areas was remarkable compared to the early period in the Barents Sea, particularly in May (Figures 8 and 9). Sea ice anomalies affected by the winter AO did not persist until late spring because there was not as much sea ice in the later period due to more open water areas. Therefore, the process by which sea ice anomalies impact SST in summer and convey climatic memory to SST anomalies was unclear in the recent period. We concluded that the Barents Sea may have lost its role in inter-seasonal linkage.

However, not only the oceanic memory effects over the Barents Sea, but also other phenomena have the potential to contribute to variations in summer AO. For example, Otomi et al. (2013) showed that the tripolar SST
anomaly pattern in the North Atlantic Ocean might be related to the AO polarity reversal from winter to the subsequent summer, focusing on the unusually hot summer of 2010. The AO index in the 2009/2010 winter was record-breaking negative with unusually cold weather over most parts of Eurasia (Cohen et al., 2010), whereas the AO index in the subsequent summer was anomalously positive. The tripolar SST anomaly pattern in the North Atlantic Ocean (warm SST anomalies over the tropics and high latitudes and cold SST anomalies over the mid-latitudes), due to the influence of the negative winter AO phase, persisted in the subsequent summer because of the large oceanic heat capacity (Otomi et al., 2013). The warm SST anomaly over the tropical North Atlantic caused a remote blocking high over Europe (Cassou et al., 2005). The blocking high induced another blocking high over the Russian Far East with eastward propagation of Rossby waves (Tachibana et al., 2010), which is similar to the pattern of the positive summer AO. However, the above process of AO polarity reversal may be specific to the case of 2010 or the process may only be seen in recent years because the SST anomaly over the tropical North Atlantic persisted until summer only during the recent period after the winter AO signal was memorized there (Figure 6). Otomi et al. (2013) only focused on the case in 2010, and therefore, whether their processes can sometimes be observed in other years has not yet been fully investigated.

In contrast, the possibility that oceanic memory effects over the Barents Sea play an important role in the connection between the winter AO and the subsequent summer AO is significant. Figure 16a shows the 21-year sliding lag-correlation coefficients between the AO index in winter and sea ice concentration in the Barents Sea in the subsequent late spring. The lag relationship between the winter AO and sea ice in the Barents Sea in late spring was significantly negative before the late 1980s, while the correlation quickly weakened after the late 1980s. The results are shown in Figures 3 and 4. In addition, Figure 16b shows the 21-year sliding lag-correlation coefficients between the AO index in winter and that in the subsequent summer, which is nearly identical to those shown in Figure 1b of Yamazaki et al. (2019). The inter-seasonal linkage between the winter AO and the subsequent summer AO was significantly positive before the late 1980s, while the correlation quickly weakened after the late 1980s. Comparing Figures 16a and 16b, the inter-seasonal linkage between the winter AO and the subsequent summer AO was synchronously broken in association with the weakening of the function that the winter AO signal persists in the sea ice anomaly over the Barents Sea until late spring. Figure 16c shows the 21-year sliding lag-correlation coefficients between the AO index in winter and SST in the tropical North Atlantic in the subsequent summer. The negative correlation between the winter AO and the summer tropical North Atlantic SST has been significant only in recent years. Therefore, the relationship between the tropical North Atlantic SST and the collapse of the inter-seasonal linkage in the late 1980s was not clear from this result.

Sea ice in the Barents Sea has rapidly decreased in recent years, which is related to the weakening of the oceanic memory effects and the collapse of the inter-seasonal linkage between the winter AO and the subsequent summer AO. However, the analyses in this study confirmed a correlation, but not causation. In the future, climate model experiments are required to investigate the influence of the Barents Sea on inter-seasonal linkage and its collapse. If Arctic warming with a decrease in sea ice in the Northern Hemisphere continues, it is unknown whether the inter-seasonal linkage will remain weak or will be reestablished in the future. Because an anomalous positive summer AO phase causes unusual weather on a planetary scale and significantly impacts people, understanding the inter-seasonal linkage may be useful for the long-term forecasting of unusual summer weather. Thus, it is essential to continue investigations of the inter-seasonal linkage.

Appendix A: The Long-Term Change of the North-South Seesaw Pattern in Terms of Spatial Distribution

To confirm the long-term change in the north-south seesaw pattern between the winter AO and atmospheric circulation in summer, in terms of spatial distribution, we also calculated the lag-correlation coefficients between the winter AO index and the 500-hPa geopotential height in the Northern Hemisphere in summer for 20 years, using a 5-year sliding correlation. During the first half of the period (Figures A1a–A1c), a seesaw-like pattern of positive correlation in the mid-latitudes, such as Europe, western Siberia, and Canada, and negative correlation in the Arctic was observed. This represents the inter-seasonal linkage between the winter AO and the subsequent summer AO. In contrast, after 1978–1997 (the center of the period was in the late 1980s), the negative signal in
Figure 16. (a) 21-year sliding lag-correlation coefficients between the winter (DJF) AO index and sea ice concentration (SIC) in the Barents Sea in late spring (AM). (b) 21-year sliding lag-correlation coefficients between the winter AO index and the summer (JJA) AO index. (c) 21-year sliding lag-correlation coefficients between the winter AO index and sea surface temperature (SST) in the tropical North Atlantic in summer. Years represent the central year of the 21-year window, for example, the correlation coefficient from 1980 to 2000 is plotted at 1990 on the x-axis. The dashed lines denote a 90% confidence level.

The Arctic disappeared, and the seesaw-like pattern was unclear (Figures A1d–A1g). Thus, the results suggest that the inter-seasonal linkage between the winter AO and atmospheric circulation in the subsequent summer changed in the late 1980s.
Data Availability Statement

The data sets can be accessed using the following links: JRA-55 (https://jra.kishou.go.jp/JRA-55/), SV NAM

Figure A1. Lag-correlation coefficients between the winter (DJF) AO index and the 500-hPa geopotential height in the Northern Hemisphere in summer (JJA) for 20 years, using a sliding correlation by 5 years. (a) 1963–1982, (b) 1968–1987, (c) 1973–1992, (d) 1978–1997, (e) 1983–2002, (f) 1988–2007, and (g) 1993–2012. The shadings indicate that the correlation coefficients exceed the confidence levels (90%, 95%, and 99%), as shown in the legend.
Acknowledgments

This study was partly supported by JSPS KAKENHI Grants 16K16349 and 20H01389, and a research grant from the Japan Geographic Data Center. The authors also thank Dr. Nozomi Kamizawa for her technical support. The grid analysis and display system (GrADS) was used to draw figures.

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