Turbulent Heat Flux, Downward Longwave Radiation, and Large-Scale Atmospheric Circulation Associated with Wintertime Barents–Kara Sea Extreme Sea Ice Loss Events

CHENG ZHENG,a MINGFANG TING,a YUTIAN WU,a NATHAN KURTZ,b CLARA ORBE,c PATRICK ALEXANDER,a,c RICHARD SEAGER,a AND MARCO Tedesco,a,c

a Lamont-Doherty Earth Observatory, Columbia University, Palisades, New York
b NASA Goddard Space Flight Center, Greenbelt, Maryland
c NASA Goddard Institute for Space Studies, New York, New York

(Manuscript received 21 May 2021, in final form 14 January 2022)

ABSTRACT: We investigate wintertime extreme sea ice loss events on synoptic to subseasonal time scales over the Barents–Kara Sea, where the largest sea ice variability is located. Consistent with previous studies, extreme sea ice loss events are associated with moisture intrusions over the Barents–Kara Sea, which are driven by the large-scale atmospheric circulation. In addition to the role of downward longwave radiation associated with moisture intrusions, which is emphasized by previous studies, our analysis shows that strong turbulent heat fluxes are associated with extreme sea ice melting events, with both turbulent sensible and latent heat fluxes contributing, although turbulent sensible heat fluxes dominate. Our analysis also shows that these events are connected to tropical convective anomalies. A dipole pattern of convective anomalies with enhanced convection over the Maritime Continent and suppressed convection over the central to eastern Pacific is consistently detected about 6–10 days prior to extreme sea ice loss events. This pattern is associated with either the Madden-Julian oscillation (MJO) or El Niño–Southern Oscillation (ENSO). Composites show that extreme sea ice loss events are connected to tropical convection via Rossby wave propagation in the midlatitudes. However, tropical convective anomalies alone are not sufficient to trigger extreme sea ice loss events, suggesting that extratropical variability likely modulates the connection between tropical convection and extreme sea ice loss events.

KEYWORDS: Arctic; Atmospheric circulation; Extreme events; Subseasonal variability

1. Introduction

Observational studies show that Arctic warming over the past few decades is most rapid during the winter season (Bekryaev et al. 2010; Screen and Simmonds 2010). Such warming is driven by many factors including greenhouse gas increases, loss of sea ice, and energy transport by the atmosphere and the ocean (Carmack and Melling 2011; Serreze and Barry 2011; Bintanja and van der Linden 2013; Pithan and Mauritsen 2014). Recently, the role of downward longwave radiation (LWR) on both the long-term trend of Arctic sea ice decline and the fluctuations of the Arctic sea ice on shorter time scales has been addressed by many studies (Francis and Hunter 2006; Graversen and Wang 2009; Graversen et al. 2011; Lee et al. 2011b; Woods et al. 2013; D. Park et al. 2015; H. Park et al. 2015a,b; Burt et al. 2016; Woods and Caballero 2016; Gong et al. 2017; Lee et al. 2017; Johansson et al. 2017; Kim and Kim 2017; Z. Wang et al. 2020). Many of these studies also showed that fluctuations of downward LWR are often associated with moisture intrusions into the Arctic. Sea ice fluctuations can also be driven by other factors including wind-induced sea ice motion (Fang and Wallace 1994; Rigor et al. 2002; Rigor and Wallace 2004; Kwok 2005; Sorteberg and Kvingedal 2006; Liptak and Strong 2014) and local temperature anomalies (Deser et al. 2000). The sea ice variability on synoptic to subseasonal time scales has not received much attention until recently. Studies show that Ural blocking is important in modulating winter Arctic sea ice on short time scales, particularly over the Barents–Kara Sea (BKS) region (Gong and Luo 2017; Luo et al. 2017; Chen et al. 2018; Cho and Kim 2021). The primary goal of this study is to identify the primary drivers of extreme Arctic sea ice loss on these time scales during winter.

Though the role of downward LWR associated with moisture intrusions has been emphasized previously (e.g., Woods and Caballero 2016; D. Park et al. 2015; Yang and Magnusdottir 2017), these studies focused on the sea ice response following strong moisture intrusion events. Similar results regarding the role of downward LWR are reached with composite studies of Ural blocking events and the associated sea ice response (Gong and Luo 2017; Luo et al. 2017; Chen et al. 2018). However, it is not clear if all extreme sea ice loss events over the BKS region are primarily driven by downward LWR. Another possible driver is the turbulent heat flux (THF), which consists of sensible heat flux (SHF) and latent heat flux (LHF), which is directly associated with atmospheric energy transport. The role of THF on sea ice melting has been suggested by Blackport et al. (2019) and Sorokina et al. (2016). The results in one recent study by Jiang et al. (2021) suggest that the amplitude of THF could be
as large as that of downward LWR during sea ice melting events in the BKS. In this study, instead of focusing on moisture intrusion events, we will start from extreme sea ice loss events and analyze the primary drivers of these events and the associated atmospheric circulation and energy fluxes. The BKS region is chosen as this is the region with the largest sea ice trend (e.g., Screen and Simmonds 2010) and winter sea ice variability (e.g., Kim et al. 2017; Zhong et al. 2018).

It is also not well understood what the precursors or preferred teleconnection patterns are associated with extreme sea ice loss events over the BKS region on synoptic and seasonal time scales. Case studies (e.g., Boisvert et al. 2016; Cullather et al. 2016; Overland and Wang 2016; B. Kim et al. 2017) and long-term analysis (Sorteberg and Walsh 2008; Rinke et al. 2017; Messori et al. 2018; Fearon et al. 2021) have shown that moisture intrusions into the Arctic can be driven by extratropical cyclones. Moisture intrusions and extreme downward LWR events in the Arctic are also related to atmospheric blocking and Rossby wave breaking (Liu and Barnes 2015; Woods et al. 2013), as well as planetary-scale waves (Papritz and Dunn-Sigouin 2020). For the BKS region, a positive North Atlantic Oscillation (NAO) phase and Ural blocking favor moisture intrusions into the region (e.g., Gong and Luo 2017; Luo et al. 2017; Zhong et al. 2018). The constructive interference of planetary waves has also been found to be related to warm episodes in different regions in the Arctic (e.g., Baggett and Lee 2015; Baggett et al. 2016; Goss et al. 2016), which can result in sea ice loss.

