The Impact of Poleward Moisture and Sensible Heat Flux on Arctic Winter Sea Ice Variability*

HYO-SEOK PARK

Korea Institute of Geoscience and Mineral Resources, Daejeon, South Korea

SUKYOUNG LEE

School of Earth and Environmental Sciences, Seoul National University, Seoul, South Korea, and Department of Meteorology, Pennsylvania State University, University Park, Pennsylvania

SEOK-WOO SON

School of Earth and Environmental Sciences, Seoul National University, Seoul, South Korea

STEVEN B. FELDSTEIN

Department of Meteorology, Pennsylvania State University, University Park, Pennsylvania

YU KOSAKA

Research Center for Advanced Science and Technology, The University of Tokyo, Tokyo, Japan

(Manuscript received 22 January 2015, in final form 16 April 2015)

ABSTRACT

The surface warming in recent decades has been most rapid in the Arctic, especially during the winter. Here, by utilizing global reanalysis and satellite datasets, it is shown that the northward flux of moisture into the Arctic during the winter strengthens the downward infrared radiation (IR) by $30\text{--}40\text{ W m}^{-2}$ over 1–2 weeks. This is followed by a decline of up to 10% in sea ice concentration over the Greenland, Barents, and Kara Seas. A climate model simulation indicates that the wind-induced sea ice drift leads the decline of sea ice thickness during the early stage of the strong downward IR events, but that within one week the cumulative downward IR effect appears to be dominant. Further analysis indicates that strong downward IR events are preceded several days earlier by enhanced convection over the tropical Indian and western Pacific Oceans. This finding suggests that sea ice predictions can benefit from an improved understanding of tropical convection and ensuing planetary wave dynamics.

1. Introduction

As the solar insolation in high latitudes rapidly weakens in late fall, sea ice over the Arctic Ocean

gradually thickens and extends farther southward until it reaches its maximum extent in early March. The processes that drive sea ice variability during the Arctic winter have not received as much attention as those for the summer, when ice–albedo feedback is thought to play an important role. However, Arctic warming has been most rapid during the winter (Bekryaev et al. 2010). Since downward infrared radiation (IR) is an important warming agent in the Arctic (Walsh and Chapman 1998; Uttal et al. 2002; Winton 2006), it is natural to ask whether fluctuations in downward IR can induce changes in the sea ice. Wintertime sea ice fluctuations have generally been attributed to wind-induced

* Supplemental information related to this paper is available at the Journals Online website: <http://dx.doi.org/10.1175/JCLI-D-15-0074.s1>.

Corresponding author address: Suhyoung Lee, Department of Meteorology, The Pennsylvania State University, University Park, PA 16802.
E-mail: slx31@psu.edu

ice motion (Fang and Wallace 1994; Rigor et al. 2002; Rigor and Wallace 2004; Kwok et al. 2005; Sorteberg and Kvingedal 2006; Liptak and Strong 2014) and temperature anomalies (Deser et al. 2000), but increased downward IR in Arctic winter was also proposed as a factor for changing sea ice thickness (Francis and Hunter 2006). In particular, recent studies have found that poleward moisture fluxes into the Arctic associated with anomalous wintertime large-scale circulations increase downward IR at the surface (Lee et al. 2011; Yoo et al. 2012a; Woods et al. 2013). While one may wonder if an increase in downward IR can reduce sea ice thickness during the Arctic winter, we will show with model calculations that the relatively thin sea ice over the Barents, Kara, and Laptev Seas (see Fig. S1 in the online supplemental material) readily decreases for typical positive downward IR anomalies.

In this study, we investigate the increase in wintertime downward IR and the associated sea ice melting (or suppression of sea ice growth). As we will show, Arctic downward IR fluctuates on a weekly time scale. We use the European Centre for Medium-Range Weather Forecasts (ECMWF) interim reanalysis (ERA-Interim; Dee et al. 2011) for the horizontal moisture flux, downward IR, and surface heat fluxes. Arctic station-based observations indicate that reanalysis data capture the observed day-to-day variability of downward IR reasonably well (Morcrette 2002). To analyze the impact of winter downward IR, with minimal impact by solar radiation, data from December to early March are analyzed. The relationships among the moisture flux, downward IR, and sea ice are analyzed using lagged composite analyses where individual events are chosen based on the occurrence of anomalously strong downward IR anomalies averaged over the domain from 70° to 90°N (see section 2). To corroborate the physical relationship between downward IR and sea ice identified in our analysis, we also examine the output from a coupled climate model.

