Atmospheric Energy Sources for Winter Sea Ice Variability over the North Barents–Kara Seas

YINGLIN TIAN, a YU ZHANG, a DEYU ZHONG, a,b MINGXI ZHANG, a TIEJIAN LI, a DI XIE, a and GUANGQIAN WANG a,b

State Key Laboratory of Hydroscience and Engineering, Department of Hydraulic Engineering, Tsinghua University, Beijing, China
Joint-Sponsored State Key Laboratory of Plateau Ecology and Agriculture, School of Water Resources and Electric Power, Qinghai University, Xining, China

ABSTRACT: Anomalous poleward transport of atmospheric energy can lead to sea ice loss during boreal winter over the Arctic, especially in the North Barents–Kara Seas (NBKS), by strengthening downward longwave radiation (DLW). However, compared with the extensive studies of latent energy sources, those of sensible energy sources are currently insufficient. Therefore, we focus on the intraseasonal sea ice loss events from the perspectives of both energy forms. First, the contributions of latent and sensible energy to DLW and sea ice reduction are quantified using the lagged composite method, a multiple linear regression model, and an ice toy model. Second, a Lagrangian approach is performed to examine sources of latent and sensible energy. Third, possible underlying mechanisms are proposed. We find that the positive anomalies of latent and sensible energy account for approximately 56% and 28% of the increase in DLW, respectively, and the DLW anomalies can theoretically explain a maximum of 58% of sea ice reduction. Geographically, the North Atlantic, the Norwegian, North, and Baltic Seas, western Europe, and the northeastern Pacific are major atmospheric energy source regions. Additionally, while the contributions of latent energy sources decrease with increasing distance from the NBKS, those of sensible energy sources are concentrated in the midlatitudes. Mechanistically, latent energy can influence sea ice decline, both directly by increasing the Arctic precipitable water and indirectly by warming the Arctic atmosphere through remote conversion into sensible energy. Our results indicate that the Rossby waves induced by latent heating over the western tropical Pacific contribute to anomalous energy sources at midlatitude Pacific and Atlantic both dynamically and thermodynamically.

SIGNIFICANCE STATEMENT: Winter sea ice retreat in the Arctic has been attributed to increasing poleward atmospheric energy transport. While latent energy sources are extensively examined in previous studies, studies on sensible energy sources remain limited. Considering both atmospheric energy forms, we detected energy sources for the intraseasonal sea ice-loss events in the winter NBKS. Geographically, the North Atlantic, the Norwegian, North, and Baltic Seas, western Europe, and the northeastern Pacific are predominant energy source regions. Mechanistically, Rossby waves in the Northern Hemisphere triggered by tropical latent heating contribute to warm and moist air intrusions into the Arctic. This work suggests that latent energy can impact Arctic sea ice directly by moistening the atmosphere and indirectly by warming the Arctic atmosphere through remote conversion into sensible energy.

KEYWORDS: Arctic; Sea ice; Energy transport; Lagrangian circulation/transport

1. Introduction

In recent decades, Arctic sea ice has undergone a dramatic decline in several aspects, including the sea ice concentration (SIC; Deser and Teng 2008), area (Comiso et al. 2017), extent (Simmonds 2015), thickness (SIT; Lindsay and Schweiger 2015), and volume (Laxon et al. 2013). Recent studies show that sea ice loss in the Arctic is largely associated with the anomalous transport of latent energy (LE; specific humidity multiplied by the latent heat of vaporization) and sensible energy (SE; atmospheric temperature multiplied by the specific heat of dry air under constant pressure) transported into the Arctic (H.-S. Park et al. 2015a; Woods and Caballero 2016; Pithan et al. 2018; Graham et al. 2019). Moreover, the meridional transport of LE and SE responsible for Arctic sea ice changes occurs mostly as anomalous synoptic pulses of warm and moist air intrusions (WaMAIs) from lower latitudes (Pithan et al. 2018). Therefore, to better understand and forecast the Arctic climate, the anomalous atmospheric energy carried by these WaMAIs has been extensively investigated in previous studies, especially by virtue of the Lagrangian approach. It has recently been proposed that the Lagrangian approach is more capable than the Eulerian method of describing the processes acting on air masses as they are transported (Komatsu et al. 2018; Pithan et al. 2018; You et al. 2021).

WaMAIs carry anomalous atmospheric energy, for which the Lagrangian approach has achieved significant advances...
with respect to its influences, origins, and transformations in transit. First, for the impacts of the anomalous energy carried by the WaMAIs into the Arctic, it has been shown that warm and moist air can emit strong downward longwave radiation (DLW), thus leading to anomalous surface warming and sea ice retreat in the Arctic (Doyle et al. 2011; D.-S. R. Park et al. 2015; H.-S. Park et al. 2015a,b; Kapsch et al. 2016; Woods and Caballero 2016; Persson et al. 2017). This is largely because moisture is a major contributor to the greenhouse effect (Held and Soden 2000), and the DLW is proportional to the fourth power of air temperature (Sridhar and Elliott 2002). Additionally, the effective longwave emissivity of the atmosphere is influenced by both the moisture (and thus the LE) and air temperature (and thus the SE). Second, regarding the energy sources and sinks along the WaMAIs into the Arctic, which correspond to the locations where atmospheric energy increases and decreases, respectively, Zhong et al. (2018) concluded that LE sources (or moisture sources) outside the Arctic, particularly in the Norwegian Sea and midlatitude Atlantic, make a major contribution to the intensified DLW, sea ice reduction, and concurrent surface warming in the wintertime Barents–Kara Seas (BKS). Third, regarding the air-mass transformation along with the transport, Ali and Pithan (2020) compiled Eulerian-based observations into a Lagrangian framework and provided direct observed evidence of the cloudy Arctic boundary layer created by WaMAIs. Dimitrelos et al. (2020) indicated that radiative cooling at the surface promoted cloud formation, the lifetime of which is sensitive to changes in moisture entrainment at the cloud top.

Nevertheless, a large part of studies on the poleward transport of atmospheric energy carried by WaMAIs are focused on the effects of LE, while those of SE have not been fully discussed. First, it was reported that the horizontal convergence of the LE flux is responsible for approximately 40%–50% of the increase in the DLW, and the remaining proportion is inferred to be partly attributed to the SE flux convergence (Woods et al. 2013; H.-S. Park et al. 2015b; Gong et al. 2017). More directly, the increase in the poleward SE flux is observed to be closely related to the polar amplification and melting of the Arctic sea ice (Lu and Cai 2009; H.-S. Park et al. 2015a,b; Alekseev et al. 2019). Therefore, to better understand the Arctic climate variation, it is of great significance to further quantify the individual contribution of the SE transport to the intraseasonal variations in the Arctic DLW and sea ice. Second, apart from the influence of the anomalously poleward SE transport, SE origins also require further investigation. Traditionally, the variations in SE have mostly been studied based on the Eulerian-based diabatic heat budget (Ueda et al. 2003; Wright and Fueglistaler 2013; Yasunaga et al. 2019); however, it is difficult to explicitly detect the geographical SE source for a designated energy sink, thus implying the necessity of the Lagrangian diagnosis for the SE source. Third, when combining the energy budget with the moisture balance equation, previous studies have proposed that latent heat release can serve as both the LE sink and SE source (Yanai et al. 1973; Wei et al. 1983; Yanai and Tomita 1998). This suggests that the LE can possibly influence the Arctic system in two ways, directly by humidifying the Arctic air and indirectly by heating the Arctic atmosphere through the remote conversion into SE. Accordingly, the conversion between LE and SE along WaMAIs also warrants further examination.

Furthermore, poleward propagation of atmospheric energy during the winter occurs mostly in filamentary structures along with some specific circulation patterns resembling the El Niño–Southern Oscillation (ENSO), Pacific–North American pattern (PNA), North Atlantic Oscillation (NAO), Arctic Oscillation, and Antarctic Oscillation patterns (Huang et al. 2012), and other large-scale patterns. Among these structures, a typical one is comprised of a cyclone embedded over the Canadian/Arctic archipelago/Greenland region accompanied by a high pressure system to its east, which is like the positive phase of NAO (NAO+) with a Ural blocking located at its eastern flank (NAO+ with the Ural blocking–like pattern; Luo et al. 2016a,b; Gong and Luo 2017; Luo et al. 2017). Moreover, this distinct circulation pattern tends to be preceded by enhanced tropical convection over the Indian and western Pacific Oceans (Lee et al. 2011; Yoo et al. 2011, 2012a,b). It has been observed that the tropical latent heating may have excited the quasi-stationary Rossby wave train in the midlatitudes, which could further extend to the Arctic when the zonal asymmetries are simultaneously enhanced by the extratropical heating (Goss et al. 2016; Park and Lee 2019). In addition, the downstream dispersion of synoptic wave energy (Drouard et al. 2015) and the strengthened stratospheric polar vortex (Jiang et al. 2017) may also have played a role in the teleconnections between the tropics, midlatitudes, and polar region. According to these findings, the underlying mechanisms of the anomalous energy transport from the midlatitudes to the Arctic should be further investigated through analysis of the associated tropical latent heating, Rossby waves, and large-scale circulation patterns.

