Rapid Arctic Sea Ice Loss on the Synoptic Time Scale and Related Atmospheric Circulation Anomalies

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(Manuscript received 15 July 2019, in final form 30 October 2019)

ABSTRACT

Large sea ice loss on the synoptic time scale is examined in various subregions in the Arctic as well as at the pan-Arctic scale. It is found that the frequency of large daily sea ice loss (LDSIL) days is significantly correlated with the September sea ice extent over the Beaufort–Chukchi–Siberian Seas, the Laptev–Kara Seas, the central Arctic, and the all-Arctic regions, indicating a link between the synoptic sea ice variability and the interannual variability of the annual minimum sea ice extent. A composite analysis reveals dipoles of anomalous cyclones and anticyclones associated with LDSIL days. Different from the well-known Arctic dipole pattern, the east–west dipoles are found over the corresponding regions of LDSIL in the Arctic marginal seas and are associated with the increasing occurrence of Rossby wave breaking and atmospheric rivers. The anticyclones of the dipoles are persistent and quasi-stationary, reminiscent of blocking. The anomalous poleward flow between the cyclone and the anticyclone enhances the poleward transport of heat and water vapor in the lower troposphere. Although enhanced downward shortwave radiation, associated with reduced cloud fraction, is found in some regions, it is not collocated with the regions of LDSIL. In contrast, enhanced downward longwave radiation owing to increasing column water vapor shows good spatial correspondence with LDSIL, indicating the importance of atmospheric rivers in LDSIL events. Lead/lag composites with respect to the onset of LDSIL episodes reveal precursor wave trains spanning the midlatitudes. The wave trains have predominantly zonal energy propagation in the midlatitudes and do not show a clear link to tropical or subtropical forcing.

1. Introduction

The recent decrease of Arctic sea ice coverage is one of the most striking indicators of global environmental change. The Arctic sea ice extent in September, as assessed from satellite observations, has changed

significantly, with the pan-Arctic extent in each of the past 13 Septembers (2007–19) all lower than in any years of the earlier satellite era (1979–2006; NSIDC 2018). The Arctic is expected to become essentially ice-free during summer by about midcentury (Notz and Stroeve 2016; Wang and Overland 2009, 2012). This ice loss has profound impacts on human activities and ecosystems in the Arctic (Post et al. 2013; Bhatt et al. 2010) and the global energy budget and hydrological cycle (Screen and Simmonds 2010; Liu et al. 2012; Bintanja and Selten 2014; Kopec et al. 2016). There is an urgent need to understand and predict sea

Supplemental information related to this paper is available at the Journals Online website: <https://doi.org/10.1175/JCLI-D-19-0528.s1>.

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DOI: 10.1175/JCLI-D-19-0528.1

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ice variations at the pan-Arctic and regional scales (Eicken 2013).

On the seasonal to interannual time scales, forecasts of September pan-Arctic sea ice extent have shown some skill at one to three months lead time (Stroeve et al. 2014), yet the improvement over anomaly persistence is modest, especially when the trend is removed (Walsh et al. 2019) or when the seasonal anomalies deviate from the long-term trend (Stroeve et al. 2014). Atmospheric predictors through teleconnection patterns have been established in statistical studies with some success (Drobot 2003; Lindsay et al. 2008). Additionally, ocean temperature initialization also contributes to skillful seasonal forecasts of sea ice, especially in the North Atlantic sub-Arctic seas (Bushuk et al. 2017; Blanchard-Wrigglesworth et al.'s 2011). However, accurate and useful predictions remain challenging due to limited knowledge of sea ice processes, imperfect data on the drivers of sea ice variability, incomplete observations of the current and past sea ice state, and inherited errors in current prediction systems (e.g., deficiencies of model physics).

On the interannual to decadal time scales, the decrease of Arctic sea ice is characterized by years of extreme ice loss, often followed by a year or two in which the sea ice extent increases but does not return to its prior level. In the post-2000 era, 2007 and 2012 stand out as such years, as do 1985, 1990, and 1995. Holland et al. (2008) examined rapid ice loss events (RILEs), which were defined as periods when the loss of 5-yr running mean September sea ice extent (SIE) exceeded 0.5 million km^2 . These events, defined also by various similar criteria, have been addressed further in the context of interannual-to-decadal changes (e.g., Döscher and Koenigk 2013; Paquin et al. 2013; Rogers et al. 2015). RILE years have accounted for most of the reduction of Arctic sea ice extent over the past several decades, and model simulations suggest that they will continue to do so in the future (Holland et al. 2006).

Much less attention has been paid to sea ice variability and extreme losses on shorter time scales (i.e., several days), and past studies on these time scales have generally focused on storm-related events. In contrast to the seasonal to decadal prediction, short-range sea ice prediction is more sensitive to the atmospheric initial conditions, particularly surface winds and temperatures (Mohammadi-Aragh et al. 2018). The thinning of Arctic sea ice in recent years can make the ice cover more vulnerable to the wind forcing and associated ocean mixing (Kwok 2018). For example, the record low sea ice extent in September 2012 (lower than any other year on record by at least $0.67 \text{ million km}^2$) has been attributed partially to the occurrence of a strong Arctic

cyclone in August 2012 (Parkinson and Comiso 2013). While the 2012 Arctic cyclone was indeed extreme (Simmonds and Rudeva 2012), another study (Zhang et al. 2013) concluded that the cyclone accounted for only $0.15 \text{ million km}^2$ of sea ice loss. In addition, Morello (2013) noted that Arctic cyclones can actually reduce ice loss if they increase snow cover during the summer season when surface albedo exerts considerable leverage on the loss of sea ice. Overall, the variability and predictability of Arctic sea ice on the weather time scale are less well studied and arguably less understood than the seasonal and longer time scales.

The present study attempts to place the extreme multiday losses of sea ice into a broader framework of atmospheric forcing, with a goal toward improved understanding and prediction on the time scale of several days to several weeks. We refer to rapid decreases of sea ice extent on the synoptic time scale as large daily sea ice loss (LDSIL) events in order to distinguish them from the interannual RILEs of Holland et al. (2008) and others. Given the thinning and increased vulnerability of sea ice, such LDSIL events may become more common. It is also reasonable to expect that these shorter-term LDSIL events occur more frequently in years that ultimately manifest as RILE years. In other words, the shorter-term LDSIL events may be a key feature of the RILE years. Additionally, skillful short-term sea ice prediction is of significant value for maritime operations. Given that numerical weather forecasts have become skillful at the range of a week to possibly two weeks, the prospects for improved prediction of LDSIL events may be more imminent than that for the prediction of RILE years.