Tropical forced planetary-scale waves can induce Arctic warming (Lee et al. 2011a,b; Lee 2012; Flournoy et al. 2016), as tropical convection acts as a heat source that excites Rossby waves that propagate into high latitudes (e.g., Hoskins and Karoly 1981). Moisture intrusion events over the Atlantic side of the Arctic are found to be associated with enhanced convection over the Maritime Continent (H. Park et al. 2015; Flournoy et al. 2016). The Madden–Julian oscillation (MJO; Madden and Julian 1971, 1972, 1994) is the dominant mode of intraseasonal tropical variability and is characterized by large-scale eastward propagating tropical convection. Previous studies have shown that the MJO can induce Arctic sea ice variability (Henderson et al. 2014) and surface air temperature variability (Yoo et al. 2012). Jiang et al. (2021) found that winter sea ice loss events in the BKS are significantly related to the MJO.

In this study, lag composite analysis is applied to the extratropical large-scale circulation and tropical convection prior to extreme sea ice loss events to investigate the associated temporal and spatial evolution of the atmospheric circulation. Although previous studies have found a significant connection between extreme sea ice loss and tropical convective anomalies (e.g., H. Park et al. 2015; Flournoy et al. 2016; Jiang et al. 2021), the evolution of the extratropical circulation that connects tropical convection and BKS sea ice loss has not been analyzed in detail, and the dynamics that link tropical convection and BKS sea ice loss are yet to be understood. The goal of the lag composite analysis presented here is to provide useful insights regarding the dynamical links between BKS extreme sea ice loss and tropical and midlatitude circulation. The data and methods used in this study will be introduced in section 2. The circulation patterns in the high latitudes during extreme sea ice loss events, as well as the role of radiative and turbulent heat fluxes, will be explored in section 3. The circulation patterns in both the extratropics and tropics prior to extreme sea ice loss events will be investigated in section 4, followed by conclusions in section 5.

2. Data and methods

a. Data

We use sea ice concentration from the Climate Data Record of Passive Microwave Sea Ice Concentration, version 3, provided by the U.S. National Snow and Ice Data Center (NSIDC). This version of NSIDC data (Peng et al. 2013) is a combination of sea ice concentration estimates from both the National Aeronautics and Space Administration (NASA) team algorithm (Cavalieri et al. 1984) and NASA Bootstrap (BT) algorithm (Comiso 1986). The data are provided on a polar stereographic grid with a 25 km × 25 km spatial resolution; the data after 1989 are used since daily data are not available before 1989. For energy budget terms and atmospheric circulation fields, the Modern-Era Retrospective Analysis for Research and Applications, version 2 (MERRA-2; Gelaro et al. 2017) is used. The hourly data for variables including surface radiative fluxes (both upward and downward longwave and shortwave), surface latent and sensible heat fluxes, skin temperature and 2-m temperature, column water vapor, and vertical integrated water vapor flux, as well as 6-hourly temperature, geopotential height, and wind on pressure levels, are used to diagnose the energy budget and large-scale circulation associated with BKS extreme sea ice loss events. The horizontal resolution of MERRA-2 data is 0.625° longitude × 0.5° latitude. Note that we also use the fifth-generation ECMWF reanalysis (ERA5; Hersbach et al. 2020) to verify the results. Because the conclusions we obtain from the two reanalyses are similar, only the results from MERRA-2 will be shown.

To investigate whether BKS extreme sea ice loss events are connected with tropical convection, we use the National Oceanic and Atmospheric Administration (NOAA) interpolated outgoing longwave radiation (OLR) (Liebmann and Smith 1996) with horizontal resolution of 2.5° longitude × 2.5° latitude, with negative OLR anomaly representing enhanced tropical convection and positive OLR anomaly representing suppressed convection. The Optimum Interpolation Sea Surface Temperature (OISST) version 2.1 (Huang et al. 2020), which provides daily sea surface temperature (SST) data at 0.25° resolution, is used to investigate high-frequency SST variability. The Oceanic Niño Index (ONI) provided by NOAA, which is the 3-month running mean Niño-3.4 index, is used to define winter El Niño–Southern Oscillation (ENSO) conditions. The Niño-3.4 index is derived from the Hadley Centre’s Sea Ice and Sea Surface Temperature dataset (HadISST1; Rayner et al. 2003). The Real-Time Multivariate MJO (RMM) index is a commonly used MJO index (Wheeler and Hendon 2004), which is based on multivariate
empirical orthogonal function (EOF) analysis of combined fields of OLR and 850- and 200-hPa zonal wind anomalies. The time series of the two leading EOFs (RMM1 and RMM2) can separate the MJO into eight different phases. The RMM index is used to define the MJO in this study.

Satellite products are also used to verify some of the results from MERRA-2. Surface downward LWR from Clouds and the Earth’s Radiant Energy System (CERES) synoptic 1° × 1° (SYN1deg) data (Rutan et al. 2015), which have been available since 2002, are used to verify the downward LWR results in MERRA-2. Since MERRA-2 does not assimilate CERES data, the satellite data can be considered as independent from MERRA-2. The NASA Atmospheric Infrared Sounder (AIRS) daily column water vapor data (Kahn et al. 2014) at 1° × 1° resolution from 2002 to 2016 are used to verify the column water vapor results from MERRA-2. Overall, results from MERRA-2 are consistent with those from the satellite products (see section 3).

b. Methods

We investigate BKS extreme sea ice loss events during 30 winter seasons [December–February (DJF)] from 1989/90 to 2018/19. Lag composite analysis is applied to investigate the role of downward LWR and surface turbulent heat fluxes on extreme sea ice loss events in the BKS, as well as the large-scale circulation associated with the events. Daily averages of MERRA-2 are first calculated from hourly data. Prior to making composites, the seasonal cycle is removed at each grid point for all variables; the seasonal cycle is defined as a 31-day moving average of the 31-yr mean (1989–2019) for each calendar day. As we focus on sea ice loss events within synoptic to subseasonal time scales, interannual variability is also removed at each grid point for all variables by subtracting their corresponding winter season averages.