In this context, this research pursues three goals: 1) to quantify the relationship among the anomalous energy advection, strengthened DLW, and sea ice loss in the Arctic based on the lagged composite method, a multiple linear regression model, and an ice toy model; 2) to identify the LE and SE sources and clarify the associated conversion between these two forms of energy based on the Lagrangian energy source detection method; and 3) to examine the possible mechanism generating the anomalous energy sources that drive the intraseasonal sea ice loss based on an analysis of the large-scale circulation, planetary waves, and latent heating. In particular, previous studies have revealed that it is necessary to focus on the wintertime North Barents–Kara Seas (NBKS) in the Arctic because the NBKS region shows the most significant sea ice loss, a high frequency of moisture injection, and close interactions between the hydrological cycle and energy budget (Cavaliere and Parkinson 2012; Parkinson 2014; Onarheim et al. 2015; Woods and Caballero 2016; King et al. 2017); consequently, we chose this region for our study.

The remainder of this paper is organized as follows: section 2 briefly introduces the data, study area, and methodology. Section 3 presents the results. The findings are discussed in section 4, and concluding remarks are presented in section 5.
2. Data, study area, and methodology

a. Data and study area

In this study, we used the European Centre for Medium-Range Weather Forecasts (ECMWF) interim reanalysis (ERA-Interim, herein ERAI; Dee et al. 2011) data for daily fields of atmospheric and oceanic variables for the December–February (DJF) season for 1979–2018. The 2D variables are the DLW ($F_{dl}$), surface net solar radiation ($F_{ns}$), surface turbulent heat flux [$F_{th}$, composed of the surface latent heat flux ($F_{lh}$) and surface sensible heat flux ($F_{sh}$)], SIC, and zonal and meridional 10-m wind. The terms $F_{dl}$ and $F_{ns}$ are defined as being positive downward, while $F_{th}$, $F_{lh}$, and $F_{sh}$ are defined as being positive upward. The 3D variables are the zonal, meridional, and vertical wind, temperature, specific humidity, geopotential height, and relative humidity from 1000 to 50 hPa in vertical intervals of 50 hPa. Moreover, the following variables are calculated, as described in Appendix A: streamfunction ($\psi$), vertical integrals of SE ($E_v$), eastward and northward SE fluxes ($I_{se}$ and $J_{se}$, respectively), SE flux convergence ($F_{con}$), LE ($E_{s}$), eastward and northward LE fluxes ($I_{le}$ and $J_{le}$, respectively), and LE flux convergence ($F_{con}$). To estimate the response of sea ice to DLW variation, sea ice thickness (SIT; $h$) from the Coupled Pan-Arctic Ice-Ocean Modeling and Assimilation System (PIOMAS; Zhang and Rothrock 2003) is used. In addition, diabatic heating data from Modern-Era Retrospective Analysis for Research and Applications (MERRA; Rienecker et al. 2011) and MERRA version 2 (MERRA-2; Gelaro et al. 2017) are used, including the diabatic heating components due to radiation ($Q_{rad}$), vertically turbulent mixing ($Q_{tur}$), moist physics ($Q_{m}$), and their sum ($Q$).

This study mainly focuses on the variation in the ice coverage in the NBKS. Figure 1 presents the composite of the SIC anomalies when the Arctic-averaged SIC is less than the 10th percentile of all the daily Arctic-averaged SIC series during 1979–2018 DJF. (b) Change rates of year-mean SIC anomalies during 1979–2018 DJF. Shading is plotted with statistical significance at the 95% confidence level. The NBKS and the BKS are bounded by the red and black solid polygons, respectively.

b. Methodology

1) ARCTIC SEA ICE EVENT DEFINITION AND LAGGED COMPOSITE ANALYSIS

In the selection of SIC decline (SIC−) events and moderate SIC (SIC=) events, seasonal cycles and trends were removed from all fields and a 3-day moving average was applied. Second, the SIC− days were detected by selecting the days on which the NBKS-mean SIC was less than the 10th percentile of the SIC series (NBKS-mean SIC in DJF 1979–2018). The SIC− days were selected similarly but limited to the 45th and 55th percentiles. The numbers of days satisfying these criteria are shown in Table 1. Third, the initial SIC− (SIC=) events were selected, which were defined as periods containing one or more consecutive SIC− (SIC=) days. Finally, valid
SIC− (SIC=) events were obtained by maintaining a separation of at least 7 days between two adjacent initial events to ensure that the selected events were independent of one another. The numbers of initial and valid SIC events during DJF 1979–2018 are shown in Table 1, and the basic characteristics of the SIC events are presented in the online supplemental material (see Text S1 and Fig. S1 therein).

Lagged composite analysis (H.-S. Park et al. 2015a) was performed to clearly illustrate the interaction between the atmosphere and surface during the SIC− events over the NBKS, with lag day 0 defined as the onset day of the sea ice event. All the fields were deseasonalized, detrended as Chen and Zhai (2017) suggest, and 3-day moving averaged.

### 2) ATTRIBUTION OF DLW ANOMALIES AND THEIR IMPACTS ON SEA ICE VARIABILITY

Following the work of D.-S. R. Park et al. (2015) and Alekseev et al. (2019), the variability in the DLW can be expressed as follows:

$$1 = \frac{\partial F_{dl}}{\partial E_i} \left( \frac{\delta E_i}{\delta F_{dl}} \right) + \frac{\partial F_{dl}}{\partial E_s} \left( \frac{\delta E_s}{\delta F_{dl}} \right) + r,$$

where $\delta$ represents the daily anomaly; the overbar ($\bar{\cdot}$) denotes the average of all samples; $\partial F_{dl}/\partial E_i$ (or $\partial F_{dl}/\partial E_s$) denotes the partial regression coefficients of $\delta E_i$ (or $\delta E_s$), which indicates the change in $\delta F_{dl}$ per unit change in $\delta E_i$ (or $\delta E_s$), with $\delta E_i$ (or $\delta E_s$) being constant; and $r$ is the residual. The derivation and details of Eq. (1) are presented in appendix B.

The first two terms on the right-hand side of Eq. (1) represent the relative contributions of $\delta E_i$ and $\delta E_s$ to $\delta F_{dl}$, respectively. In addition, to estimate the difference in the effectiveness of $\delta E_s$ and $\delta E_i$ on $\delta F_{dl}$, a new coefficient is defined as

$$C_{nor} = \frac{\partial F_{dl}}{\partial E_i} \left( \frac{\delta E_i}{\delta F_{dl}} \right)^{-1},$$

where $C_{nor}$ is the normalized regression coefficient between $\delta F_{dl}$ and $\delta E_i$.

The impact of DLW anomalies on the change in the SIT is quantified as follows (Eisenman and Wettlaufer 2009):

$$\Delta h_{dl} = -\frac{\epsilon}{\rho_{ic} L_f} \sum_{i=t_0}^{t_k} \delta F_{dl,i},$$

where $t_0$ and $t_k$ are the start and end days, respectively; $\Delta h_{dl}$ is the cumulative change in the SIT from $t_0$ to $t_k$ caused by the DLW change; $\delta F_{dl,i}$ is the daily DLW anomaly; $\rho_{ic}$ is the ice density; $L_f$ is the latent heat of fusion; and $\epsilon = 1 - [1/(1 + \gamma h_0)]$ is the sensitivity of ice to the heat flux, which depends on the initial ice condition ($h_0$) and thermodynamic scale thickness ($\gamma = 0.7$ m, as in Eisenman 2012). When quantifying the individual contribution from anomalies of SE (or LE) to SIT changes, $\delta F_{dl}$ is replaced by $(\partial F_{dl}/\partial E_i)\delta E_{i,t}$ or $(\partial F_{dl}/\partial E_s)\delta E_{s,t}$.

### 3) DIAGNOSIS OF ATMOSPHERIC ENERGY SOURCES

The examination of the atmospheric energy sources includes the following six steps. The core idea is to identify the connection between the sources and sinks of LE (diabatic SE) based on sequential specific humidify data (diabatic heating) along the air trajectories.