In the present study, we focus on two primary questions: (i) How does short-term sea ice loss relate to the annual minimum SIE (i.e., September SIE)? (ii) What large-scale atmospheric circulation anomalies and their associated physical processes contribute to LDSIL events? We examine sea ice variability and LDSIL on the regional as well as the pan-Arctic scales. To the extent that the atmospheric circulation provides dynamic (advection, deformation) and thermodynamic (melt) forcing of sea ice loss, the atmospheric signatures of LDSIL events can be expected to vary in different sub-Arctic and pan-Arctic regions. Moreover, the response of the atmospheric circulation to sea ice loss is likely sensitive to the location of sea ice loss (Pedersen et al. 2016), further highlighting the need to understand regional sea ice variability. Although our focus is on the short time scales (several days), the investigation of the abovementioned two questions will help advance our understanding of the variability and predictability of Arctic sea ice on the longer time scales.

FIG. 1. Different subregions of the Arctic as defined for the Multi-Sensor Arctic Sea Ice Extent (MASIE) product (https://nsidc.org/data/masie/browse_regions). The six regions examined in this study are the Beaufort–Chukchi–East Siberian Seas (i.e., BeaChSi, 1 + 2 + 3), the Kara–Laptev Seas (LapKar, 4 + 5), the Barents Sea (Barents, 6), the Greenland Sea (Grnland, 7), the central Arctic (CentArc, 11), and the all-Arctic (AllArct, 1 + 2 + 3 + 4 + 5 + 6 + 7 + 8 + 11).

The remainder of the paper is organized as follows. Data and methodology are described in [section 2](#). The season of interest is described in [section 3](#). The link between LDSIL and the September SIE is examined in [section 4](#), and the atmospheric processes associated with LDSIL are investigated in [section 5](#), followed by a summary in [section 6](#).

2. Data and methodology

a. Sea ice data and preprocessing

The primary data source is the daily gridded sea ice concentration during 1979–2017 from the National Snow and Ice Data Center (dataset NSIDC-0051, <https://nsidc.org/data/NSIDC-0051>), derived from passive microwave retrievals at a spatial resolution of 25 km. The Arctic domain is divided into 16 regions following the Multisensor Analyzed Sea Ice Extent regionalization ([Fig. 1](#)). Daily SIE for each region is computed as the cumulative area of all grid cells with a sea ice concentration of at least 15% following the conventional definition of ice extent (e.g., [Parkinson and Cavalieri 2012](#)).

Several regions are consolidated based on their geographical proximity, seasonal cycle similarity, and the strong correlations between their interannual variations (see Table S1 in the online supplemental material): the Beaufort–Chukchi–East Siberian Seas (regions 1–3), and Laptev–Kara Seas (regions 4–5). An “all-Arctic” area is defined consisting of regions 1–8 and 11. In addition to the combined regions, the Barents Sea (region 6), the Greenland Sea (region 7), and the central Arctic (region 11) are also examined. Several regions (10 = Hudson Bay, 12 = Bering Sea, 13 = Baltic Sea, 14 = Sea of Okhotsk, 15 = Yellow Sea, 16 = Gulf of Alaska) were omitted in this study because they are geographically separated from the main Arctic sea ice pack. We also omitted region 9, the Canadian Archipelago, because it consists mainly of straits and channels that constrain the ice so that its response (at least dynamically) to atmospheric forcing is quite different from the open Arctic and sub-Arctic seas. By contrast, region 8 (Baffin Bay and the Labrador Sea), while largely separated from the main ice pack of the central Arctic, generally has a marginal ice zone or an ice edge that responds to dynamic and thermodynamic forcing from the

FIG. 2. (a) The response function of the 3- (green), 5- (red), and 7-day (yellow) low-pass filters; (b) the raw time series of the CentArc SIE (black squares; 10^6 km^2) in June 1979 and the time series after the application of the 3- (green), 5- (red), and 7-day (yellow) low-pass filters.

atmosphere. We therefore include region 8 in the all-Arctic total ice extent but do not present results separately for it.

Sea ice data derived from satellite retrievals are subject to artificial, high-frequency fluctuations due to changes in cloud patterns (Comiso et al. 2017). To mitigate this impact, a low-pass filter (Doblas-Reyes and Deque 1998) is applied to the daily SIE time series. We tested the 3-, 5- and 7-day low-pass filters (Fig. 2a), and the 5-day low-pass filter was adopted because it smooths out high-frequency fluctuations of SIE but does not dampen the multiday variability as much as the 7-day low-pass filter (Fig. 2b). In addition, the SIE data are available every other day before July 1987 and available every day afterward. The low-pass filter effectively fills the missing data before July 1987.

To focus on the synoptic-scale SIE variability (i.e., from several days to two weeks), the linear trend and seasonal cycle are both removed from the 5-day low-pass filtered data. The linear trend and seasonal cycle are defined for the time period of 1979–2017 for each calendar day using the 31-day running mean daily data, which helps avoid discontinuities across calendar months. As shown in Fig. 3, seasonal cycles are apparent in all regions, with minima in September. The September SIE varies from 0.1 to 0.2 million km^2 in the Greenland and Barents Seas to more than 3 million km^2 in the central Arctic, and the all-Arctic SIE is between 5 and 6 million km^2 in September. Strong seasonality and regional differences are also found in the interannual variability of SIE. The interannual variability, measured by standard deviation, is greatest during the winter in the Greenland and Barents Seas and during the late summer in the other regions. The small summer variability in the Greenland and Barents

Seas is a consequence of the low SIE in these regions in late summer, while the winter variability is small in the other regions (Beaufort–Chukchi–East Siberian, Laptev–Kara) owing to their nearly complete winter sea ice coverage. Finally, Fig. 3 shows that all regions have experienced negative trends of SIE. The trend has a similar seasonality as the interannual variability: the negative trend is strongest in winter in the Greenland and Barents Seas and is strongest in summer in the other regions; the trend is weak when a region is nearly ice-free or has nearly complete sea ice coverage. All the subsequent sea ice analyses are based on the detrended 5-day low-pass filtered SIE data with the seasonal cycle removed except for the monthly mean SIE data used in Fig. 6, which are derived from the raw daily SIE data.

b. Atmospheric variables

The atmospheric data are obtained from the 6-hourly ERA-Interim reanalysis (<https://www.ecmwf.int/en/forecasts/datasets/reanalysis-datasets/era-interim>). To reduce the data volume and to have a temporal frequency consistent with the SIE data, we subsample the reanalysis data at daily intervals by using the 0000 UTC fields only. Additionally, longwave and shortwave radiative fluxes are derived from the accumulated values from the ERA-Interim 12-h forecasts (Graversen et al. 2011; Lee et al. 2017). To account for the diurnal cycle, the daily mean radiative fluxes are computed by averaging the 12-h forecasts initialized at 0000 and 1200 UTC. As with the SIE data, the seasonal cycle is removed from all the atmospheric variables, but no detrending or low-pass filtering is applied as the linear trend of the atmospheric circulation is much weaker than that of the Arctic sea ice.