The BKS sea ice concentration (SIC) is calculated by averaging SIC over 70°–82°N, 15°–100°E (Zhang et al. 2018). The BKS sea ice loss events are defined as the time period when the BKS SIC anomaly falls below negative one standard deviation for at least 5 consecutive days with no less than 5 days apart between two events. There are 25 events that fit the criteria during the 30 winter seasons. Lag zero (the beginning of the event) is defined as the first day when BKS SIC falls below −1 standard deviation. Sensitivities to the sea ice criterion are tested using standard deviations of −0.8 and −0.5, and the overall results are similar to that using −1 standard deviation except for a higher number of events and weaker amplitudes. Note that our conclusions remain the same if we define lag zero as the day of minimal sea ice for each event.

The temporal evolution of the sea ice anomaly associated with extreme sea ice loss events is similar to that found by Gong et al. (2020) with a composite analysis based on maximum latent heat index over the Greenland, Barents, and Kara Seas.

3. Energy budget and circulation patterns during the sea ice loss events

To understand the patterns of sea ice loss and the associated atmospheric processes, composite maps of SIC, vertically integrated water vapor flux, column water vapor, and 500-hPa geopotential height (Z500) from day −4 to day +4 with a 2-day interval are shown in Fig. 2. Significant sea ice losses (Figs. 2b–e) can be seen around day −2 over the BKS region reaching a maximum sea ice loss at day +4. Although SIC anomalies could be related to wind-driven sea ice motion, there are no regions with SIC gain near the SIC loss region over BKS since day 0 (Figs. 2c–e), indicating that the net decrease of SIC over the BKS region is mostly driven by thermodynamic loss rather than dynamical transport, which redistributes the sea ice locally. Strong northward moisture fluxes (Fig. 2f) develop along the 0° longitude line to the southwest of BKS region at day −4, intensifying quickly by day −2. A large amount of moisture influx reaches the BKS region at day zero (Fig. 2h), resulting in a large column water vapor increase over the entire BKS region at day zero (Fig. 2m). The Z500 field shows a dipole pattern (Figs. 2p–t), with negative anomalies over Greenland, extending into eastern

![BKS sea ice concentration](image_url)
Canada, and positive anomalies centered slightly south of the BKS region from day −4 to +2. The circulation pattern is to some extent similar to that in Gong and Luo (2017), Luo et al. (2017), and Chen et al. (2018), who found that Ural blockings are related to sea ice loss over the BKS region. However, the Z500 anomalous pattern is located more northward compared with typical composites of Ural blocking patterns, as the positive Z500 center is over the Kara Sea rather than the Ural

FIG. 2. (a)–(e) Composites of sea ice concentration anomalies (percentage) during lag day −4, −2, 0, 2, and 4, respectively. (f)–(j) As in (a)–(e), but for vertical integrated water vapor flux (kg m s⁻¹). (k)–(o) As in (a)–(e), but for column water vapor (kg m⁻²). (p)–(t) As in (a)–(e), but for 500-hPa geopotential height (m). The black contour in (k)–(t) shows the significance level of 95% from the results of Student's t test. The dark green box depicts the BKS region (70°–82°N, 15°–100°E).
Mountains. Similar to what has been discussed in these previous studies, this dipole pattern drives warm and moist air from the Atlantic into the BKS region along the 0° longitude line. Note that the spatial patterns of the composite water vapor fluxes and Z500 are very similar to those presented in previous studies (H. Park et al. 2015; Papritz and Dunn-Sigouin 2020) where the composites were based on strong moisture intrusion events. The similarity suggests that extreme sea ice loss events are strongly linked to strong moisture intrusions.

Composite maps of surface downward LWR, SHF, and LHF, as well as horizontal temperature advection at 950 hPa, are further examined in Fig. 3 to determine the dominant drivers of the sea ice loss events. A negative sign is added to both SHF and LHF, so that positive values of both downward LWR and turbulent heat flux mean that the atmosphere heats up the surface. As wintertime SWR is negligible and surface emitted LWR is considered as a response rather than a forcing, the main focus here is on the downward LWR and turbulent heat fluxes. A strong downward LWR anomaly can be found over the BKS region, which peaks around day zero (Fig. 3c). The downward LWR anomaly weakens and moves farther northward and eastward at days +2 to +4 (Figs. 3d,e). The strong downward LWR is largely associated with the strong moisture intrusions (Figs. 2g,h), as pointed out in previous studies. The SHF anomalies (Figs. 3f–j) appear earlier than the downward LWR and with a larger amplitude, particularly in the western part of the BKS, while LHF anomalies (Figs. 3k–o) show similar spatial patterns and temporal evolution as SHF but with a lesser amplitude. The spatial patterns of downward LWR, SHF, and LHF are similar to that in Luo et al. (2019), with larger amplitudes of SHF and LHF compared with downward LWR in the western part of the region. This suggests that apart from LWR, turbulent heat fluxes also play an important role during extreme sea ice loss events. As anomalous sensible heat fluxes are strongly linked to near-surface atmospheric temperature anomalies that are mostly forced by near-surface horizontal temperature advection (V · VT) over this region during winter (Clark and Feldstein 2020), the patterns of 950-hPa temperature advection are also shown in Fig. 3. As expected, the pattern of 950-hPa temperature advection agrees well with the SHF, suggesting that the atmospheric circulation patterns associated with extreme sea ice loss not only cause moisture intrusions but also produce strong heat transport into the BKS region. Note that 950 hPa is shown here as temperature advection within the troposphere peaks near the surface, which is consistent with Clark and Feldstein (2020). Also, downward SHF anomaly is driven by the horizontal temperature advection but not dynamical warming due to vertical motion, as rising motion is associated with the events (not shown). The moisture transport associated with moisture intrusions also leads to an increase in near-surface specific humidity, which drives anomalous downward latent heat flux.