(i) Generating the backward Lagrangian trajectories

The first step is to identify the atmospheric energy sources is to generate the backward Lagrangian trajectories. In this paper, the Hybrid Single-Particle Lagrangian Integrated Trajectory model (HYSPLIT) model (Draxier and Hess 1998) is chosen to trace the backward air particle trajectories. The length of time to trace back is one of the key parameters of the Lagrangian analysis, and its determination is described in the online supplemental material (Text S2 and Fig. S2). As shown in Fig. S2, air particle tracing for as long as 12 days is sufficient to represent the related atmospheric energy variation. Detailed information regarding the Lagrangian trajectory generation is presented in appendix C, and the total trajectory amounts released for different SIC events are provided in Table 1.

(ii) Calculating changes in LE and SE

Second, changes in the LE and SE along the trajectories are calculated using the following equations:

$$\frac{de_i}{dt} = \frac{L_e dq}{dt},$$

$$\frac{de_s}{dt} = \frac{\epsilon}{\rho_{ic} L_f} \sum_{i=t_0}^{t_k} \delta F_{dl,i},$$

where $e_i = L_e dq$ denotes the LE that an air particle carries while moving to a specific location; $e_s = \rho_{ic} (\theta - \theta_{min})$ denotes the renewable SE that the air particle carries while moving to a specific location, where $\theta_{min} = 214$ K represents the climatological wintertime minimum of $\theta$ in the global troposphere during 1979–2018, and the definition and calculation of $e_s$ are presented in the online supplemental material (Text S3 and Fig. S3); $de_i$ denotes the variation in the LE; $de_s$ denotes the variation in the SE due to diabatic heating; and $dt$ is the time interval, which is set as a constant value of 3 h and dropped in

<table>
<thead>
<tr>
<th>Percentiles</th>
<th>No. of days</th>
<th>No. of initial events</th>
<th>No. of valid events</th>
<th>No. of trajectories</th>
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</thead>
<tbody>
<tr>
<td>0–10th</td>
<td>352</td>
<td>72</td>
<td>65</td>
<td>365,518</td>
</tr>
<tr>
<td>0–5th</td>
<td>176</td>
<td>40</td>
<td>39</td>
<td>219,290</td>
</tr>
<tr>
<td>0–15th</td>
<td>528</td>
<td>95</td>
<td>86</td>
<td>483,484</td>
</tr>
<tr>
<td>45th–55th</td>
<td>352</td>
<td>190</td>
<td>154</td>
<td>863,705</td>
</tr>
</tbody>
</table>
the following part for simplicity. Also, $L_v$ is the latent heat of vaporization; $dq$ is the variation in the specific humidity; $p_0 = 1000$ hPa is the reference pressure; $p$ is the pressure; $\kappa = R/c_p$ is the Poisson constant, in which $R$ is the gas constant and $c_p$ is the specific heat of dry air under constant pressure; and $Q$ is the total diabatic heating, which can be separated into three individual components pertaining to moist physics $Q_{lh}$, radiation $Q_{rad}$, and turbulent mixing $Q_{tur}$. The diurnal cycle is filtered from $Q_{rad}$ and $Q$ using the 24-h moving average.

(iii) Identifying locations of energy uptake and release

Third, locations on the trajectories are divided into energy uptake and release locations (location refers to the region to which the air particle travels to every 3 h). Locations where the LE increases ($\Delta l > 0$) and decreases ($\Delta l < 0$) are classified as sources and sinks of LE, respectively, while locations where the SE diabatically increases ($\Delta e > 0$) and decreases ($\Delta e < 0$) are classified as diabatic sources and sinks of SE, respectively. Furthermore, positive $\Delta l$ and $\Delta e$ are classified as atmospheric energy uptakes ($u_{le}$ and $u_{se}$, respectively), while the absolute values of the negative $\Delta l$ and $\Delta e$ are treated as atmospheric energy losses ($l_{le}$ and $l_{se}$, respectively).

(iv) Discounting the contribution of energy uptakes

Fourth, a discount process is performed in which $l_{le}$ is assumed to proportionally reduce the contribution of any $u_{le}$ upstream to it.

If location $n$ is an LE source, the fractional contribution of all previous $u_{le}$ to the whole $e_l$ of location $n$ is calculated as

$$F_m = \frac{u_{le,m}}{e_{l,n}} (m \leq n, \Delta e_{l,m} > 0, \Delta e_{l,n} > 0), \quad (6)$$

where $m$ and $n$ denote the locations along the trajectories and $F_m$ denotes the fractional contribution of the source location $m$. In contrast, if location $n$ is an LE sink, all previous $u_{le}$ are discounted to the net value and $F_{u_{le}}$ is calculated as

$$u'_{le,m} = u_{le,m} - l_{le,m}F_m (m < n, \Delta e_{l,m} > 0, \Delta e_{l,n} < 0), \quad (7)$$

$$F_m = \frac{u'_{le,m}}{e_{l,n}} (m < n, \Delta e_{l,m} > 0, \Delta e_{l,n} < 0), \quad (8)$$

where $u'_{le,m}$ is the net uptake of the LE at the source location $m$. After traversing all locations along the trajectories every 3 h, the vertical integral of the final $u'_{le}$ is considered the LE source ($S_{le}$).

The diabatic uptake of SE is discounted in a similar way to obtain the net $u_{se}$ ($u'_{se}$) by using the following equations:

$$f_m = \frac{u_{se,m}}{e_{s,n}} (m \leq n, \Delta e_{s,m} > 0, \Delta e_{s,n} > 0), \quad (9)$$

$$u'_{se,m} = u_{se,m} - l_{se,m}f_m (m < n, \Delta e_{s,m} > 0, \Delta e_{s,n} < 0), \quad (10)$$

$$f_m = \frac{u'_{se,m}}{e_{s,n}} (m < n, \Delta e_{s,m} > 0, \Delta e_{s,n} < 0), \quad (11)$$

where $m$ and $n$ denote the locations along the trajectories; $f_m$ denotes the $e_s$ fractional contribution of the source location $m$; and $u'_{se,m}$ is the net SE uptake of the source location $m$.

Considering the different efficiencies of the influence of the LE and SE anomalies on DLW variation, the vertical integral of $C_{net}u'_{le}$ is termed the SE source due to diabatic heating ($S_{se}$), and the sum of $S_{le}$ and $S_{se}$ is termed the atmospheric energy source $S_{ae}$.

(v) Decomposing the sensible energy source

Fifth, $S_{se}$ is decomposed into the partial $S_{ae}$ contributions of latent heat release ($S_{lh}$), radiative heating ($S_{rad}$), and turbulent heating ($S_{tur}$). The ratio among $S_{lh}$, $S_{rad}$, and $S_{tur}$ of the specific location is set constant as the ratio among $Q_{lh}$, $Q_{rad}$, and $Q_{tur}$. A sample calculation for the Lagrangian detection and decomposition of SE sources is presented in the online supplemental material (Text S4 and Fig. S4).

(vi) Calculating the anomalies of the attributed atmospheric energy sources

Finally, the anomalies of the attributed atmospheric energy sources are calculated by subtracting the event-mean value during the SIC= events of the corresponding variable.

Source detection of LE can be considered as an extension of the Lagrangian moisture source diagnostic of Sodemann et al. (2008). Thus, we validated SE detection only, as described in the online supplemental material (Text S5; Figs. S5 and S6). Definitions of terms and acronyms are provided in Table 2.

3. Results

a. Coupled air–surface variations during sea ice decline events

To gain an insight into the coupled air–surface system in the Arctic during SIC= events, the spatiotemporal variations for a series of variables, including the composites of 500-hPa geopotential height, latent energy heat flux, sensible energy flux, downward longwave radiation, and sea ice concentration anomalies, are shown in Fig. 2, and their NBKS-mean anomalies are shown in Fig. 3.

In the first column in Fig. 2, 4 days before SIC declines, a synoptic cyclone is anchored at Greenland Island (65°N, 40°W), and a blocking high occurs in the Ural Mountains (60°N, 60°E). The Greenland low and Ural high form a pair of stable cyclone–anticyclone circulation structures that transports more LE and SE poleward from the Atlantic Ocean (second and third columns in Fig. 2). Consequently, as shown in Fig. 3a, the NBKS-mean convergence of the atmospheric energy significantly increases. Meanwhile, both LE and SE at the NBKS significantly increase 2 days before the onset of SIC= events, which is accompanied by strengthened DLW. Generally, the spatial patterns of anomalous LE, SE, and DLW resemble one another (second, third, and fourth columns of Fig. 2), which is consistent with the findings reported previously (Woods et al. 2013; H.-S. Park et al. 2015a; Zhong et al. 2018).