Mean, Standard Deviation and Linear Trend of SIE (1979–2017)

FIG. 3. The mean (green curves; 10^6 km^2) ± 1 standard deviation (shading) and the linear trend (orange curves; $10^6 \text{ km}^2 \text{ decade}^{-1}$) of SIE as a function of calendar days. The left ordinate is for the mean and standard deviation, and the ordinate of the linear trend is shown in orange on the right.

Recent studies suggested that atmospheric rivers (ARs) play an important role in poleward water vapor transport into the Arctic, which may subsequently cause sea ice loss (Yang and Magnusdottir 2017; Gong et al. 2017; Mattingly et al. 2018; Hegyi and Taylor 2018). Atmospheric rivers are narrow corridors of intense horizontal water vapor transport in the atmosphere. Although covering only about 10% of Earth's surface at a given latitude, they contribute to more than 90% meridional water vapor transport in the midlatitudes (Zhu and Newell 1998; Ralph et al. 2004). Here we employ the global AR dataset developed by Guan and Waliser (2015), which provides the daily frequency of ARs at $1.5^\circ \times 1.5^\circ$ resolution derived from the ERA-Interim reanalysis.

ARs are often related to Rossby wave breaking (RWB; Payne and Magnusdottir 2014; Hu et al. 2017). RWB is characterized by irreversible overturning of potential vorticity (PV) contours, and it occurs when the weak background PV gradient is not sufficient to support the linear propagation of large-amplitude waves or when the wave phase speed approaches the mean flow speed. RWB is marked by enhanced turbulent mixing in the vicinity of the breaking wave and by PV and moisture filaments that may extend a long distance from the location of RWB. In our study, the frequencies of cyclonic RWB and anticyclonic RWB (distinguished by the orientation of the PV overturning) are derived separately using the 350-K isentropic PV from the ERA-

Interim reanalysis following the algorithm developed by Strong and Magnusdottir (2008). The location of an RWB event is determined by the centroid of the high PV tongue, and a grid is marked with RWB occurrence if a high-PV tongue associated with RWB covers that grid point.

Additionally, the wave activity flux analysis is employed to estimate the energy propagation of Rossby waves. We follow the formulation by Takaya and Nakamura (2001), which can be applied to either stationary or migratory quasigeostrophic eddies on a zonally varying basic flow. The wave activity flux vector is independent of the wave phase and is parallel to the local group velocity of wave perturbations in the WKB limit.

3. Season of interest

Although large short-term sea ice loss can occur in all seasons, we focus on the summer season for its potentially strong link to the annual minimum SIE in September. The autocorrelations of monthly mean SIE in September and those in preceding months provide an indication of the time scales of the seasonal variations (Table 1). Significant autocorrelations (above the 95% confidence level) are found as far back as in June in most regions, but do not extend further back beyond June except in the Laptev–Kara Seas. The substantial decrease in autocorrelations from June to May is consistent

TABLE 1. Correlation coefficients between the September SIE and SIE in previous months during 1979–2017. The correlation coefficients exceeding the 99% confidence level are in bold italic, and the correlation coefficients exceeding the 95% confidence level are in italic.

Corr	BeaChSi	LapKara	Barents	Grnland	CentArc	AllArct
January	0.00	0.00	<i>0.40</i>	0.15	0.25	0.23
February	0.11	0.17	0.27	0.18	<i>0.42</i>	<i>0.44</i>
March	0.09	0.08	0.21	−0.05	<i>0.43</i>	0.17
April	0.11	<i>0.39</i>	0.22	−0.09	−0.11	0.22
May	0.28	<i>0.45</i>	0.17	0.08	−0.06	0.23
June	<i>0.53</i>	<i>0.49</i>	<i>0.37</i>	<i>0.35</i>	−0.01	<i>0.51</i>
July	<i>0.69</i>	<i>0.70</i>	<i>0.65</i>	<i>0.62</i>	0.38	<i>0.78</i>
August	<i>0.91</i>	<i>0.89</i>	<i>0.93</i>	<i>0.92</i>	<i>0.80</i>	<i>0.93</i>

with the “springtime predictability barrier” (Lindsay et al. 2008; Day et al. 2014; Bushuk et al. 2018). Additionally, significant correlations are found in January or February in the Barents Sea, the central Arctic and the all-Arctic regions, possibly related to the sea ice growth-to-melt re-emergence (Blanchard-Wrigglesworth et al. 2011; Bushuk and Giannakis 2015).

A more relevant metric for the present study is the correlation between September SIE and the antecedent month-to-month changes in SIE in the same region. As shown in Table 2, the month-to-month SIE changes correlate significantly with September SIE in most months from May onward in most regions. An exception is the Barents Sea, which is generally ice-free in September (Fig. 3). The strong correlations suggest that the low SIE in September is owing to the accumulative sea ice loss in preceding summer months, but the SIE changes prior to May generally do not correlate significantly with September SIE except for some occasional negative values. It is possible that the negative, significant correlations in late winter and spring are related to late-forming, thin sea ice prone to rapid ice cover reduction. However, it is also possible that these negative correlations are simply due to sampling, in which case the correlation tables indicate that the variability of SIE from June onward is the prime contributor to late-summer sea

ice extent (sea ice thickness, which may have a longer persistence across seasons, is not considered here). We will thus focus on the extended summer season from June to September. Restricting the analysis to one season also helps detect robust atmospheric circulation anomalies, which may vary from season to season.

4. Large daily sea ice loss and its link to September sea ice extent

The 5-day sea ice change ($dSIE$) is calculated as a metric of synoptic sea ice variability. The $dSIE$ on a certain day n is defined as the difference in SIE between two days later ($n + 2$) and two days before ($n - 2$): $dSIE(n) = SIE(n + 2) - SIE(n - 2)$. The histograms of $dSIE$ during June–September 1979–2017 are plotted for the six regions on the same scale of areal loss or gain in Fig. 4. While the symmetry is apparent in all the regions, the width of the distributions varies substantially. In the central Arctic, which is nearly completely covered with ice concentration of more than 15% for much of the summer season (Fig. 3), 5-day changes of zero (or close to zero) dominate the distribution. Similarly, the Barents and Greenland Seas, which are ice-free for much of the summer season, also have relatively narrow distributions. The Beaufort–Chukchi–Siberian and

TABLE 2. Correlation coefficients between the September SIE and the monthly SIE change during 1979–2017. The correlation coefficients exceeding the 99% confidence level are in bold italic, and the correlation coefficients exceeding the 95% confidence level are in italic.