There are notable temporal and spatial differences between downward LWR and SHF anomalies in Fig. 3. As noted earlier, SHF develops earlier than downward LWR, as a large SHF anomaly is already present around day −4, whereas the strongest downward LWR occurs between day −2 and day 2 (also see Fig. 4), with the largest anomaly over the BKS regions at day zero. The largest amplitude of the SHF is over the sea ice loss region (Figs. 3f–h) and the open ocean, while downward LWR covers the sea ice loss region and extends farther to the north. On interannual time scales, turbulent heat flux associated with wintertime sea ice loss often shows an upward anomaly over the sea ice loss region, and the upward turbulent heat flux anomaly is considered as a response to sea ice loss, since open water allows more turbulent heat flux from the ocean to the atmosphere (Kim and Kim 2017). In contrast, here we find that turbulent heat fluxes have a downward anomaly during extreme sea ice loss events on synoptic and subseasonal time scales. Thus, THF acts as a forcing to sea ice loss rather than a response. Blackport et al. (2019) and Sorokina et al. (2016) found that the warm Arctic–cold Siberia pattern is associated with reduced sea ice and a downward THF anomaly over the BKS region, suggesting that the atmosphere is driving the sea ice variability via THF over the BKS. Our results here more clearly illustrate that THF plays a major role in sea ice loss events on subseasonal time scales. In general, the advection of warm and moist air into the region due to the large-scale atmospheric circulation, which leads to both anomalous downward THF and downward LWR, is the driver of the extreme sea ice loss events.

The sea ice loss and energy budget associated with the events averaged over the BKS region are summarized in Fig. 4a, which more quantitatively shows the day-to-day evolution of the surface energy terms. A 3-day running average is performed to smooth the temporal evolution. The composite mean SIC (thick solid red line) drops from 0 at around day −6 to about 7% loss at day +4. The SWR anomaly (gray line) is near zero throughout the period as it is negligible during winter. A positive downward LWR anomaly (green line) develops from day −4 and peaks at around days 0 to +1. The peak downward LWR leads the peak sea ice loss by 3–4 days, consistent with previous studies (e.g., Woods and Caballero 2016; H. Park et al. 2015). Downward SHF (light-blue line) and LHF (dark-blue line) develop at around day −10 and reach their peaks at around days −2 to 0. Both positive SHF and LHF develop earlier than downward LWR by a few days, and positive SHF has a larger amplitude than downward LWR before they both peak near day −1, as the difference between SHF and downward LWR is statistically significant from day −6 to day −3 at the 5% level. Averaging over the BKS region, the peak amplitude of SHF is comparable to the amplitude of downward LWR, whereas the THF (dashed purple line), the sum of SHF and LHF, is clearly larger than downward LWR. The total surface energy budget, which is the sum of turbulent heat fluxes and incoming and outgoing SWR and LWR, is shown by the dashed yellow line. During days −8 to −4, the surface gains energy from the atmosphere mostly through turbulent heat fluxes, while after day −4, both turbulent heat fluxes and downward LWR contributes. The total energy budget peaks at around days −2 to 0, corresponding to when the SIC has the sharpest decrease. At days +4 to +6, SHF and LHF anomalies drop to zero, and surface emitted LWR (dashed black line) tends to balance the downward LWR. The anomalous energy budget terms cancel each
other, as shown by the total energy budget. As there is little net gain of energy at the surface, sea ice loss stops after days 1 to 6. After that, sea ice concentration slowly recovers to the normal level. Note some of the large SHF and LHF anomalies, especially over the western part of the BKS region, take place over sea ice–free regions that would not directly contribute to the sea ice melting [e.g., Luo et al. (2019)]. Therefore, we also examine the surface energy budget excluding the sea ice free areas by selecting only the grid points with a large sea ice decrease (Fig. S2 in the online supplemental material).
The THF in this case reduces slightly from that for the entire BKS average, but still possesses a comparable amplitude as downward LWR, confounding the importance of THF in melting the sea ice. Over the grid points with large sea ice decrease, the amplitude of LHF is relatively small compared to SHF and downward LWR, showing that LHF is playing a minor role in melting the sea ice.

Note that although Jiang et al. (2021) did not focus on the role of downward LWR and THF during the sea ice melting stage (they focused instead on the atmospheric response to sea ice loss), our results here are nevertheless consistent with their Fig. 1, which shows that downward THF occurs a few days ahead of downward LWR. Turbulent heat flux (THF; the sum of sensible and latent heat flux) is shown by the dashed purple line. The total energy budget at the surface is shown in the dashed yellow line. Units are W m\(^{-2}\).

Fig. 4. (a) Temporal evolution of composites of SIC and surface energy fluxes averaged over the BKS region (70\(^\circ\)-82\(^\circ\)N, 15\(^\circ\)-100\(^\circ\)E; also see Fig. 1). The thick solid red line shows the SIC anomalies (as in Fig. 1). Surface downward LWR, net downward SWR, negative sensible heat flux, and latent heat flux are shown by the solid green, gray, light blue, and dark blue lines, respectively. Surface emitted LWR is shown by the dashed black line. Turbulent heat flux (THF; the sum of sensible and latent heat flux) is shown by the dashed purple line. The total energy budget at the surface is shown in the dashed yellow line. Units are W m\(^{-2}\).

(b) Water vapor flux convergence (kg m\(^{-2}\) day\(^{-1}\)), column water vapor (kg m\(^{-2}\)), and 950-hPa temperature advection (K day\(^{-1}\)) are shown by the thick dashed red, orange, and magenta lines. Surface downward LWR, negative sensible heat flux, and latent heat flux are shown by the thin lines. The vertical axis is different for water vapor flux convergence, column water vapor, and 950-hPa temperature advection compared with energy fluxes, which is shown on the right side of the panel. The time series in (a) and (b) are smoothed by applying a 3-day running mean.

The evolution of 950-hPa temperature advection and water vapor flux convergence are similar to each other with a peak at around days \(-2\) to \(-1\), which is not surprising as the same atmospheric circulation pattern (e.g., Z500 pattern in Fig. 2) drives both temperature advection and moisture flux into the BKS region. As SHF (LHF) is a direct response to the atmospheric sensible heat (moisture) transport, SHF (LHF) peaks at the same time as the 950-hPa temperature (moisture) advection. The change in column water vapor, on the other hand, is the integral effect of the water vapor flux convergence. Thus, there is a delay of \(1\) to \(2\) days in the peaking of column water vapor compared to that of water vapor flux convergence. The downward LWR (green line) is determined by both air temperature and column water vapor, and the peak of downward LWR agrees well with that of the column water vapor. Therefore, THF (both SHF and LHF) develops a few days earlier than the downward LWR and has a larger amplitude than downward LWR before their peaks, suggesting that THF is not a result of increased downward LWR, but rather a main driver of sea ice loss. Note that we also use AIRS column water vapor to verify the reanalysis data (Fig. S3) and the results are very similar to MERRA-2, which may not be surprising since MERRA-2 assimilates AIRS.
observations (Gelaro et al. 2017). We have also attempted to use THF data derived from AIRS (Boisvert et al. 2016) to verify our results. However, as AIRS THF data over the BKS region are not available during about 80% of the events, likely due to cloud cover, it is challenging to compare the composites of THF between MERRA-2 and AIRS.