Following the processes described above, sea ice in the NBKS began to decrease at lag day 0. Negative SIC anomalies vary with DLW anomalies in terms of the spatial structure (Fig. 2) and regional mean (Fig. 3a) with a 2-day lag, consistent with H.-S. Park et al. (2015a). In addition, the absolute
<table>
<thead>
<tr>
<th>Variable</th>
<th>Acronym</th>
<th>Unit</th>
<th>Source</th>
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<tbody>
<tr>
<td>Absolute temperature</td>
<td>$T$</td>
<td>K</td>
<td>3D variables from ERAI</td>
</tr>
<tr>
<td>Pressure</td>
<td>$p$</td>
<td>hPa</td>
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<tr>
<td>Horizontal wind velocity</td>
<td>$V$</td>
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<td>Zonal wind velocity</td>
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<td>Meridional wind velocity</td>
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<tr>
<td>Specific humidity</td>
<td>$q$</td>
<td>kg kg$^{-1}$</td>
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<td>Surface pressure</td>
<td>$p_s$</td>
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<td>Surface downward longwave radiation</td>
<td>$F_{dl}$</td>
<td>W m$^{-2}$</td>
<td></td>
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<tr>
<td>Surface turbulent heat flux</td>
<td>$F_{th}$</td>
<td>W m$^{-2}$</td>
<td></td>
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<tr>
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<td>$F_{lat}$</td>
<td>W m$^{-2}$</td>
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<td>SIC</td>
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<td>PIOMAS</td>
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<td>m</td>
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<tr>
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<td>$Q$</td>
<td>K s$^{-1}$</td>
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<td>K s$^{-1}$</td>
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<td>J kg$^{-1}$ K$^{-1}$</td>
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<td>Eastward sensible energy fluxes</td>
<td>$I_{se}$</td>
<td>J m$^{-1}$ s$^{-2}$</td>
<td></td>
</tr>
<tr>
<td>Vertical integrals of latent energy</td>
<td>$E_l$</td>
<td>J m$^{-2}$</td>
<td></td>
</tr>
<tr>
<td>Northward latent energy fluxes</td>
<td>$J_{le}$</td>
<td>J m$^{-1}$ s$^{-2}$</td>
<td></td>
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<tr>
<td>Eastward latent energy fluxes</td>
<td>$I_{le}$</td>
<td>J m$^{-1}$ s$^{-2}$</td>
<td></td>
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<tr>
<td>Latent energy flux convergence</td>
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<td>J m$^{-2}$ s$^{-1}$</td>
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<tr>
<td>Sensitivity of ice to the heat flux</td>
<td>$\epsilon$</td>
<td>—</td>
<td>Calculated in the impact quantification of $F_{dl}$ anomalies</td>
</tr>
<tr>
<td>Cumulative change in sea ice thickness</td>
<td>$\Delta h_{dl}$</td>
<td>m</td>
<td>of $F_{dl}$ anomalies</td>
</tr>
<tr>
<td>Partial regression coefficient of $\delta E_i$</td>
<td>$\partial F_{dl}/\partial E_i$</td>
<td>—</td>
<td>Calculated in the attribution estimation of $F_{dl}$ anomalies</td>
</tr>
<tr>
<td>Normalized partial regression coefficient of $\delta E_i$</td>
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<td>—</td>
<td></td>
</tr>
<tr>
<td>Relative contributions of $\delta E_i$ to $\delta F_{dl}$</td>
<td>$\partial F_{dl}/\delta E_i$</td>
<td>%</td>
<td></td>
</tr>
<tr>
<td>Relative contributions of $\delta E_i$ to $\delta F_{dl}$</td>
<td>$\partial F_{dl}/\delta E_i$</td>
<td>%</td>
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</tr>
<tr>
<td>Latent energy that the air particle carries when moving to a specific location</td>
<td>$e_l$</td>
<td>J kg$^{-1}$</td>
<td>Calculated in the diagnosis of atmospheric energy sources</td>
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<tr>
<td>Renewable sensible energy that the air particle carries when moving to a specific location</td>
<td>$e_s$</td>
<td>J kg$^{-1}$</td>
<td></td>
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<tr>
<td>Latent energy uptake</td>
<td>$u_{te}$</td>
<td>J kg$^{-1}$</td>
<td></td>
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<tr>
<td>Sensible energy uptake due to the diabatic heating</td>
<td>$u_{se}$</td>
<td>J kg$^{-1}$</td>
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</table>
value of energy fluxes (Fig. 3b) shows that the DLW anomalies dominate the surface energy budget, the regional-mean net solar radiation anomalies are not sufficient to force the SIC decrease, and the surface turbulent heat fluxes are suppressed while the atmospheric energy converges; these results were also demonstrated by Lesins et al. (2012) and Gong et al. (2017). Interestingly, as shown in the last column in Fig. 2, significant sea ice reduction also occurs around the Chukchi Sea (70°N, 180°W), Baffin Bay and the Labrador Sea (65°N, 80°W), and the Greenland Sea (75°N, 15°W), which are located in the other region with thinner sea ice cover in the winter Arctic (Fig. S7) and are thus the more sensitive to external forcing. However, the drivers of these reductions are different from those of the NBKS. In Fig. S8, while the NBKS is largely influenced by the DLW (second column in Fig. S8), the other three regions are impacted by the surface turbulent heat fluxes (first column in Fig. S8) and do not exhibit increased atmospheric energy (second and third columns in Fig. 2).

Generally, the atmospheric circulation preconditioning SIC−events favors the transportation of large amounts of LE and SE into the Arctic, originating from the midlatitude North Atlantic Ocean and passing through the Norwegian Sea and the Fram Strait between Greenland and Svalbard. These anomalous atmospheric energies are subsequently delivered to the surface by the amplified DLW, resulting in sea ice loss in the Arctic.

### b. Attribution of DLW anomalies and their impacts on sea ice variability

Since the DLW increase is essential for sea ice loss, a detailed investigation of its attribution and impact is advisable. In Fig. 3b, the NBKS-mean DLW anomalies tend to become positive from 6 days before the onset days of the SIC−events and last 13 days (from lag days −6 to 7). Therefore, the attribution of DLW anomalies and their impacts on sea ice variability are determined based on this period (hereafter the strengthened DLW period).

With Eqs. (A8) and (A9), the correlation between the daily NBKS-mean anomalies of DLW, LE, and SE can be described by a multiple linear regression model as

\[
\delta F_{dl} = 3.62\delta E_{le} + 0.47\delta E_{se} + 1.76, \quad r^2 = 0.89, \quad (12)
\]

where \(r^2 = 0.89\) indicates a strong linear dependence between \(\delta E_{le}, \delta E_{se}\), and \(\delta F_{dl}\) in the NBKS during the strengthened DLW period. Equation (12) shows that \(\delta F_{dl}\) increases linearly as \(\delta E_{le}\) or \(\delta E_{se}\) increases, implying that the DLW is intensified for delivering these anomalous energies from the atmosphere to the undersurface. Quantitatively, according to Eqs. (2) and (12), \(\delta E_{le}\) and \(\delta E_{se}\) contribute 56% \((\delta F_{dl}/\delta E_{le}/\delta F_{dl})\) and 28% \((\delta F_{dl}/\delta E_{se}/\delta F_{dl})\) to \(\delta F_{dl}\), respectively, largely consistent with the previous results. Alekseev et al. (2019) estimated the contributions of LE (40%) and SE (34%) to interannual temperature variability in the Arctic during winter; H.-S. Park et al. (2015b) and Gong et al. (2017) concluded that the LE flux convergence explains approximately 50% of the interannual DLW trends. The discrepancies might be caused by different spatial and temporal scales as well as the slight difference in the temperature and DLW changes. With Eq. (2), \(C_{nor}\) is 0.13, which represents a higher efficiency of influence of \(\delta E_{le}\) on \(\delta F_{dl}\) than that of \(\delta E_{se}\).