2. Data and methods

a. Data

The ERA-Interim dataset is 6-hourly on 16 pressure levels between 1000 and 30 hPa, with a horizontal resolution of $1.5^\circ \times 1.5^\circ$. To cross-check our findings from the ERA-Interim, we also analyzed data from the Japanese 25-year Reanalysis Project (JRA-25; Onogi et al. 2007) produced by the Japan Meteorological Agency and the Central Research Institute of Electric Power Industry, Japan. The JRA-25 global model has a spectral resolution of T106 and 40 vertical layers with the top at 0.4 hPa. Arctic sea ice concentration data are from the U.S. National Snow and Ice Data Center (NSIDC) where the NASA Team algorithm

(Swift and Cavalieri 1985) was used. This algorithm estimates sea ice concentration from satellite-derived passive microwave brightness temperatures. These data are provided on a polar stereographic grid with $25 \text{ km} \times 25 \text{ km}$ resolution. We regridded these data onto a regular $1.0^\circ \times 1.0^\circ$ grid. The NSIDC sea ice concentration data are available every two days from the year 1979 to 1989, and then daily since 1990. Because we are focusing on daily variations of sea ice, we used the data that begin in 1990. For Arctic sea ice thickness, the coupled Pan-Arctic Ice Ocean Modeling and Assimilation System (PIOMAS; Zhang and Rothrock 2003) is used. PIOMAS consists of a 12-category thickness and enthalpy distribution sea ice model coupled with the Parallel Ocean Program (POP) ocean model (Smith et al. 2010).

To identify the signal of tropical convection, we use daily-mean outgoing longwave radiation (OLR) data from the National Oceanic and Atmospheric Administration (NOAA); satellite-based data are interpolated onto a $2.5^\circ \times 2.5^\circ$ grid (Liebmann and Smith 1996). Both for the reanalysis and NSIDC sea ice concentration data, we used boreal winter (from 1 December to 10 March) data for the 23-yr period spanning December 1990–March 2012. Unlike the conventional definition of winter (December–February), our definition of Arctic winter extends to 10 March. This is mainly because Arctic sea ice cover reaches its maximum extent in early March, after which the intensifying solar radiation starts to rapidly melt the sea ice edge.

To test the ideas gained from the observational analysis, we also examined output from the coupled climate model GFDL CM3.0 (Donner et al. 2011) that was developed at the NOAA/Geophysical Fluid Dynamics Laboratory (GFDL) and has participated in phase 5 of the Coupled Model Intercomparison Project (CMIP5). This model uses a finite-volume dynamical core with a horizontal resolution of approximately 200 km. The sea ice component of CM3.0 is the GFDL Sea Ice Simulator (SIS), which is the same sea ice model used in CM2.1. The SIS is a dynamical model with three vertical layers, one for snow and two for sea ice, and five ice thickness categories. In this study, we analyze a transient twentieth- to twenty-first-century simulation forced by historical and representative concentration pathway 4.5 (RCP4.5) forcing. Here, we examine the first ensemble member (r1). For the model output, we take advantage of the availability of the data over a longer time period: from the year 1990 to 2030.

b. Methods

1) LAGGED COMPOSITE ANALYSIS

Composite calculations are performed in order to investigate the impact of downward IR on the intra-seasonal variability of sea ice concentration during the

Arctic winter. First, the seasonal cycle is removed at each grid point; the seasonal cycle is defined as a 30-day moving average of the 23-yr mean for each calendar day. The deseasonalized downward IR is zonally and meridionally averaged from 70° to 90°N (cosine weighting is applied to each latitude) to obtain an Arctic-mean downward IR. This daily downward IR time series is then detrended for each calendar day (from day 1 to day 365). A downward IR event is defined as a time period when the downward IR value exceeds 1.0 standard deviation (σ) for three or more consecutive days. If the beginning of an event occurs within 7 days of the end of the preceding event, then the latter event is discarded. This procedure identifies 36 events during 1 December–10 March over the 23 years (1990–2012). Lag zero is defined as the onset day (i.e., the first day of the event). These lag-zero days are used to construct composites of all variables considered in this study. Prior to generating the composites, a 3-day moving average is applied to filter out noise associated with day-to-day fluctuations. The same procedure is also applied to the GFDL CM3.0 output to identify the effect of downward IR events on sea ice thickness changes.

2) A TOY MODEL

Following simple modeling studies (Eisenman 2012; Thorndike 1992), the changes in sea ice thickness can be written as a function of the anomalous downward IR:

$$\delta h = -\epsilon \delta Q / (\rho L_f), \quad (1)$$

where δh denotes sea ice thickness change per unit time, δQ the anomalous downward IR (W m^{-2}), ρ the density of sea ice, and L_f the latent heat of fusion for sea ice. The quantity ϵ is a measure of efficiency: $\epsilon = 1 - [\gamma/h]^{-1}$, where γ is the thermodynamic scale thickness (0.7 m; Eisenman 2012) and h the initial sea ice thickness. Therefore, the efficiency ϵ is larger the thinner the sea ice. In other words, relatively thin sea ice is more sensitive to the changes in downward IR. This is because the diffusive heat flux through the ice is inversely proportional to sea ice thickness (Thorndike 1992). The value of γ controls the degree of the sensitivity. For h , we used the climatological mean sea ice thickness provided by PIOMAS.