Corr	BeaChSi	LapKara	Barents	Grnland	CentArc	AllArct
February–January	0.01	0.20	−0.19	−0.02	0.06	<i>0.43</i>
March–February	−0.10	−0.17	−0.12	<i>−0.41</i>	−0.13	<i>−0.53</i>
April–March	0.11	<i>0.33</i>	0.08	−0.05	<i>−0.42</i>	0.06
May–April	0.28	0.26	−0.02	0.23	0.03	0.05
June–May	<i>0.54</i>	<i>0.49</i>	0.23	<i>0.44</i>	0.09	<i>0.34</i>
July–June	<i>0.64</i>	<i>0.69</i>	0.11	<i>0.53</i>	<i>0.49</i>	<i>0.65</i>
August–July	<i>0.84</i>	<i>0.43</i>	−0.22	<i>0.64</i>	<i>0.81</i>	<i>0.65</i>
September–August	<i>0.42</i>	<i>0.47</i>	0.30	0.25	<i>0.90</i>	<i>0.60</i>

Histogram(%) of SIE Change: 1979–2017

FIG. 4. Histogram of sea ice extent change ($dSIE$; 10^6 km^2 per 5 days) in different regions.

Laptev–Kara Seas show a considerable range, with tails extending beyond ± 0.1 million km^2 per five days. Interestingly, these two regions show slight skewness, with large negative values more common than large positive values.¹ Finally, the all-Arctic distribution is the broadest of all, with some 5-day changes exceeding 0.3 million km^2 .

An LDSIL day is defined as a day when $dSIE$ falls below the 5th percentile of the $dSIE$ distribution in that region; the exact value of the $dSIE$ threshold varies from region to region (Fig. 4). The relevance of LDSIL to extreme sea ice days is illustrated by the evolution of SIE in the summer of 2012, a year of record-low Arctic SIE (Fig. 5). Also shown for comparison are the long-term mean (close to zero after detrending and removal of the seasonal cycle) and ± 1 standard deviation. For the four illustrated regions, the interannual standard deviation of SIE generally increases as the summer progresses. SIE is well outside the normal range in summer 2012, with negative anomalies exceeding two standard deviations in each of the four regions at certain times of the season. The extreme low SIE days,

when SIE reaches a minimum more than one standard deviation below the mean, are highlighted in red. Some interesting features of SIE variability can be learned from this illustrative example. First, sea ice loss is not a gradual process throughout the season after the removal of the seasonal cycle. Instead, it is characterized by abrupt, large losses over short time periods. Second, many extreme low SIE days are preceded by LDSIL days. This suggests that extreme low anomalies may be the cumulative effects of repeated LDSIL, especially when LDSIL days occur in succession. In fact, multiday durations of LDSIL events are more common than single-day durations (see Fig. S1). Durations of 4–6 days are not unusual, although the most common ones are 2–3 days.

The relation between the LDSIL frequency and the annual minimum SIE (i.e., September SIE) is summarized in Fig. 6. The summer seasonal frequency of LDSIL is defined as the number of LDSIL days during June–September in a year. Although derived from detrended SIE data, the LDSIL frequency shows a positive trend in the Beaufort–Chukchi–Siberian Seas and in the central Arctic, where the ice edge has been located for longer periods in recent years due to large sea ice retreat. By contrast, a negative trend is discernible in the LDSIL frequency over the Barents and Greenland Seas, consistent with their longer periods of ice-free conditions in recent years (note the different ranges of SIE between different regions).

¹ Since the seasonal cycle and linear trend have been removed from the SIE data, $dSIE = 0$ in the histogram represents the average SIE change related to the seasonal cycle and the linear trend (i.e., sea ice loss in most of the summer), instead of no change in SIE. The negative tails represent sea ice loss much more rapid than the seasonal cycle.

FIG. 5. Time series of detrended SIE anomalies (green curves; 10^6 km^2) in summer 2012 over four regions. The blue and red circles denote LDSIL days and minimum SIE days, respectively (see the definitions in the text). The long-term mean ± 1 standard deviation is shown in black curves enclosed by gray shading.

Strong negative correlations above the 99% confidence level are found between the LDSIL frequency and the detrended September SIE (orange lines) in all regions except the Barents and the Greenland Seas. In the Barents Sea, the absence of September sea ice in recent years makes the correlations essentially meaningless. The corresponding correlation for the nondetrended September SIE (in green) remains significant for the Beaufort–Chukchi–Siberian, the central Arctic and the all-Arctic seas, and the correlation is especially strong (-0.93) for the central Arctic. It is noteworthy that September 2007 and 2012, both having record-breaking low SIE, are characterized by the frequent occurrence of LDSIL during June–September in the Beaufort–Chukchi–Siberian Seas, the central Arctic and the all-Arctic seas. The strong correlations indicate that synoptic SIE variability contributes significantly to the annual minimum SIE and highlight the connection of sea ice variability between the weather and climate time scales.

5. Atmospheric circulation anomalies related to LDSIL

a. Composites of daily anomalies

To examine the atmospheric circulation anomalies associated with LDSIL, the composites of daily anomalies

of 500-hPa geopotential height (H500), AR frequency, and vertically integrated water vapor flux (IVF) from the surface to 300 hPa are constructed for all LDSIL days for the six regions described earlier. In the Arctic marginal seas (Figs. 7a–d), a dipole of H500 anomalies is found. The dipole is anchored over the region experiencing LDSIL, with a cyclone and an anticyclone to the west and east, respectively. The dipole has an equivalent barotropic structure extending throughout the troposphere (not shown). The geopotential gradient between the anomalous cyclone and anticyclone implies an anomalous poleward flow, which transports warm, moist air into the region of large sea ice loss. The IVF vectors indeed show strong poleward transport of water vapor between the anomalous cyclone and anticyclone, associated with an increase in AR occurrence (Figs. 7a–d).