To further explore the role of atmospheric energy in shrinking sea ice, the cumulative SIT change during the strengthened DLW period, the portion caused by the DLW, and the individual contributions from LE and SE to SIT changes are

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**Table 2. (Continued)**

<table>
<thead>
<tr>
<th>Variable</th>
<th>Acronym</th>
<th>Unit</th>
</tr>
</thead>
<tbody>
<tr>
<td>Latent energy loss</td>
<td>(l_{le})</td>
<td>J kg(^{-1})</td>
</tr>
<tr>
<td>Sensible energy loss due to the diabatic heating</td>
<td>(l_{se})</td>
<td>J kg(^{-1})</td>
</tr>
<tr>
<td>(c), fractional contribution of location (m)</td>
<td>(F_m)</td>
<td>%</td>
</tr>
<tr>
<td>Net latent energy uptake</td>
<td>(u_{le})</td>
<td>J kg(^{-1})</td>
</tr>
<tr>
<td>(c), fractional contribution of location (m)</td>
<td>(f_m)</td>
<td>%</td>
</tr>
<tr>
<td>Net sensible energy uptake due to the diabatic heating</td>
<td>(u_{se})</td>
<td>J kg(^{-1})</td>
</tr>
<tr>
<td>Latent energy source</td>
<td>(S_{le})</td>
<td>W m(^{-2})</td>
</tr>
<tr>
<td>Sensible energy source due to the diabatic heating</td>
<td>(S_{se})</td>
<td>W m(^{-2})</td>
</tr>
<tr>
<td>Composited atmospheric energy source</td>
<td>(S_{at})</td>
<td>W m(^{-2})</td>
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<td>(S_{le})</td>
<td>W m(^{-2})</td>
</tr>
<tr>
<td>Partial (S_{se}), contributions of the radiative heating</td>
<td>(S_{rad})</td>
<td>W m(^{-2})</td>
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<tr>
<td>Partial (S_{se}), contributions of the turbulent heating</td>
<td>(S_{dur})</td>
<td>W m(^{-2})</td>
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<td>Trajectory densities</td>
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<td>—</td>
</tr>
<tr>
<td>Intensity of diabatic heating along the traces</td>
<td>(Q_{dl})</td>
<td>W m(^{-2})</td>
</tr>
</tbody>
</table>

**Other abbreviations**

- NBKS: Northern Barents–Kara Seas
- WaMAIs: Warm and moist air intrusions
- MEP: Midlatitude eastern Pacific
- MNA: Midlatitude North Atlantic
- PNA: Pacific–North American pattern
- NAO: North Atlantic Oscillation

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FIG. 2. From left to right, composites of the (first column) anomalous geopotential height at 500 hPa, (second column) anomalous latent energy flux \([I_{i}, J_{i}]\), black vectors with \(E_{i}\) (shading), (third column) anomalous sensible energy flux \([I_{se}, J_{se}]\), black vectors with \(E_{s}\) (shading), (fourth column) anomalous \(F_{d}\), and (fifth column) anomalous SIC for lag −4, −2, 0, +2, +4, and +6 days, shown from top to bottom, respectively, during the SIC− events. Shading and vectors are plotted with statistical significance at the 95% confidence level, and the vectors are shown when energy flux anomalies are significant in both directions. The NBKS and BKS are bounded by blue and black solid polygons, respectively.
calculated. From the perspective of the regional mean, the increased DLW can theoretically account for at most 58% of the transient SIT decline during the strengthened DLW period in the NBKS. In terms of the spatial distribution, the negative SIT anomalies (Fig. 4a) are most prominent over the western Kara Sea and the northern Barents Sea. Comparably, the SIT anomalies caused by DLW changes (Fig. 4b) have a similar spatial distribution, although with a lower amplitude and a wider range, especially outside of the NBKS region. This phenomenon occurs because positive DLW anomalies over the Atlantic sector of the Arctic Ocean are largely offset by negative anomalies of the upward turbulent heat fluxes (Fig. S8). Furthermore, the increase in LE is observed to contribute to the SIT decrease by enhancing DLW over a large part of the northwestern NBKS (Fig. 4c). In comparison, the contribution of the increase in SE is located farther south, covering the central and southern Barents Sea (Fig. 4d).

c. Source detection for increased atmospheric energy

Given the finding that the added LE and SE supplies are originally responsible for sea ice loss in the Arctic, the sources of the anomalous LE and SE are further examined. Based on the atmospheric energy source detection method described in section 2, the horizontal distribution of the attributed energy source anomalies is shown in Fig. 5, and the meridional and zonal averages of the energy sources over regions north of 30°N are explicitly shown in Figs. 6a–d.

Figure 5a depicts the horizontal distribution of the composite atmospheric energy source anomalies (δSsc). Regions with δSsc exceeding 80 W m⁻² include the Barents Sea, Norwegian Sea, and the North Sea. Anomalies of the atmospheric energy source exceed 50 W m⁻² over the western coast of continental Europe, the Mediterranean, Black, and Caspian Seas, the midlatitude North Atlantic (MNA), and the midlatitude eastern Pacific (MEP). Additionally, inner continental Europe, the Greenland Sea, Baffin Bay, and eastern North America account for certain percentages of δSsc, with values larger than 5 W m⁻².

Figures 5b and 5c present the spatial characteristics of anomalous latent energy sources (δSle) and anomalous diabatic sensible energy sources (δSle). The term δSle reaches its maximum value in the southern Barents Sea and gradually decreases along the Norwegian, North, and Baltic Seas. In contrast, for δSse, two heating centers are located at the MNA and the MEP. The MNA has δSse exceeding 50 W m⁻², and the high values extend widely to Scandinavia and the Greenland Sea. In the MEP, δSse also surpasses 50 W m⁻², but the coverage of the positive δSse in this region is limited both meridionally and zonally compared with the MNA.

In contrast to latent energy sources, which are mainly due to evaporation, sensible energy sources are considerably more complicated because they consist of different components with different physical processes and various distribution features. Therefore, the anomalies of the partial contributions of δSsa associated with the latent heat release (δSsl), radiation (δSsr), and turbulent mixing (δSsm) are further extracted, as shown in Figs. 5d–f. In general, the spatial pattern of Fig. 5d resembles that of Fig. 5c, indicating that the anomalous diabatic sensible energy source for the SIC– events is largely dominated by the latent heat release, especially for the two
heating centers. In addition, as shown in Fig. 5e, enhanced turbulent heating contributes significantly to $\delta S_{\text{th}}$ of the MNA but slightly to that of the MEP. For radiative heating (Fig. 5f), a general negative contribution is observed.

After meridional averaging, as shown in Figs. 6a and 6c, the meridional average of latent energy sources monotonically decreases with increasing distance from the NBKS (30°–50°E). In comparison, the sensible energy source and its component due to latent heating are characterized by a multipeak pattern, with the strongest peak group near 50°W–10°E matching the MNA and the second most identifiable peak group near 150°–120°W matching the MEP. For the zonal mean, presented in Figs. 6b and 6d, the sensible energy sources are pronouncedly amplified in 40°–55°N but suppressed north of 70°N, while the latent energy sources are anomalously positive in 65°–75°N. Combined with Fig. 5, it can be considered that the NBKS itself contributes to the added latent energy supply but provides less sensible energy. Furthermore, as shown in Figs. 6c and 6d, the global mean $\delta S_{\text{th}}$ is slightly reduced by the anomalies of the other two diabatic heating components, and anomalies of $\delta S_{\text{th}}$ and $\delta S_{\text{rad}}$ are almost opposite at all longitudes, indicating negative feedback among different components of diabatic heating.

In addition, to test whether the source detection is sensitive to the definition of SIC—events, the percentile threshold is changed from the 10th to the 5th and 15th percentiles. The corresponding zonal and meridional averages are drawn in Fig. 6, and the spatial patterns of all attributed energy source anomalies, as shown in Fig. 5, are presented in Figs. S9 and S10. Our results suggest that the atmospheric source diagnosis varies slightly with the percentile threshold of the SIC—days.