Another possible factor that can affect the downward LWR is cloud cover, as shown by Sokolowsky et al. (2020) that the contribution from liquid and ice water on downward LWR could be similar due to increased water vapor during moisture intrusion events in the Arctic. Liu et al. (2018) found a significant contribution from clouds to downward LWR associated with wintertime moisture intrusion events in the Arctic, as cloud amount and cloud water content increases during moisture intrusion events. Here, we compare downward LWR in all-sky and clear-sky conditions from both reanalysis (MERRA-2) and the satellite product (CERES) in Fig. S4. The results are very consistent between MERRA-2 and CERES, with about 1–2 W m⁻² difference between all-sky and clear-sky conditions, compared with the peak of downward LWR of 15–16 W m⁻². This suggests that clouds play a minor role in modulating downward LWR during BKS extreme sea ice loss events. It is not entirely clear why clouds do not have a significant contribution to downward LWR in our study, in apparent contradiction to Liu et al. (2018). However, Liu et al. (2018) focused on water vapor intrusion events rather than sea ice loss events, which is different from our study.

Previous studies (e.g., Luo et al. 2017; Chen et al. 2018) have found that Ural blocking events can induce large sea ice loss over the BKS region, where downward LWR plays a major role in surface energy flux in these events while the influence of THF is secondary, which seems to contradict with our results. However, the sea ice loss associated with Ural blocking in Luo et al. (2017) and Chen et al. (2018) centers around Yuzhny Island and Severny Island, which is over the eastern part of the BKS region, whereas the sea ice loss in our study is mainly over the western and northern part of the BKS region (e.g., Fig. 2e). To examine the sensitivity of our results to the location of sea ice loss, we split our original BKS region into the western (70°–82°N, 15°–50°E) and eastern BKS (70°–82°N, 50°–100°E), and repeat the analysis for these two regions separately. Similar to previous studies, for eastern BKS (Fig. S5) the large negative SIC anomaly (Fig. S5a) is near Yuzhny Island and Severny Island, the large-scale circulation pattern (Fig. S5b) resembles Ural blocking, and surface energy budget (Fig. S5) is dominated by downward LWR. However, for western BKS (Fig. S6) the large negative SIC anomaly (Fig. S6a) is over the northern Barents Sea, and the positive Z500 center is over the eastern BKS (Fig. S6b), which is different from the typical Ural blocking. THF plays a much more important role than downward LWR in the energy budget over the western region (Fig. S6f).

The differences between the eastern and western regions are due to the fact that large THF (mostly SHF) can only reach the western region (Figs. S5d and S8d; see also Fig. 3), while downward LWR consistently covers the entire BKS region (Figs. S5c and S8c; see also Fig. 3). It appears that effective low-level temperature advection can only reach the western region, due to the advection of warm air from the North Atlantic (Figs. S5e and S8e; see also Fig. 3), possibly because of the cold surface temperature over the eastern part of the BKS region, which could give rise to an inversion that prevents the warm air from reaching the surface. When considering the BKS as a whole, the sea ice loss (Fig. 2e), is dominated by the sea ice loss events over the western region (Fig. S6a), with the maximum sea ice loss over northern Barents Sea. Therefore, over the entire BKS region, THF plays a more important role than downward LWR during extreme sea ice loss events.

However, it is not entirely clear how THF contributes to the melting of sea ice during these events. A warmer atmosphere above the sea ice can contribute directly to the melting of sea ice through heat exchange with the atmosphere; in addition, it can also contribute to sea ice melting through indirectly warming the ocean water surrounding the sea ice. It is also not clear whether SIC would modulate the amplitude of THF anomaly. During the extreme sea ice loss events, THF contributes more over grid points with lower SIC than those with higher SIC (not shown), possibly because 1) ocean water is more interactive with the atmosphere than sea ice and 2) lower SIC could reduce low-level inversion which leads to larger THF anomaly. These processes should be explored further in the future.

While in a composite sense both downward LWR and SHF are equally important to extreme sea ice loss events, it is interesting to ask the following question: how essential are the downward SHF and moisture intrusions for individual extreme sea ice loss events? The probability density functions (PDFs) of SIC, column water vapor, SHF, LHF, and downward LWR associated with the events are shown in Fig. 5 respectively in blue bars, compared with winter climatological PDF in red bars. The blue bars for SIC are a 9-day average centered at the local minimum of SIC during each event. As expected, during extreme sea ice loss events the PDF shifts to the left compared with the climatological 9-day averaged SIC PDF. All the loss events have at least a negative 3% anomaly of SIC. The column water vapor and downward LWR is averaged from days 4 to 4, and downward SHF as well as LHF is averaged from days 6 to 2 to capture the maximum anomalies associated with the events. All events are associated with above normal or near normal column water vapor, downward SHF, and downward LHF. This further suggests that sensible heat flux associated with positive temperature advection and downward LWR associated with moisture intrusion are both essential for sea ice loss events to occur.