To estimate the cumulative changes in sea ice thickness associated with the strong downward IR events, the anomalous daily δh is integrated from lag day -8 to day $+8$:

$$\Delta h \equiv \sum_{i=-8}^{+8} \delta h_i = -\frac{\epsilon}{\rho L_f} \sum_{i=-8}^{+8} \delta Q_i \delta t_i, \quad (2)$$

where Δh is the cumulative change (in h) from lag day -8 to day $+8$, and the subscript i is the index for the lag days. Thus, δt_i is the time step corresponding to each lag day. Here, δt_i is one day for all lag days. A 16-day time integration is used because downward IR events, on average, start about 4–8 days before they peak and end about 8 days later.

3. Results

Figure 1 (middle) shows that the Arctic downward IR strengthens rapidly over several days, and this is preceded by a poleward moisture flux (vectors in the left column) from both the Norwegian and Bering Seas, with the moisture flux in the former region being more pronounced. This relative timing suggests that anomalously strong downward IR may be caused by poleward moisture transport. The vertically integrated moisture flux convergence, multiplied by the latent heat of vaporization L_v ($=2.26 \times 10^6 \text{ J kg}^{-1}$), compares reasonably well with the downward IR in spatial structure if the former is compared with the latter with a 2-day lead time. Both the moisture flux convergence (multiplied by L_v) and the downward IR range up to 40 W m^{-2} . Because Arctic clouds have a relatively low amount of liquid water in winter, an increase in cloud liquid water can strengthen the downward IR (Gorodetskaya and Tremblay 2008; Chen et al. 2006). Therefore, poleward moisture fluxes can have a large impact on downward IR. Moreover, poleward moisture fluxes are accompanied by a poleward sensible heat flux (e.g., warm air advection from the south), further strengthening the downward IR (Skific and Francis 2013). Figure 2 shows that the relationship between the poleward moisture flux and downward IR is not just limited to the composite events, but also holds within the entire 1990–2012 wintertime period at a correlation of 0.53. The lag correlations demonstrate that the moisture flux leads the downward IR by 1–3 days.

Downward IR events are followed by anomalously low sea ice concentration (SIC), especially over the Greenland, Barents, and Kara Seas [Fig. 1 (right)]. Over a 6-day period, the SIC in these regions decreases by 10%, an amount comparable to or slightly smaller than one standard deviation of the corresponding daily SIC time series [Fig. S2 (left panel) in the online supplemental material]. This daily variation of SIC is only about 50%–60% of the interannual variability of SIC, which is defined as one standard deviation of the December through February (DJF)-mean SIC [Fig. S2 (right panel) in the online supplemental material]. However, if strong downward IR events occur multiple times in a single winter season, because of their cumulative effect the seasonal change would be larger than that caused by an individual event.

FIG. 1. Composites of the anomalous (left) moisture flux, (middle) downward IR, and (right) sea ice concentration for (from top to bottom) lag -4, -2, 0, and +4 days. In the left column, vectors show the vertically integrated moisture flux (kg m s^{-1}) and the shading is the estimated latent heat release (W m^{-2}). For the moisture flux convergence and downward IR, statistically significant (Student's t test; $p < 0.05$) values are shaded. The black contours in the right column indicate the regions where the sea ice reduction is statistically significant (Student's t test; $p < 0.05$).

ERA-interim

JRA-25

FIG. 2. (a) Scatterplot between the zonal-mean poleward moisture flux at the Arctic Circle (abscissa; averaged between 70° and $75^{\circ}N$) and the Arctic-mean downward IR (ordinate; averaged between 70° and $90^{\circ}N$), with the former leading the latter by 2 days. (b) Lag correlations between the poleward moisture flux and downward IR. Negative lags indicate moisture flux leading downward IR. Both for (a) and (b), the winter days (1 Dec–10 Mar) from 1990 to 2012 are used, and the deseasonalized daily poleward moisture flux and downward IR are smoothed with a 3-day moving average. (c),(d) As in (a),(b), but for JRA-25.

Therefore, it is possible that downward IR events are an important contributor for interannual SIC variability.