4. Discussion

a. Feedback mechanisms during sea ice decline events

Figure 2a indicates slight lags between local variables involved in SIC—events in the NBKS, which implies a high degree of cross-coupling between the related processes. Considering the situation that has been described in section 3a in which DLW is the driver of sea ice loss and atmospheric energy is a driver of DLW, some feedback mechanisms are further discussed in this section. The time lag shown in Figs. 2 and 3 can be explained by the sea ice melting process (Thorndike 1992): when the DLW is enhanced, the surface temperature of the sea ice is modified to balance the downward energy surplus, accompanied by the suppression of ice formation; subsequently,
FIG. 5. Composites of (a) anomalous atmospheric energy sources ($\delta S_{ae}; \text{W m}^{-2}$), (b) anomalous latent energy sources ($\delta S_{le}; \text{W m}^{-2}$), (c) anomalous diabatic sensible energy sources ($\delta S_{se}; \text{W m}^{-2}$), (d) anomalous partial $S_{ae}$ contribution of latent heat release ($\delta S_{lh}; \text{W m}^{-2}$), (e) anomalous partial $S_{ae}$ contribution of turbulent heating ($\delta S_{tur}; \text{W m}^{-2}$), and (f) anomalous partial $S_{ae}$ contribution of radiative heating ($\delta S_{rad}; \text{W m}^{-2}$) [see section 2b(3) for details]. The NBKS is highlighted by the blue solid polygon.
Fig. 6. Meridional averages of anomalies over regions north of 30°N for (a) the composites of latent energy sources ($\Delta S_{\text{le}}$; W m$^{-2}$) and diabatic sensible energy sources ($\Delta S_{\text{se}}$; W m$^{-2}$); (c) the composites of partial $S_{\text{se}}$ contributions of latent heat release ($\Delta S_{\text{lh}}$; W m$^{-2}$), turbulent heating ($\Delta S_{\text{tur}}$; W m$^{-2}$), and radiative heating ($\Delta S_{\text{rad}}$; W m$^{-2}$); (e) the composites of trajectory densities 12 days prior to the SIC events ($\Delta TD$, the number of trajectories per square of 1° × 1°) and diabatic heating intensity ($\Delta Q_{\text{se}}$, the total diabatic heating or cooling that particles go through per square of 1° × 1° divided by the corresponding number of trajectories; W m$^{-2}$). (b),(d),(f) Zonal averages for the same variables of (a), (c), and (e), respectively. The solid, dashed, and dotted lines represent the results of SIC events selected by the percentile threshold from the 10th, 5th, and 15th percentiles, respectively. The blue shadings indicate two key latent energy sources (the MEP at 150°–120°W, 40°–55°N and the MNA at 50°W–10°E, 40°–55°N), and the yellow shadings indicate the key latent energy source (20°–60°E, 65°–75°N). Notably, both the positive and negative values in the subfigures are anomalies but not absolute values, thus not indicating actual heating or cooling. Negative values are not shown in (e) and (f) for better visualization.
when the ice is warmed to the freezing point, melting occurs and downward conductive heat is established in the ice slab; finally, until the cumulative change in the ice thickness reaches a certain value, the ice concentration is significantly impacted.

Moreover, the time lags are minor because of the feedback mechanisms of sea ice loss, in particular, the positive ice-albedo feedback (Francis et al. 2005; Burt et al. 2016) during winter when the surface albedo feedback mechanism (Lindsay and Zhang 2005; Screen and Simmonds 2010; Stroeve et al. 2012) is weakened. In the case of the SIC– events selected in this research, before lag day 0, the increase in atmospheric energy precedes DLW amplification and, thus, sea ice retreat. After lag day 0, the surface turbulent heat fluxes, which have been inhibited by horizontal atmospheric energy convergence, progressively increase. This phenomenon can be regarded as the manifestation of exposure of more ocean to the air since the insulating power of ice diminishes (Zhong et al. 2018). The strengthened turbulent heat humidifies and warms the near-surface air, which then becomes a stronger emitter of longwave radiation, resulting in further surface temperature gain and sea ice loss. The two processes before and after lag day 0 comprise the positive ice-albedo feedback, in which the changes in the ice thickness, DLW, and atmospheric energy transport are mutually reinforcing and finally contribute to the high degree of cross-coupling between the air and land fields related to SIC– events.

However, it is noteworthy that the ice-albedo feedback does not always exhibit the abovementioned process. Burt et al. (2016) highlighted a different situation: in a warming climate, local diabatic cooling weakens, leading to surface warming and sea ice loss, followed by increasing local evaporation, DLW, and further sea ice reduction. Burt et al. (2016) did not observe any enhanced poleward moisture transport but did observe enhanced evaporation due to the decrease in the SIT and increase in the surface wind speed. The difference can likely be attributed to the time scale and study area [Burt et al. (2016) focused on century variations in the whole Arctic].

Furthermore, previous studies show that the DLW enhancement driven by the atmospheric energy convergence can also influence the interannual SIC variation (D.-S. R. Park et al. 2015). However, Fig. 3a shows that the SIC anomaly recovers relatively quickly, which appears unable to affect the long-term changes. Based on Fig. 4 and the ice melting process of Thorndike (1992), we suspect that the SIT anomaly might have played a role in bridging sea ice variations in different time scales because it is more directly determined by the surface energy surplus. Therefore, further analysis is performed based on the nondetrended SIT data, as well as SIC data for comparison). In Figs. S1c and S1d, based on the Mann–Kendall test, significant decreasing interannual trends are observed after the year 2000 in the year-mean intensity of SIC– events (composites of the SIC anomalies on lag day 0 of the SIC– events), year-mean SIC anomalies, and year-mean SIT anomalies, which suggests that the year 2000 is a key point in time. Then, the composite variations in the nondetrended SIT and SIC anomalies during the SIC– events before and after the year 2000 are respectively investigated, and the results are shown in Fig. S11.

In Fig. S11, from 1979 to 1999, both nondetrended SIC and SIT anomalies recovered when the SIC– events ended. Nevertheless, after the year 2000, while SIC is always restored to its normal level after the lag day 20, SIT barely increases enough to regain its original pre-SIC– event value. This suggests that anomalous energy intrusions during the intraseasonal SIC– events transiently decrease the SIT in the NBKS from the year 2000 to 2018, which is not pronounced enough to affect the synoptic SIC simultaneously but ultimately manifests as the interannual SIC tendencies through the accumulation and strengthened positive ice-albedo feedback.

b. Analysis of sensible energy source formation

For climate change in the Arctic, previous studies have indicated the importance of the vigorous LE supply in the Atlantic Ocean, North Sea, and Norwegian Sea, mainly caused by a “NAO+ with the Ural blocking”–like large-scale circulation pattern (Luo et al. 2016a,b; Gong and Luo 2017; Luo et al. 2017; Zhong et al. 2018). In addition to the evidence along the lines of the studies above, Figs. 5 and 6a–d further show that the MNA (50°–70°E, 40°–55°N) and MEP (150°–120°W, 40°–55°N) are two major sources of anomalous SE transportation toward the Arctic through the Atlantic gate. To more comprehensively understand the new finding, these two SE sources are explored based on the trajectory density (δTD; number of trajectories per 1° × 1° square) and diabatic heating intensity along the traces (δQse; total diabatic heating or cooling that particles go through per 1° × 1° square divided by the corresponding number of trajectories).

Figures 6e and 6f show the anomalies (difference between the value during SIC– events and SIC= events) of the trajectory density and diabatic heating intensity. Zonally, the anomalous trajectory density and diabatic heating intensity in the MNA reach the largest and second largest peaks, respectively, implying that the MNA is most frequently visited by air particles before the SIC– events while also concurrently providing relatively more diabatic heating. Comparably, the MEP does not have obviously positive anomalies of trajectory density. However, the anomalies in diabatic heating intensity over the MEP are almost 3 times those in the MNA, indicating that although the MEP is not passed as frequently as the MNA, the extremely strong diabatic heating there makes it the second most important anomalous SE source. Meridionally, both δTD and δQse peak between zonal circles of 40° and 55°N (Fig. 6f), where the MNA and the MEP are located.

Furthermore, since the latent heat release dominates the MEP (Figs. 5c,d), the anomalies of tropospheric latent heating are presented in Fig. 7. As is described in section 2b, the daily series of tropospheric latent heating are deseasonalized, detrended, and 3-day moving averaged. By comparing Figs. 7a and 7c, it can be found that from 12 to 8 days before the SIC– events begin, a significant positive anomaly of latent heating occurs in the western tropical Pacific; from lag days −7 to −3, the latent heating anomaly becomes positive over the MEP while tropical heating still continues. Nevertheless, the MNA does not have a centralized and pronounced center of latent heating during either time window above, which is
consistent with the anticyclone overlying the region (Figs. 2 and 7d). Figures 6, 7a, and 7e suggest that before the SIC events are triggered, more diabatic heating caused by moisture condensation modulates the diabatic heating intensity over the MEP.