The enhanced poleward water vapor transport in the vicinity of large SIE loss in Figs. 7a–d is associated with a decrease in poleward water vapor transport and AR occurrence in other regions. Therefore, the anomalies of the zonal mean poleward water vapor transport into the Arctic are not necessarily large. Given the impacts of poleward heat and water vapor transport on SIE (see more discussion in section 5b), it implies that sea ice melt may be reduced in other regions while it is enhanced between the dipole centers. This opposition

SIE (Sep) and LDSIL days (JJAS)

FIG. 6. September SIE (curves; 10^6 km^2) and the LDSIL frequency (black bars; days per season) during 1979–2017. The orange and green curves represent the detrended and nondetrended September SIE, respectively. The correlations of SIE with the LDSIL frequency are shown at the top of each panel in the corresponding colors. The ordinate of a variable is shown in the same color as the plot of the variable.

likely contributes to the weak correlations between the SIE of adjacent regions (Table S1).

The composite for the central Arctic (Fig. 7e) shows a somewhat different pattern. For this region, the geopotential height dipole anomaly is north–south oriented, with the poleward moisture flux anomaly occurring in the northeastern North Atlantic. The IVF anomalies are southward and southwestward elsewhere in the polar cap, associated with reduced AR activity. The enhanced poleward water vapor flux over the North Atlantic is consistent with Yang and Magnusdottir (2017), who showed the preferential occurrence of springtime ARs in the Atlantic sector contributing to the Arctic sea ice loss.

A similar north–south dipole pattern as seen in the central Arctic is found for the LDSIL days over the all-Arctic region (Fig. 7f). The anticyclone of the dipole pattern is consistent with reduced storm activity on the Pacific side of the pole. The equatorward water vapor transport is more extensive than the poleward anomalies, and significantly enhanced poleward IVF is absent south of 70°N . Since the all-Arctic $d\text{SIE}$ is the combination of various regions, it is possible that the poleward and equatorward water vapor flux anomalies are cancelled out when multiple regions are summed. The lack of strong poleward water vapor transport anomalies is consistent with some recent studies that emphasize the

importance of local processes on the Arctic amplification and pan-Arctic sea ice loss (e.g., Stuecker et al. 2018). Meanwhile, the strong equatorward flow east of Greenland may enhance sea ice transport out of the Arctic via the Fram Strait. It is also worth mentioning that the location and orientation of the dipole patterns in Figs. 7e and 7f differ from the commonly known Arctic dipole (see Fig. 2 in Wang et al. 2009) by its north–south orientation over the Atlantic sector.

The thermodynamic structures of the dipole anomalies are illustrated in Fig. 8. The vertical cross sections along the poleward flow over the marginal seas (Figs. 8a–d) reveal strong and nearly collocated positive anomalies of specific humidity and air temperature with maxima in the lower troposphere except over the Greenland Sea, where maximum warming occurs in the middle troposphere poleward of the positive specific humidity anomalies. The maximum warming near the surface differs from the maximum midtropospheric warming that would be expected from moist isentropic ascent (Laliberté and Kushner 2013; Fajber et al. 2018) and indicates the impacts of strong horizontal advection. Additionally, the maximum warming and moistening in the lower troposphere favor increased downwelling longwave radiation and are more conducive to sea ice loss than the maximum midtropospheric warming. For the central Arctic and all Arctic, a cross section is shown through the center of

FIG. 7. Composite anomalies of H500 (green contours; intervals: 10 m) above the 95% confidence level and the vertically integrated water vapor flux (IVF) vectors ($10 \text{ kg m}^{-1} \text{ s}^{-1}$). The IVF vectors are only shown for significant anomalies equatorward of 80°N with a magnitude exceeding $10 \text{ kg m}^{-1} \text{ s}^{-1}$. Pink and blue shadings depict significant positive and negative AR frequency anomalies, respectively (without showing the exact magnitudes). The anomalies are composited for all LDSIL days for each region.

the anomalous anticyclone (Figs. 8e,f). Maximum warm anomalies occur around 900 hPa near the center of the anticyclone, likely associated with adiabatic descent (Ding et al. 2017).

The frequency of occurrence of Rossby wave breaking was also examined. Since the PV or moisture filaments associated with RWB may last a few days and contribute to an LDSIL day, the composites of RWB frequency are constructed within ± 2 days of all LDSIL days, consistent with the definition of $d\text{SIE}$ (the drawback of this method is that it will include some days more than once if LDSIL occurs consecutively or with an interval less than 5 days). Both cyclonic and anticyclonic RWB frequencies were examined but the signature of the former is much weaker (see Fig. S2) and is not discussed here for brevity. For the composites based on regional-scale LDSIL days (Figs. 9a–e), anticyclonic RWB frequency is enhanced significantly in the midlatitude band between 40° and 60°N (Fig. 9) in the longitude sector of the dipole cyclone and/or its upstream (west) region, and

reduced RWB frequency is present in other longitude sectors or latitude bands. The composite for the all-Arctic LDSIL shows a region of enhanced RWB occurrence near the Barents Sea (Fig. 9f), similar to Fig. 9a, while the other regional features are lost in the all-Arctic composite. The longitude of the maximum positive RWB anomalies, however, is not well aligned with the regions of LDSIL or enhanced AR frequency. This is likely related to the RWB detection method, which defines the centroid of a high-PV tongue as the location of RWB, while the low-PV and high-moisture streamers can rotate and extend eastward and poleward over a long distance from the high PV tongues. Overall, the positive RWB anomalies indicate enhanced mixing between the midlatitudes and the Arctic in the vicinity of regional-scale LDSIL.

b. Lead/lag composites of daily data

To examine the temporal evolution of the atmospheric circulation anomalies and identify possible

Vertical Cross Sections of T and SH

FIG. 8. Vertical cross sections along various longitudes for LDSIL over different regions. Air temperature anomalies are shown in black contours with interval of 0.3 K (zero contour omitted); specific humidity anomalies are shown in red contours with intervals of 0.1 g kg^{-1} , and the anomalies above the 95% confidence level are highlighted in shading.

precursor signals for LDSIL, lead/lag composites were constructed with respect to the onset of LDSIL episodes. For this purpose, an LDSIL episode is defined as an LDSIL event lasting for at least four consecutive days, and the first day of the episode is defined as the onset day. Such multiday LDSIL events are likely associated with persistent and robust circulation anomalies and result in large reductions of SIE. To better illustrate anomalies in the tropics and subtropics, the 500-hPa streamfunction is shown instead of geopotential height. Wave activity flux vectors were calculated based on the composite mean streamfunction anomalies and the long-term mean zonal and meridional wind fields to estimate the wave energy propagation. For brevity, only the composites for the Beaufort-Chukchi-Siberian Seas and the Laptev-Kara Seas are shown (Figs. 10 and 11).