4. Large-scale atmospheric circulation prior to the events

a. All BKS extreme sea ice loss events

Next, we examine the large-scale atmospheric circulation precursors to the BKS extreme sea ice loss events by compositing 300-hPa streamfunction (SF300) anomalies from day 14 to day 0 with a 2-day interval (Fig. 6). The wave activity fluxes (Takaya and Nakamura 2001) at 300 hPa, which
indicate the propagation of the Rossby waves, are shown in vectors on top of the SF300 anomalies. Tropical OLR anomalies are plotted in the subpanels. As shown in Fig. 2, there is a clear dipole pattern of SF300 over the high-latitude North Atlantic that is established from day −24 to day 0 (Figs. 6f–h). On day −28 (Fig. 6d), a Rossby wave train propagates from the North Pacific to North America. At day −6 (Fig. 6e), the wave train propagates northward into North America and then into the North Atlantic and gives rise to the positive SF300 anomaly over the midlatitude North Atlantic and negative SF300 anomaly over Canada. From days −4 to 0, the wave train propagates farther northeastward over the North Atlantic and forms the dipole pattern that is essential to the northward heat and moisture transport associated with extreme sea ice loss events (the pattern also resembles a positive NAO, which has been found to be associated with extreme sea ice loss events over BKS; Lu et al. 2016; Luo et al. 2017; Gong and Luo 2017; Gimeno et al. 2019; Yang and Magnusdottir 2017). While the cause of the wave train originally seen at day −8 is not entirely clear, there are some interesting tropical heating signals characterized by a dipole pattern of convective anomalies with enhanced convection over the Maritime Continent and tropical western Pacific and suppressed convection over the tropical central Pacific from days −14 to −8. This is consistent with previous studies (H. Park et al. 2015; Flournoy et al. 2016) regarding the role of tropical convection and moisture intrusions into the Arctic. The enhanced convective anomalies propagate eastward from the Maritime Continent (days −14 to −12; Figs. 6a,b) to the central and western Pacific (days −2 to 0; Figs. 6g,h), which resembles the evolution of the MJO. Note that the interannual variability that could influence the tropical convective anomalies and extratropical circulation has been removed from these composites. However, the composites in Fig. 6 are not sensitive to the removal of interannual variability, as the results with interannual variability included (without subtracting seasonal averages; Fig. S7) are very similar to Fig. 6.
Are extreme sea ice loss events over the BKS region related to the MJO? To address this question, we show first the evolution of the RMM index phase and amplitude associated with the 25 extreme sea ice loss events (days 215 to 21) in Fig. 7a. Each line represents an individual sea ice loss event. The lines in Fig. 7a are concentrated on the right half of the plot (phases 3–7). In particular, from days 29 to 25 (green colors), most of the events are in RMM phases 4–7, when the convective anomalies are over the Maritime Continent and the tropical western Pacific. Is this connection with MJO statistically significant? We averaged the RMM index (RMM1 and RMM2) in the 25 extreme sea ice loss events on each day, and then calculated the amplitude of the averaged RMM index \( \sqrt{\text{RMM1}^2 + \text{RMM2}^2} \). The phase and amplitude of the averaged RMM index for extreme sea ice loss events are shown in Table 1. A Monte Carlo bootstrapping using random winter days is applied to test the significance of the RMM amplitude for days corresponding to the sea ice loss events. The amplitude is significant at 5% level from days 210 to day 0, when the mean RMM is mostly in phases 5–7. The results confirm that extreme sea ice loss events are significantly connected with the MJO. Jiang et al. (2021) also noted the connection between the RMM index
and extreme BKS sea ice loss, although their emphasis is on the atmospheric response to sea ice loss.

Although extreme sea ice loss events are significantly connected with the MJO in a composite sense, it is worth noting that an MJO event does not always trigger extreme sea ice loss. In Fig. 7b, the evolution of the RMM index during all 30 winter seasons is plotted in the gray lines. Only a fraction of the MJO events in phases 4–7 are connected with sea ice loss. We further investigate the connection between MJO and sea ice losses by making composites of SIC anomalies during different RMM phases and lag times. The BKS SIC anomalies during different RMM phases when the amplitude of the RMM index is greater than 1 for the 30 winter seasons are shown in Fig. 7c. The lags here are defined as the number of days after a day when RMM index amplitude is larger than 1 in a specific RMM phase. Significant negative SIC anomalies can be seen about 5–10 days after RMM phases 5–6, consistent with Fig. 7a and Table 1. However, the composite SIC associated with MJO phases 5 and 6 (Fig. 7c) only amounts to a decrease of sea ice by less than 1.5%, while the averaged peak of extreme sea ice loss events is about 7% (Fig. 4a). Therefore, MJO phases 5–6 may not be the only trigger for extreme sea ice loss in the BKS. Note that our results in Fig. 7c are similar to the findings in Henderson et al. (2014).

b. Active and non-active MJO events

There are a few events with small RMM amplitudes (Fig. 7a) even though the mean MJO signal is significant. So, we further separate extreme sea ice loss events into two groups, one associated with an active MJO and the other with a non-active MJO. The non-active MJO group is defined as those sea ice loss events with a 3-day running averaged RMM index amplitude smaller than 1 for more than 5 days from day −15 to day 0. The 3-day running average is applied to smooth the MJO index. The rest of the sea ice loss events are considered to be associated with an active MJO. The evolution of the RMM index from the two groups is shown in Fig. 7b, with active MJO events in solid lines (16 events) and non-active MJO events in dashed lines (9 events). The active MJO events generally have large RMM amplitudes and go through phases 4–7 during day −15 to day 0. The non-active MJO events are clustered at the center of the figure with the RMM amplitude smaller or close to 1 during this time period.
As ENSO can modulate the MJO extratropical teleconnections (e.g., Moon et al. 2011; Lee and Woolnough 2019), the connection between MJO and extreme BKS sea ice loss events may vary in different ENSO conditions. Table 2 lists the ONI in each winter season with each extreme sea ice loss event. El Niño winters are defined as the winters with ONI above 0.5°C, whereas La Niña winters are when ONI is below −0.5°C, for at least 5 consecutive months, following the NOAA ENSO definition. The rest of the winters are considered ENSO neutral. The active MJO sea ice loss events can happen in any ENSO conditions (El Niño, La Niña, or neutral), with the averaged ONI leaning toward El Niño. Within the nine non-active MJO sea ice loss events, five of them happen during La Niña winters, and no events happen during El Niño winters. Thus, MJO events do not seem to be necessary to trigger extreme sea ice loss events during La Niña or non–El Niño winters. Note that our findings here are different from J. Wang et al. (2020), who found no significant connection between summer rapid Arctic sea ice loss events and the tropics, which may be because summer teleconnections induced by tropical (MJO or ENSO) convective anomalies are weaker compared to winter (e.g., Jenney et al. 2019; Yang and Delsole 2012). The evolution of extratropical large-scale circulation (SF300) and tropical convective anomalies associated with active and non-active MJO events will be discussed in the following subsections.