In summary, the amplified atmospheric energy advection into the NBKS is the manifestation of the large-scale circulation pattern (dynamic driver) and regional diabatic heating variation (thermodynamic driver). While the former pattern dominates the latent and sensible energy sources over the
MNA, the latter pattern largely shapes the sensible energy source regions over the MEP. In previous studies, the latent energy released before arriving at the Arctic was regarded as not affecting the target region. However, the results of this paper suggest that latent energy released before entering the NBKS is not simply dissipated but instead contributes to increased atmospheric energy in the Arctic (and thus strengthens DLIW and melts the sea ice there) by remote conversion into sensible energy and its subsequent transport. In this way, our study implies a new possible linkage between the tropics and extratropics.

c. Possible interactions between the tropics and the extratropics

Figures 7a and 7e suggest a lead-lag relationship between tropical heating and midlatitude heating, which has also been observed by Park and Lee (2019). They generalized this possible linkage as a “heating-circulation relay”: tropical latent heating can trigger the poleward propagating Rossby wave, which then transports warmer and moister air to the midlatitude North Pacific and leads to more latent heating in the region. Therefore, to assess the possibility that atmospheric energy source regions may be associated with tropical variation, composites of Rossby wave sources, streamfunction anomalies at 200 hPa, and geopotential height anomalies at 500 hPa are further examined.

The Rossby wave source is shown in Fig. 7b to describe the relationship between tropical diabatic heating and the general circulation at the upper-level troposphere. The Rossby wave source is defined as $-\nabla \cdot (\mathbf{v} \xi)$, where $\mathbf{v}$ is the divergent wind velocity and $\xi$ is absolute vorticity (Sardeshmukh and Hoskins 1988), meaning that the change in vorticity $[-\nabla \cdot (\mathbf{v} \xi)]$ or anomalous divergence $[-\nabla \cdot (\mathbf{v} \xi)]$ can influence the Rossby wave propagation. Tending to be balanced by the adiabatic cooling due to ascent, the tropical diabatic heating is usually accompanied by upper-level outflows; thus it has the potential to trigger the Rossby wave. From lag days $-12$ to $-8$, when anomalous latent heating occurs in the troposphere over the western tropical Pacific (Fig. 7a), a significant Rossby wave source at the upper level is observed in eastern Asia (Fig. 7b), which is located north and slightly east of the tropical heating center. On the one hand, lower-level moisture horizontally converges and upward convects in the western tropical Pacific, releasing latent heat in the troposphere and inducing upper-level divergences; on the other hand, the edge of this region is located at the subtropics, therefore featuring a larger vorticity gradient due to the jet stream. As a result, an anomalously positive Rossby wave source is triggered around the region $100^\circ$–$150^\circ$E, $30^\circ$–$40^\circ$N, indicating the response of Rossby wave propagation to the anomalous tropical heating.

Following the anomalies of the Rossby wave source, distinct teleconnection patterns can be observed. Figures 7c and 7d further present the streamfunction anomalies at the 250 hPa and geopotential height anomalies at 500 hPa from lag days $-7$ to $0$, respectively. In Fig. 7c, the Rossby wave train occurs in the North Hemisphere, and the transient waves overlap the stationary waves in eastern Asia, the tropical Pacific, the midlatitude Atlantic, and Canada. Stationary waves have been found to be the major contributor to poleward atmospheric energy transport in the Northern Hemisphere during wintertime, especially for sensible energy (Peixoto and Oort 1992). Therefore, we examined the stationary wave index (SWI; Goss et al. 2016) to quantify the consistency between the transient and stationary waves (SWI values larger than 1.0 represent constructive interference between transient waves and the climatological stationary waves). From lag days $-7$ to $0$, the SWI index varies from 0.45 to 0.86 (plot not shown), presenting that the transient waves to some extend interfere with the stationary waves and relatively amplify the poleward transport of the atmospheric energy. Along with the poleward propagation of the Rossby wave, potential height anomalies in the North Hemisphere begin to carry some features of the PNA and NAO+ patterns. In Fig. 7d, high pressure systems persist over the middle North Pacific Ocean north to the Bering Sea, in the vicinity of Hawaii, over the mountainous region of western North America, and over the western Atlantic Sea north to the Labrador Sea, while low pressure structures are located in the Aleutian Islands, the southeastern United States, Greenland, and Iceland. Finally, from lag days $-7$ to $-3$, following this Rossby wave train and the PNA-like pattern, the latent heat release over the MEP increases (Fig. 7e), which is consistent with the model result of Park and Lee (2019). The time window between lag days $-7$ and $-3$ is chosen instead of from lag days $-7$ to $0$ because trajectories pass the eastern midlatitude Pacific during the former time window.

Based on the preceding discussions, both the dynamic driver and thermodynamic driver introduced at the end of section 4b are likely to be originally triggered by tropical heating. Dynamically, as Stan et al. (2017) generalized, tropical convective activity can induce a Rossby wave source, which consists of the strong divergent flow oriented up by the gradient of vorticity and finally modulates the midlatitude large-scale circulation patterns. In this manner, the trajectory densities of the MNA are increased. Thermodynamically, the relatively constructive stationary-transient wave interference led by tropical heating favors more atmospheric energy transport from the tropics to the extratropical regions, which brings warmer and moister air into the MEP and triggers more latent heat release over there, thus shaping the high intensity of diabatic heating at the MEP.

5. Summary and conclusions

The sea ice reduction events in the Arctic, one of the most pronounced characteristics of global climate change, are caused by a suite of factors that are multiply intertwined. In this study, to provide a holistic view of the underlying mechanisms, first, the interaction between various physical fields associated with the SIC– events is analyzed based on the lagged composite method, a multiple linear regression model, and a toy ice model. Next, a Lagrangian method is proposed to detect the atmospheric energy sources concerning both latent and sensible energy. Finally, the mutually influenced dynamic and thermodynamic forcings are clarified. Generally, triggered by anomalous convection in the tropics, distinct
large-scale circulations and anomalous heating sources form in the midlatitude, which favor poleward atmospheric energy advection, result in the DLW rise, and finally increase the sea ice retreat in the NBKS. Specifically, our major conclusions are as follows:

1) In terms of quantifying the relationship among energy transport, DLW change, and sea ice variability, theoretically, at most 58% of the wintertime sea ice decline in the NBKS during the SIC–events selected in this research is driven by DLW variability, which is further attributed to the enhanced poleward advection of atmospheric LE (56%) and SE (28%). Compared with the increase in LE, which mainly induces sea ice reduction over a large scale of the northwestern NBKS, the contribution of the increase in SE is in more southern regions.

2) For the composited atmospheric energy sources, the midlatitude North Atlantic, the Norwegian, North, and Baltic Seas, western Europe, the northeastern Pacific, and western North America are the major anomalous energy source regions for the sea ice loss over the wintertime NBKS. While the anomalies of latent energy sources peak locally and decrease with increasing distance from the NBKS, the anomalies of sensible energy sources are negative locally and have two main heating centers over the MNA and MEP. In addition, latent heating is the main physical component of anomalous sensible energy sources.

3) Mechanistically, our research to some extent complements the mechanism of the poleward transport of atmospheric energy. After enhanced convection occurs in the western tropical Pacific, significant constructive stationary-transient interference follows, contributing to the distinct large-scale circulation (dynamic driver) and strengthening the conversion between LE and SE in the northeastern Pacific (thermodynamic driver). While the dynamic driver forms more frequent WaMAIs to the Arctic, the thermodynamic driver increases the sensible energy carried by the WaMAIs. As a result, the atmospheric energy of the Arctic is added, which then increases the DLW and reduces the Arctic sea ice. In return, thinner ice enables more energy to be advected from the ocean to the atmosphere, thus further warming the surface and melting the ice.

When examining and interpreting the possibility of tropical–extratropical interactions that are related to sea ice retreat in the Arctic, we have relied on lead–lag reanalysis, Rossby wave source theory, and transient-stationary wave construction theory. However, whether the enhanced latent heat release over the MEP is triggered by the latent heating over the western tropical Pacific through Rossby waves has not been directly proven in this work. To directly prove this heat–circulation–heat relay, model-based study is necessary. Further, comparison of the sea ice variation in the Arctic can be made for situations 1) without any anomalies, 2) with an increase in latent heat release only over the western tropical Pacific, 3) with an increase in latent heat release only over the MEP, and 4) with an increase in latent heat release only over the MNA, to quantify the contribution of different energy sources to sea ice retreat in the Arctic. Additionally, only diabatic heating is taken into consideration in our Lagrangian energy detection because we focus on the mutual influence between hydrological circulation and energy transportation. In the future, the adiabatic heating can be included based on the difference between absolute temperature and potential temperature along trajectories, which corresponds to our planned objective in the next stage.

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Data availability statement. This study utilizes data from the European Centre for Medium-Range Weather Forecasts (ECMWF) interim reanalysis (ERAI; Dee et al. 2011), Coupled Pan-Arctic Ice-Ocean Modeling and Assimilation System (PIOMAS; Zhang and Rothrock 2003), Modern-Era Retrospective Analysis for Research and Applications (MERRA; Rienecker et al. 2011), and Modern-Era Retrospective Analysis for Research and Applications, version 2 (MERRA-2; Gelaro et al. 2017).