Twenty-two LDSIL episodes were identified for the Beaufort-Chukchi-Siberian Seas. Four days before the onset of LDSIL episodes, a weak but detectable wave

train extends from the North Pacific over Canada (Fig. 10). On day -2, the upstream cells over the northeastern Pacific and Canada Archipelago become stronger. Meanwhile, another wave train extends eastward from the Iberian Peninsula across northern Eurasia to 105°E. The downstream portion of the second wave train strengthens on day 0 from 45°E to 90°W. In particular, an anticyclonic cell over the Ural Mountain region intensifies, and the cyclone-anticyclone couplet develops over the Beaufort-Chukchi-Siberian Seas, which resembles the pattern in Fig. 7a in some respects (note that the samples are different). The cyclonic cell weakens from day 0 to day 2 and falls below the 95% confidence level on day 4, but the anticyclonic cell strengthens and extends eastward from day 0 through day 4 before weakening on day 6. Meanwhile, negative streamfunction anomalies develop south of Alaska from day 0 to day 6 and are associated with large eastward wave activity fluxes, resembling the feature prior to the LDSIL onset (from days -4 to 0).

Composite of H500 and ARWB (+/-2 days)

FIG. 9. Composite anomalies of anticyclonic RWB (ARWB) frequency (shading; %; only anomalies above the 95% confidence level are shown) within ± 2 days of LDSIL days superimposed on the anomalies of H500 (black contours; contour intervals are 15 m with the zero contour omitted). Note that a larger domain is plotted than in Fig. 7.

The Laptev–Kara composites (Fig. 11) show a wave train pattern along 40° – 60° N, extending from the Bering Sea to 15° W on day -4 and day -2 . The cyclone–anticyclone dipole over the Laptev–Kara Seas becomes well defined on day 0, accompanied by the weakening of the upstream wave train. The cyclonic cell of the couplet strengthens from day 0 to day 2 and then weakens gradually; the anticyclonic cell reaches its peak intensity on day 4 and remains prominent on day 6.

The lead/lag composites for the other regions show dipole patterns and precursor wave train patterns anchored to the corresponding regions of LDSIL (not shown). Despite the different geographic locations, there are some noticeable similarities. First, although some significant streamfunction anomalies are present in the subtropics on certain days (such as on day -4 in Fig. 10), the anomalies are rather patchy and their link to the midlatitude wave train is not clear. The wave activity flux shows a nearly zonal energy propagation path confined to the midlatitudes. This suggests that the precursor wave trains are likely tied to the internal dynamics of the mid- and high-latitude atmosphere, instead of being excited by tropical or subtropical forcing, which differs from the seasonal time scale (Ding et al. 2014; Grunseich and Wang 2016). Another striking feature of the sequences is the abrupt development of the cyclone–anticyclone couplets (with little

signature prior to the event). Additionally, the anticyclonic cell is nearly stationary and shows strong persistence, which is clearly shown in the Hovmöller diagram of the 500-hPa streamfunction in Fig. 12. Significant anomalies of the streamfunction last for 10 days over the Laptev–Kara Seas, the central Arctic, and the all-Arctic region. This persistent and nearly stationary nature of the anticyclone is reminiscent of blocking. The role of blocking in Arctic sea ice loss is emphasized by Wernli and Papritz (2018), and the persistence of blocking anticyclones can be partly attributed to diabatic heating (Pfahl et al. 2015; Wernli and Papritz 2018; Zhang and Wang 2018).

To further examine the physical processes leading to sea ice loss, we show in Fig. 13 the composite anomalies of column water vapor (CWV), surface downward longwave radiation, midlevel cloud fraction and surface downward shortwave radiation averaged over days 0–5 of the LDSIL episodes. We examined the low, midlevel, and high cloud fractions² from the ERA-Interim separately. The midlevel

² The low, midlevel, and high clouds represent clouds in the following layers, respectively, in σ coordinates: $1.0 > \sigma > 0.8$, $0.8 \geq \sigma \geq 0.45$, and $\sigma < 0.45$.

Lead/lag Composites of ψ_{500} and Wave Activity Flux: BeaChSi

FIG. 10. Lead/lag composite anomalies of 500-hPa streamfunction (contours; $10^4 \text{ m}^2 \text{ s}^{-1}$; anomalies exceeding the 95% confidence level are shaded) and wave activity flux vectors (shown equatorward of 80°N for magnitude larger than $5 \text{ m}^{-2} \text{ s}^{-2}$) with respect to the onset of LDSIL episodes (day 0) for the Beaufort–Chukchi–East Siberian Seas. The sample size is 22.

cloud fraction has a better agreement with downward shortwave radiation than low or high clouds, and only the midlevel cloud fraction is shown for brevity. Three regions, the Beaufort–Chukchi–Siberian Seas, the Laptev–Kara Seas and the all-Arctic seas, are shown as illustrative examples.

Significant positive anomalies of CWV and surface downward longwave radiation are found locally in association with the LDSIL over the Beaufort–Chukchi–Siberian Seas. These positive anomalies occur in the western portion of the anomalous anticyclone, roughly collocated with the anomalous poleward flow. The CWV anomalies are largest ($>2 \text{ mm}$) where the moisture flux enters the Arctic, with a decreasing magnitude as the flux approaches the Pole. The spatial pattern of downward longwave radiation anomalies is broadly consistent with

that of CWV anomalies, with the magnitude exceeding 12 W m^{-2} . The correspondence between the two fields strongly suggests that water vapor distribution is an important modulator for the longwave radiation anomalies, consistent with previous studies (e.g., Screen and Simmonds 2010; Kapsch et al. 2013; Cullather et al. 2016). Additionally, the anomalously warm poleward flow in the lower troposphere (Fig. 8) also enhances the downward longwave radiative flux, consistent with Liu and Schweiger's (2017) finding that the advection of warm, humid air contributes to earlier and more frequent sea ice melt. Interestingly, no coherent, significant anomalies of cloud fraction or downward shortwave radiative flux are found, although the role of shortwave radiation has been emphasized in some previous studies (Kay et al. 2008; Perovich and Richter-Menge 2009).

Lead/lag Composites of ψ_{500} and Wave Activity Flux: LapKara

FIG. 11. As in Fig. 10, but for the Laptev–Kara Seas with a sample size of 27.