1) ACTIVE MJO EVENTS

The composites of active MJO sea ice loss events are shown in Fig. 8. Since active MJO events can happen in all ENSO conditions, the composites including interannual variability (Fig. S8) are similar to Fig. 8, except for a weak El Niño signal in the tropics (consistent with the mean ONI in Table 2). Note that the significance test in these composites, particularly to reject the null hypothesis that extreme sea ice loss events are not connected to midlatitude Rossby wave trains, is based on a relatively small sample size (16 for active MJO events and 9 for non-active MJO events). The tropical signals in active MJO events (Fig. 8) are similar to that in all sea ice loss events (Fig. 6), with a larger amplitude. The convective anomalies highlight the dipole pattern with enhanced convection over Maritime Continent and the western Pacific, and the suppressed convection over the central to western Pacific (MJO phases 5–6), when the subtropical streamfunction anomalies over the North Pacific are developing (days −12 to −8). The main features of the midlatitude circulation in Fig. 8 are similar to those in Fig. 6, as synoptic-scale Rossby wave trains propagate from the North Pacific, across North America and into the North Atlantic from days −8 to 0, which give rise to the dipole circulation pattern near the BKS region. This suggests that much of the signal in Fig. 6 is contributed by the MJO-related tropical heating. At lag zero (Fig. 8h), there is a strong positive SF300 anomaly over the BKS region, which resembles a blocking pattern. This connection between MJO phases 5–6 and the blocking pattern near the BKS region has been noted in previous studies (Cassou 2008; Henderson et al. 2016). Our results are consistent with one of the mechanisms proposed by Cassou (2008) that MJO phases 5–6 give rise to a positive SF300 anomaly over the west coast of North America (days −8 to −6; Figs. 8d,e), and then the Rossby wave train propagates eastward and northward into northern Europe and forms a blocking pattern.
2) **NON-ACTIVE MJO EVENTS**

For non-active MJO events (Fig. 9), the composites are made without removing interannual variability to include the impact of La Niña, as more than half of the non-active MJO events are in La Niña winters (Table 2). Note that the evolution of the extratropical circulation is similar if interannual variability is removed (Fig. S9), but the composites without removing interannual variability (Fig. 9) highlight the La Niña impact in the tropics. A strong dipole pattern of SF300 anomalies near the BKS region can be seen at day zero (Fig. 9h), which is consistent with Figs. 2 and 6. A Rossby wave train that develops over the North Pacific at day $-10$ (Fig. 9c) propagates eastward into North America and the North Atlantic, and then propagates northeastward into the BKS region during days $-8$ to $-6$. Note that although the negative streamfunction anomalies within this wave train are weak in Fig. 9, these negative signals become significant when interannual variability is removed (Fig. S9), showing that this synoptic-scale wave train is important on subseasonal time scales. The wave train gives rise to the positive SF300 anomaly over the BKS region associated with extreme sea ice loss events. During days $-6$ to $-4$, there are also some signals of northeastward propagation of Rossby waves from the North Pacific into North America, which develops the negative SF300 anomaly over Canada. This negative SF300 anomaly enhances at day $-2$ and moves eastward at day 0, which helps form the dipole pattern of circulation anomalies associated with extreme sea ice loss events. In addition, this negative SF300 anomaly also reinforces the positive SF300 anomaly over the BKS region via Rossby wave propagation as shown by the wave activity flux. This reinforcement of the dipole pattern is better captured when the interannual variability is removed (Fig. S9). Note that the daily composites shown in Fig. 9 differ significantly from the seasonal mean pattern associated with a typical La Niña, due to the fact that Fig. 9 contains daily synoptic-scale waves that are averaged out in a seasonal mean response.
c. Discussion

Common features of the convective anomalies can be found in both active MJO events and non-active MJO events. For the active MJO events, the subtropical circulation anomalies over the North Pacific (negative streamfunction anomaly over the subtropical western Pacific and positive anomaly over the west coast of North America), which are part of the Rossby wave train, develop at day −8 (Fig. 8d). The associated MJO signal, which is in MJO phases 5–6 (Table 1), has a dipole pattern of convective anomalies with suppressed convection over the central to western tropical Pacific and enhanced convection over the Maritime Continent (Figs. 8a–e). This dipole pattern associated with the MJO phases 5–6 on subseasonal time scales also resembles the convective features associated with La Niña (e.g., Fig. 9). As only 4 of the 25 events are not associated with this dipole pattern of convective anomalies (MJO or La Niña; see Table 1), this pattern is nearly a necessary condition for extreme BKS sea ice loss events as it is detected significantly in both groups of events.

How does this dipole pattern of tropical convective anomalies lead to extreme BKS sea ice loss events? In the composites of both active and non-active MJO events (Figs. 8 and 9), the most prominent features are Rossby wave trains that propagate from the eastern North Pacific across North America into the North Atlantic from day −8 to day 0. These features highlight the enhanced synoptic-scale transient wave activity in the midlatitudes, but not the quasi-stationary planetary-scale waves directly forced by tropical convection. These transient features can be driven by the large-scale circulation (e.g., enhancement or shift of the midlatitude jet) which could be forced by tropical forcing. The anomalies of 300-hPa zonal wind with respect to winter climatology during days −15 to
0 of active and non-active MJO extreme sea ice loss events are shown in Figs. 10a and 10b. Both composites show an enhancement and a northward shift of the jet at the exit region of the Pacific jet (eastern North Pacific and west coast of North America), North America, and the North Atlantic (150°W–30°E). This enhancement and northward shift of the jet corresponds well with the transient synoptic-scale Rossby wave features from the eastern North Pacific to the North Atlantic in Figs. 8 and 9, suggesting that the synoptic-scale Rossby wave trains that connect the North Pacific and the BKS sea ice loss are a result of enhanced storm track activity related to driving the enhancement and northward shift of the jet (e.g., Zheng et al. 2018). Many of the jet anomalies can be attributed to the tropical forcing. The composite of 300-hPa zonal wind of 0–15 days after all days that the MJO is in phase 5 with an amplitude larger than 1 is shown in Fig. 10c. Note that from Table 1, the period 0–15 days after MJO phase 5 corresponds to about days −15 to 0 of BKS extreme sea ice loss events (Fig. 10a). The northward shift of the jet from the eastern North Pacific to the North Atlantic (150°W–30°E) is very significant when compositing all days associated with MJO phase 5 (Fig. 10c), showing that many of the jet anomalies during the BKS sea ice loss events with active MJO (Fig. 10a) are forced by the tropical convection associated with MJO phase 5. The composite of the jet during all days in La Niña winter is shown in Fig. 10d. Consistent with other panels, there is a northward shift of the jet over the eastern North Pacific. However, the northward shift of the jet is less clear over the North Atlantic.