The 2D and 3D ERAI reanalysis data are openly available at https://apps.ecmwf.int/datasets/data/interim-full-daily/levtype=sfc/ and https://apps.ecmwf.int/datasets/data/interim-full-daily/levtype=pl/, respectively. The flux fields are downloaded at step 12 h at 0000 and 1200 UTC, including the DLW, surface net solar radiation, surface turbulent heat flux, latent heat flux, and sensible heat flux. Other variables are downloaded at step 0 h at 0000, 0600, 1200, and 1800 UTC. The daily field of SIT is openly available at https://pscfiles.apl.washington.edu/zhang/PIOMAS/data/v2.1/hiday/. The diabatic heating datasets from MERRA and MERRA-2 are openly available at https://goldsmr3.gesdisc.eosdis.nasa.gov/data/MERRA/MAT3CPTDT.5.2.0/ and https://goldsmr5.gesdisc.eosdis.nasa.gov/data/MERRA2/M2T3NPTDT.5.12.4/, respectively. The variables named temperature tendency from moist physics, temperature tendency from radiation, and temperature tendency from turbulence are selected.

APPENDIX A

Calculation of the Atmospheric Variables

The integrated atmospheric energy and its fluxes are obtained using the following equations:

\[
E_i = -\frac{1}{g} \int_{\rho_i}^{\rho_0} L_\omega q \, dp, \quad (A1)
\]

\[
lpt(I_{\rho_0}, I_\omega) = \left(-\frac{1}{g} \int_{\rho_i}^{\rho_0} L_\omega q \, dp \right) - \left(-\frac{1}{g} \int_{\rho_i}^{\rho_0} L_\omega q \, dp \right). \quad (A2)
\]
\[
FC_{le} = \frac{1}{g} \int_{p_l}^{p_0} \nabla \cdot (L_c q V) dp, \tag{A3}
\]
\[
E_i = - \frac{1}{g} \int_{p_l}^{p_0} C_p T dp, \tag{A4}
\]
\[
(I_{se}, J_{se}) = \begin{pmatrix} - \frac{1}{g} \int_{p_l}^{p_0} C_p T dp \n - \frac{1}{g} \int_{p_l}^{p_0} C_p T dp \end{pmatrix}, \tag{A5}
\]
\[
FC_{se} = \frac{1}{g} \int_{p_l}^{p_0} \nabla \cdot (C_p TV) dp. \tag{A6}
\]

where \( E_l \) is the vertical integral of latent energy; \((I_{se}, J_{se})\) are the eastward and northward latent energy fluxes, respectively; \( FC_{se} \) is the latent energy flux convergence; \( E_i \) is the vertical integral of sensible energy; \((I_{se}, J_{se})\) are the eastward and northward sensible energy fluxes, respectively; \( FC_{se} \) is the sensible energy flux convergence; \( L_v = 2.26 \times 10^6 \) J kg\(^{-1}\) is the latent heat of vaporization; \( C_p = 1005 \) J kg\(^{-1}\) K\(^{-1}\) is the specific heat of dry air under constant pressure; \( g = 9.8 \) m s\(^{-2}\) is the acceleration of gravity; \( p_l = 200 \) hPa is considered as the mean pressure of the top of the troposphere; \( q \) is the surface pressure (hPa); \( q \) is the specific humidity (kg kg\(^{-1}\)); \( T \) is the absolute temperature (K); \( p \) is the surface temperature; \( u, v \) and \( w \) represent the horizontal wind velocity and its zonal and meridional components (m s\(^{-1}\)), respectively; and \( V \) is the Laplace operator.

The streamfunction \( (\psi) \) satisfies the following equation:

\[
\begin{align*}
\frac{\partial \psi}{\partial y} &= \frac{\partial u}{\partial x} \\
\frac{\partial \psi}{\partial x} &= - \frac{\partial u}{\partial x},
\end{align*}
\tag{A7}
\]

where \( x \) and \( y \) denote coordinates in the Cartesian coordinate system; \( \psi \) is calculated using the Python package “windspharm” (Dawson 2016).

**APPENDIX B**

**Derivation and Application of the Multiple Linear Regression Model of \( F_{dl} \)**

The empirical DLW evaluation equation can be expressed as

\[
F_{dl} = e \sigma T^4, \tag{B1}
\]

where \( F_{dl} \) is surface downward longwave radiation; \( T \) is the absolute temperature; \( \sigma \) is the Stefan–Boltzmann constant; and \( e = e(\mathrm{cl}, e, T) \) is the effective longwave emissivity of the atmosphere, which is empirically a function of the cloudiness \( (\mathrm{cl}) \); water vapor pressure at 2 m \( (e) \), and temperature \( (T) \).

By applying the total differential operators on both sides of Eq. (B1), the variations in \( F_{dl} \) can be expressed as a function of \( E_l \) (or specific humidity) and \( E_i \) (or temperature) changes, yielding the following:

\[
\delta F_{dl} = \frac{\partial F_{dl}}{\partial E_l} \delta E_l + \frac{\partial F_{dl}}{\partial E_i} \delta E_i + \text{RES}. \tag{B2}
\]

Equation (B2) represents the multiple linear regression model of \( F_{dl} \), where \( \delta \) represents an anomaly; \( \delta F_{dl} \) denotes the predicted value; \( \delta E_l \) and \( \delta E_i \) denote two predictors; \( \frac{\partial F_{dl}}{\partial E_l} / \delta E_l \) and \( \frac{\partial F_{dl}}{\partial E_i} / \delta E_i \) denote partial regression coefficients of \( \delta E_l \) and \( \delta E_i \), respectively; and \( \text{RES} \) denotes the intercept term, which can be attributed to cloudiness, surface temperature, and other factors.

By dividing both sides by \( \delta F_{dl} \), the following expression can be obtained:

\[
1 = \frac{\partial F_{dl}}{\partial E_l} \frac{\delta E_i}{\delta F_{dl}} + \frac{\partial F_{dl}}{\partial E_i} \frac{\delta E_i}{\delta F_{dl}} + R. \tag{B3}
\]

Based on the condition that 1) during the strengthened DLW period (\( \delta F_{dl} > 0 \)) and 2) being within the NBKS, 897 samples are selected in total. By substituting \( \Delta F_{dl} / \delta F_{dl} \) and \( \Delta E_i / \delta F_{dl} \) into Eq. (B3), Eq. (1) is obtained, where \( \Delta E_i / \delta F_{dl} \) and \( \Delta E_i / \delta F_{dl} \) indicate \( [\sum_{n=0}^{897}(\Delta E_{in}/\delta F_{dl})]/897 \) and \( [\sum_{n=0}^{897}(\Delta E_{in}/\delta F_{dl})]/897 \), respectively.

**APPENDIX C**

**Generation of Lagrangian Trajectories**

The HYSPLIT model is used to generate backward air trajectories and is driven by the multilevel (from 1000 to 50 hPa in vertical intervals of 50 hPa) variables (zonal, meridional, and vertical wind, relative humidity, temperature, and geopotential height) and the surface variables (surface air temperature, zonal and meridional 10-m wind) from the ERAI. These data are provided to the HYSPLIT model at a temporal resolution of 6 h and a spatial resolution of 1.0° longitude × 1.0° latitude.

The starting points are homogeneously distributed on 3D grids, which can be projected to a horizontal surface with 338 quasi-equidistant points in the target region, and are vertically distributed among 17 levels from 1000 to 200 hPa in intervals of 50 hPa (including only the pressure layers above the surface). Therefore, these points can represent tropospheric air particles with the same fraction of the entire air mass. Next, trajectories are released from these starting points at 00:00 on the onset day of the SIC− (65 events) and SIC+ events (154 events). Under this situation, 365 518 (865 705) Lagrangian trajectories should be released for SIC− events (SIC = events), determined as the number of horizontal points × the number of vertical levels × the number of events = 338 × 17 × 65 (154). When the surface pressure is less than 1000 hPa, the number of vertical levels is less than 17.

The output parameters of the HYSPLIT model include the latitude, longitude, pressure, height, potential temperature, specific humidity, and mixing depth every 3 h. Moreover, based on MERRA-2 data, the diabatic heating values \( (Q, Q_{LH}, Q_{RAD}, \text{and } Q_{TUR}) \) along the tracks are obtained by interpolating the MERRA-2 grid data onto locations with a 3-h interval on traces. Since the MERRA-2 dataset begins in 1980, interpolation for events in 1979 is performed based on the corresponding data from MERRA.