The composites for the Laptev–Kara Seas are shown in the middle row of Fig. 13. Similar to the Beaufort–Chukchi–Siberian Seas, positive anomalies of CWV and downward longwave radiative flux are found in the western half of the anomalous anticyclone, within the region of the anomalous poleward flow. Different from the Beaufort–Chukchi–Siberian Seas, midlevel cloud fraction is significantly reduced at the center and southern part of the anomalous anticyclone, accompanied by enhanced downward shortwave radiation, and midlevel cloud fraction is enhanced around the center of the anomalous cyclone with reduced downward shortwave radiative flux. Although the magnitude of shortwave radiation anomalies is comparable to that of longwave radiation anomalies, the regions of greatest shortwave radiation anomalies do not correspond to the region of LDSIL or the regions of greatest sea ice cover changes (Fig. S3). The spatial offset in the case of the

Laptev–Kara Seas and the lack of strong shortwave radiation anomalies in the case of the Beaufort–Chukchi–Siberian Seas are consistent with previous studies on seasonal sea ice variability: earlier sea ice melt is associated with enhanced downward longwave radiation owing to higher water vapor content and warmer lower troposphere, but notably not associated with enhanced downward shortwave radiation (Kapsch et al. 2013, 2016; Mortin et al. 2016; Liu and Schweiger 2017; Crawford et al. 2018).

Enhanced CWV and downward longwave radiative flux are found in the northern half of the anomalous anticyclone for the LDSIL episodes over the all-Arctic region (bottom row in Fig. 13). Although midlevel cloud fraction is reduced near the center of the anticyclone, no significant anomalies of downward shortwave radiative flux are found due to the lack of significant changes in the total cloud fraction. The changes in sea ice cover are very scattered (Fig. S3).

FIG. 12. Hovmöller diagram of 500-hPa streamfunction anomalies (contour intervals $15 \times 10^4 \text{ m}^2 \text{ s}^{-1}$; the zero contours are omitted). Positive and negative anomalies exceeding the 95% confidence level are highlighted in pink and blue, respectively.

Overall, Fig. 13 suggests that enhanced downward longwave radiation plays an important role in LDSIL. The longwave radiation anomalies are tied to the warm, moist air mass advected from lower latitudes and not accompanied by significant cloud cover changes. Although short-wave radiation anomalies of comparable magnitude may result from cloud-radiation feedback, they do not occur in the same regions of LDSIL and do not seem to play an important role in LDSIL.

c. Seasonal mean anomalies

Previous studies suggested that the influence of atmospheric circulation anomalies on sea ice depends on the preconditions of sea ice (e.g., Döscher et al. 2014; Kwok 2018). In particular, sea ice thinning in preceding months or seasons makes sea ice more susceptible to atmospheric forcing. Given the lack of reliable long-term sea ice thickness data, we examine whether the Arctic atmosphere is anomalously warm and moist in the summer or the preceding spring precondition

LDSIL. Seasonal anomalies of geopotential height and column water vapor are constructed for the six years with the highest LDSIL seasonal frequency of occurrence during June–September (Table S2). No significant coherent increase in CWV or thickness is found poleward of 70°N in the preceding spring (March–May) except a warm, moist patch centered around 140°E in the composites for the Beaufort–Chukchi–Siberian Seas and the central Arctic (Fig. S4).

The June–September seasonal mean composites of H500 are shown in Fig. 14. Compared to Fig. 7, one may argue that the seasonal mean composite anomalies share some similarities with the daily composites for the Beaufort–Chukchi–Siberian, Laptev–Kara, and all-Arctic regions. The seasonal mean dipole patterns in those regions, however, are much weaker and less well-defined than the corresponding daily composites in Fig. 7. This may not be a surprise because the LDSIL days only account for 5% of a season on average and the associated anomalies can be easily masked out in

FIG. 13. Composite anomalies of (first column) CWV (shading; mm), (second column) download longwave radiation (dlwr; shading; W m^{-2}), (third column) midlevel cloud cover (MCC; shading; %), and (fourth column) downward shortwave radiation (dswr; shading; W m^{-2}) averaged over days 0–5 for the (top) Beaufort–Chukchi–Siberian, (middle) Laptev–Kara, and (bottom) all-Arctic seas. The anomalies exceeding the 95% confidence level are highlighted in shading, and the black contours represent H500 (contour intervals are 15 m with the zero contour omitted).

the seasonal mean. The differences between the two time scales highlight the need to examine the synoptic-scale variability and processes for a better understanding of the interannual variability of SIE. Figure 14 also shows the seasonal mean CWV anomalies. The positive (negative) anomalies of CWV are loosely collocated with positive (negative) anomalies in 950–500-hPa thickness, and the increase in CWV is associated with the higher specific humidity in a warmer air column (Ding et al. 2017). It is also worth noting that the H500 anomalies for the LDSIL over the Beaufort–Chukchi–East Siberia Seas are similar to the linear trend of geopotential height during 1979–2014 (Ding et al. 2017), consistent with the positive trend of the LDSIL frequency in this region (Fig. 6).

Although no coherent, significant signals in the CWV are found in the preceding spring, the results do not mean that sea ice preconditions do not play a role in LDSIL.

For example, the LDSIL frequency during summer shows a positive trend over the Beaufort–Chukchi–East Siberian Seas and the central Arctic (Fig. 6), and LDSIL days tend to occur earlier in the season in recent years over the Laptev–Kara Seas, the central Arctic and the all-Arctic seas (Fig. 15). However, a similar trend, or a change in the seasonality, is absent in the occurrence of the dipole patterns for those regions (Fig. S5). Taken together, the results suggest that the dynamic changes of the atmospheric circulation probably do not play the primary role in the trends of the LDSIL frequency (Deser and Teng 2008), and that sea ice thinning and the thermodynamic changes of the atmospheric circulation (i.e., higher specific humidity in warmer air) may be the major players. It is also interesting to note the negative trends of LDSIL occurrence over the Barents and the Greenland Seas, especially in late summer (Fig. 15), which can be attributed to the prolonged sea ice-free period and thus

Composite of H500, CWV and dZ: JJAS

FIG. 14. The June–September (JJAS) seasonal mean composite anomalies of H500 (contour intervals are 10 m with the zero contour omitted) and CWV (shading; mm) for the six years with the highest LDSIL seasonal frequency. Only significant CWV anomalies (exceeding the 95% confidence level) are shown. Circles and asterisks highlight regions of significant positive and negative anomalies of 500–950-hPa thickness (exact values not shown), respectively.

weak summer sea ice variability over these regions in recent years.