Note that this dipole pattern of tropical convective anomalies is not sufficient to generate extreme BKS sea ice loss events. As ENSO is a phenomenon on interannual time scales, this dipole pattern of convective anomalies should exist during most of the time in La Niña winters. However, extreme sea ice loss events are not significantly more frequent during strong La Niña winters, as there are 8 events in La Niña winters, 7 events in El Niño winters, and 10 events in ENSO neutral conditions. Additionally, no extreme sea ice loss event happened during the 1998/99 winter, which is a very strong La Niña winter. One possible reason is that La Niña forces a northward shift of the jet over the eastern North Pacific, but does not force the northward shift of the jet over eastern North America and the North Atlantic (Fig. 10d). From Figs. 10a and 9b, an enhancement and a northward shift of the jet over the North Atlantic are connected with both active MJO and non-active MJO sea ice loss events, which drives the synoptic-scale wave activity that leads to extreme BKS sea ice loss. As discussed in section 4a, for all the MJO events that propagate from phases 4 to 7 (associated with this dipole convective pattern), only a few are associated with extreme BKS sea ice losses (Fig. 7b). Note that previous studies (Cassou 2008; Henderson et al. 2016) found that blocking frequency is significantly increased over northern Europe (similar to Fig. 8h) after MJO phase 6. However, though this connection is significant, blocking over northern Europe (or a Scandinavian blocking weather regime) can only be detected in about 35%–40% of the cases after MJO phase 6 (Henderson et al. 2016; Cassou 2008), meaning that the blocking pattern does not always emerge after MJO phase 6. Therefore, it is not surprising that only a few of the MJO events are associated with extreme sea ice losses in Fig. 7b. The circulation pattern associated with extreme sea ice loss events in Fig. 10 resembles a positive NAO pattern, consistent with previous studies (e.g., Gong and Luo 2017; Luo et al. 2017). The positive NAO can be forced by not only tropical heating (e.g., Cassou 2008; Lin et al. 2009) but also extratropical variability such as anticyclonic Rossby wave breaking (e.g., Rivière and Orlanski 2007), suggesting that both tropical and extratropical forcing could modulate the extreme BKS sea ice loss events.

The discussion above suggests that, in addition to the dipole tropical convection anomalies, there must be other factors that modulate the connection between the tropical convection and extreme BKS sea ice loss events. We argue that internal variability of the extratropical circulation, which is not necessarily forced by the tropics, likely plays a crucial role in modulating the propagation of the synoptic Rossby waves train that
originates from the North Pacific and induces the dipole flow pattern near the BKS region (Figs. 8 and 9). Early studies have shown that the extratropical basic state can modulate the propagation of the Rossby waves (Hoskins and Ambrizzi 1993; Ting and Sardeshmukh 1993). More recent studies have also shown that the extratropical basic state modulates the teleconnection patterns induced by the MJO (Roundy et al. 2010; Henderson et al. 2017; Lin and Brunet 2018; Zheng and Chang 2020; J. Wang et al. 2020). Henderson et al. (2016) also suggests that the connection between northern Europe block and the evolution of the MJO convective anomalies prior to phases 5–6 is it worthwhile to further explore the factors that could modulate the connection between tropical convection and BKS extreme sea ice loss in future studies.

5. Conclusions

In this study, we analyze the surface energy flux and the large-scale circulation associated with extreme sea ice loss events over the Barents–Kara Seas (BKS) during winter. We identified 25 extreme sea ice loss events during the 30-yr period. The composites of the 25 events show that these extreme sea ice loss events are associated with strong moisture intrusions in the BKS region, which is consistent with previous studies that emphasize the role of enhanced downward longwave radiation (LWR) associated with moisture intrusions. However, the composites of surface energy flux suggest that a downward LWR anomaly is not the only important factor to drive sea ice loss during these events, as the amplitude of the sensible heat flux anomaly is comparable to that of the downward LWR anomaly, and the SHF leads downward LWR by a few days. The role of turbulent heat flux associated with sea ice loss events has not received much attention in previous studies, and we highlight this point in our analysis, showing that THF is as important as downward LWR. The reason why SHF leads downward LWR is that the SHF anomaly is a direct response to atmospheric heat transport, while the downward LWR anomaly is related to total column water vapor, which takes a few extra days to build up after increased atmospheric moisture transport.

The composites of large-scale circulation prior to BKS extreme sea ice loss events show that these events on average are significantly connected with the Madden–Julian oscillation (MJO). More specifically, they are related to a dipole pattern of tropical convective anomalies with enhanced convection over the Maritime Continent and suppressed convection over the central to western tropical Pacific. The 25 extreme sea ice loss events are further divided into two groups. For the group of events with active MJO, about 5–10 days before the events, the MJO is in phases 5–6, which have the dipole convective pattern discussed above. For the group with non-active MJO, these events tend to happen during La Niña winters, which also feature the same dipole convective pattern. This dipole pattern of tropical convective anomalies drives an enhancement and a northward shift of the jet from the eastern North Pacific to the North Atlantic. The synoptic-scale Rossby wave pattern that connects the North Pacific and BKS sea ice loss is likely a result of an enhanced and northward shift of storm track activity, which is associated with the jet anomalies. However, this dipole convective pattern itself is not sufficient to give rise to BKS extreme sea ice loss events, since many MJO events that propagate through phase 5–6 do not induce extreme sea ice loss events, and there is no significant increase of extreme sea ice loss events during La Niña winters. We hypothesize that variability in extratropical background flow, likely an enhanced and northward shifted jet over the North Pacific to North Atlantic, could modulate the connection between the tropical dipole convective anomaly and extreme sea ice loss in the BKS. We will further investigate this hypothesis in future studies.

Acknowledgments. We thank four anonymous reviewers for their constructive comments that improved the clarity of the manuscript. This work is supported by NASA Award 80NSSC20K1254. CZ, MT, and YW are also supported by NSF Award OPP-1825858. We thank Dr. Linette Boisvert for sharing the sensible and latent heat flux derived from AIRS data. The RMM index data are provided by the Bureau of Meteorology website (http://www.bom.gov.au/climate/mjo/). The Niño-3.4 index and ONI are available at https://climatedataguide.ucar.edu/climate-data/nino-sst-indices-nino-12-3-34-4-oni-and-tni.

REFERENCES