6. Summary

Large sea ice loss on the synoptic time scale, termed large daily sea ice loss (LDSIL), is examined for different subregions over the Arctic as well as the pan-Arctic scale using daily sea ice extent data from 1979 to 2017. The criterion for an LDSIL in a region is a 5-day sea ice change that is below the 5th percentile in that region. After removal of the seasonal cycle, sea ice loss is characterized by strong variability on the synoptic time scale instead of a gradual decrease throughout the season, and the days of extremely low sea ice extent (with a SIE minimum at least one standard deviation below the mean) are often preceded by LDSIL days. The frequency of LDSIL occurrence during June–September has a significant negative correlation with the September SIE over the Beaufort–Chukchi–East Siberian Seas, the Laptev–Kara Seas, the central Arctic, and the pan-Arctic scale. Years of record-low SIE are characterized by frequent occurrence of LDSIL. The results

suggest that extreme synoptic-scale events occurring on ~5% of the days of the antecedent period make a disproportionately large contribution to the annual minimum SIE and highlights the weather–climate connection in terms of sea ice variability.

Composite daily anomalies were constructed to examine the atmospheric circulation anomalies associated with LDSIL days. A robust dipole pattern of 500-hPa height anomalies was found geographically locked to the subregion experiencing large sea ice loss. An interesting feature of the dipole pattern is the persistence and quasi-stationary nature of the anticyclone. Composite mean anomalies show that the anticyclone may last for up to 10 days and resembles a blocking event (Wernli and Papritz 2018). Poleward flow is enhanced between the anomalous cyclone and anticyclone, which increases the moisture and heat transport into the Arctic. Atmospheric rivers occur more frequently between the dipole centers in the Arctic marginal seas, while poleward moisture transport and AR frequency are reduced in other regions. The atmospheric river signature found in this study is a robust feature of LDSIL events across the different subregions (but not

LDSIL Occurrence

FIG. 15. LDSIL occurrence (red blocking) from 1979 to 2017. The vertical axis shows calendar days from 1 Jun (day 1) to 30 Sep (day 122).

for the pan-Arctic scale), suggesting a need to include poleward moisture fluxes in diagnostic assessments of regional sea ice loss. Furthermore, RWB frequency increases significantly in the midlatitudes upstream (west) of the dipole pattern, indicating that the enhanced mixing between the midlatitudes and the Arctic contributes to Arctic sea ice loss. The finding is consistent with previous studies that emphasize the role of moisture transport in hindering sea ice growth in winter and promoting sea ice melt in spring–summer (Park et al. 2015; Hegyi and Taylor 2018; Mortin et al. 2016; Liu and Schweiger 2017). Additionally, increasing local evaporation owing to reduced sea ice cover may help further increase column water vapor (Screen and Simmonds 2010; Liu et al. 2012; Bintanja and Selten 2014) and accelerate sea ice loss.

In contrast, the composites for LDSIL over the all-Arctic region are not associated with enhanced poleward moisture transport or increasing occurrence of

ARs, since regional moisture transport anomalies are partially cancelled out when the subregions are summed together. This finding is consistent with some recent studies that emphasize local feedback processes instead of remote impacts on Arctic sea ice loss as a whole. Meanwhile, our findings also suggest that the physical processes controlling the regional and pan-Arctic-scale sea ice variability may be different.

With regard to the temporal evolution of atmospheric circulation and the associated radiative forcing, significant but weak midlatitude wave trains are found preceding LDSIL episodes. Wave activity flux analysis reveals a predominantly zonal energy dispersion along the wave train, and the link of the wave train to tropical or subtropical anomalies is tenuous. This suggests that the LDSIL episodes are likely tied to the midlatitude atmospheric internal variability instead of tropical or subtropical forcing.

Surface downward longwave and shortwave radiative fluxes following the onset of LDSIL episodes were also

examined. Enhanced downward longwave radiation was found in the regions of LDSIL. The downward longwave radiation anomalies are closely related to the changes in CWV and mainly occur in the periphery of the anomalies anticyclone where the pressure gradient is largest and an anomalous poleward flow is observed. Although downward shortwave radiation anomalies of comparable magnitude were found in the composites for some regions, the greatest shortwave radiation anomalies do not collocate with LDSIL, suggesting that shortwave radiation anomalies may not play an important role in LDSIL. However, shortwave radiation may help amplify the Arctic warming after sea ice cover and the surface albedo are reduced as suggested by some previous studies (e.g., Kapsch et al. 2016).

The summer seasonal mean atmospheric fields for the years of frequent LDSIL occurrence show similarities to the daily composites in some regions, but the dipole pattern of an anomalous cyclone and anticyclone is less well defined, and significant CWV anomalies are absent for most regional composites. This highlights the need to examine the synoptic variability for a better understanding of seasonal anomalies. In particular, global models generally have difficulty in representing the spatial distribution, frequency, and duration of blocking (e.g., Davini and D'Andrea 2016), and such model deficiencies may contribute to the underestimated sea ice variability and trend in a global model (e.g., Winton 2011; Rosenblum and Eisenman 2017).

While our focus has been primarily on the thermodynamic effects of atmospheric circulation anomalies on sea ice loss, other processes also induce the Arctic sea ice variability. For example, surface winds drive sea ice deformation and transport (e.g., Parkinson and Comiso 2013; Ogi and Wallace 2007; Mills and Walsh 2014), and dynamic and thermodynamic forcings by the atmosphere are generally correlated, that is, southerly winds drive sea ice poleward while advecting positive air temperature and moisture anomalies over sea ice (Yu et al. 2019). Moreover, oceanic heat anomalies induced by enhanced vertical mixing and horizontal advection may cause sea ice thinning and precondition LDSIL (e.g., Årthun et al. 2012; Zhang et al. 2013). Ocean heat transport through the Bering Strait has been shown to precondition sea ice in the Chukchi and Beaufort Seas (Serreze et al. 2016), while antecedent ocean temperatures serve as predictors of interannual variations of sea ice in the North Atlantic sub-Arctic (Bushuk et al. 2017). Variations of quantitative as well as qualitative assessments of the roles of different physical and dynamical processes in LDSIL warrant further study. In addition, although no significant correlations are found between the

atmospheric conditions in the preceding spring season and the summertime LDSIL frequency, preconditions in terms of sea ice thinning cannot be ruled out. It is possible that sea ice thinning makes sea ice more susceptible to the atmospheric forcing and contributes to a positive trend in the LDSIL frequency in some regions. In this case, LDSIL events may occur more frequently in a warmer climate.

Acknowledgments. This study was supported by the Office of Naval Research through Grant N000141812216. This study benefits from the discussion with Drs. Qinghua Ding and Edward Blanchard-Wrigglesworth, and we are grateful to Dr. Bin Guan for making the atmospheric river dataset available.

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