We further explore the relationship between downward IR and SIC by calculating the sea ice thickness change caused by the observed downward IR anomaly. The changes in sea ice thickness associated with the anomalous downward IR [δQ in (1) and (2)], integrated from day -8 to day $+8$, with the initial sea ice thickness corresponding to its climatological value, is presented in Fig. 3a. The spatial pattern of the change in sea ice thickness is generally consistent with the downward IR anomalies (Fig. 1), with the sea ice thickness change being particularly large over the regions where the mean

sea ice thickness is relatively thin, such as the Barents–Kara Seas. For an initial ice thickness of less than 4–5 cm, this downward IR can completely melt the sea ice. Therefore, the downward IR anomaly can help to explain the SIC decline during a single downward IR event (right column of Fig. 1). The sensitivity of the sea ice thickness change Δh to the initial sea ice thickness h for a given downward IR forcing ΔQ ,¹ based on (1), is

¹ For Fig. 3b, ΔQ is given as $24\ MJ\ m^{-2}$, a value close to the anomalous downward IR over the Barents–Kara Seas integrated from the lag day -8 to day $+8$.

(a) thickness changes by downward IR

(b) sensitivity of thickness changes to mean sea-ice thickness

FIG. 3. (a) Changes in sea ice thickness (cm) associated with winter downward IR events, calculated with (1). (b) The sensitivity of sea ice thickness changes to the initial sea ice thickness for the cumulative downward IR forcing of 24 MJ m^{-2} .

presented in Fig. 3b. During the last decade, Arctic sea ice has been thinning (Hansen et al. 2013; Renner et al. 2014), leading to an increased thermal efficiency ϵ in (1). Therefore, the anomalous downward IR forcing (ΔQ) has become more effective in reducing sea ice thickness.

A climate model simulation (see section 2a) supports the above observation-based estimation. Figure 4 presents model analysis that was constructed following the same procedure used to produce Fig. 1. Comparing these two figures, it can be seen that the simulated downward IR events are in good agreement with those from the reanalysis in spatial pattern and magnitude; the downward IR anomaly maximum is close to 40 W m^{-2} , and occurs over the Barents–Kara Seas [Fig. 4 (left)]. At the end of these strong downward IR events, specifically at day +8, the simulated sea ice thins by 5–6 cm over a wide area of the Arctic Ocean, including the Barents–Kara, Greenland, and Chukchi Seas [Fig. 4 (middle)]. The magnitude of these model-simulated sea ice thickness changes is comparable to the observation-based sea ice thickness changes calculated from (1). Over the marginal sea ice zone, especially over the Barents Sea [Fig. 4 (right)], the simulated SIC pattern during the downward IR events resembles the sea ice thickness anomaly pattern, suggesting that sea ice thinning driven by downward IR anomalies contributes to the SIC decline. Toward the interior of the Arctic Ocean where sea ice is relatively thick, the thinning would have no consequence on SIC, and this is borne out in lag day +8 anomalies.

Despite the positive downward IR over the entire Arctic, there are patches of slightly positive sea ice thickness anomalies (e.g., the Laptev Sea), suggesting

that other processes, such as wind-induced sea ice convergence, are at work. This is to be expected because positive Arctic downward IR anomalies are concurrent with poleward circulation anomalies (Lee et al. 2011; Yoo et al. 2012b; Woods et al. 2013). The effect of wind-driven ice drift on the redistribution of sea ice thickness is evaluated by utilizing the daily sea ice velocity provided by the GFDL CM3. Specifically, the time rate of change in thickness caused by ice drift can be estimated as

$$\frac{\partial h}{\partial t} = - \left[\frac{\partial}{\partial x}(uh) + \frac{\partial}{\partial y}(vh) \right], \quad (3)$$

where u and v denote zonal and meridional ice-drift velocities respectively. This sea ice volume flux convergence is calculated for every time step and a deviation from the long-term climatological mean is calculated. The resulting anomalies are integrated forward in time from lag day -8 to lag day j to estimate the cumulative change in sea ice thickness at different stages during the downward IR events:

$$\Delta h(j) \equiv \sum_{i=-8}^j \delta h_i = - \sum_{i=-8}^j \left[\frac{\partial}{\partial x}(uh) + \frac{\partial}{\partial y}(vh) \right]'_i \delta t_i. \quad (4)$$

Here, a prime (') denotes a deviation from the long-term climatological mean.

Figure 5 shows the total thickness change (Fig. 5a) and the ice drift–induced thickness change (Fig. 5b). As can be seen, the total and ice drift–induced cumulative thickness changes remain similar until lag day 0 (first

FIG. 4. GFDL CM3.0: Composites of the anomalous (left) downward IR (W m^{-2}), (middle) sea ice thickness (cm), and (right) sea ice concentration (%) for (from top to bottom) lag -2, 0, +2, and +8 days. The black contours in the middle and right columns indicate the regions where the sea ice changes are statistically significant (Student's t test; $p < 0.05$).

FIG. 5. GFDL CM3.0: Composites of the (a) anomalous sea ice thickness (cm) and (b) thickness changes by ice drift (cm) for (left) lag 0, (middle) lag +8, and (right) lag +16 days. Thickness changes (cm) by sea ice drift as a function of time, averaged over (c) the Atlantic sector (0° – 60° E) and (d) the entire Arctic (0° – 360°). The left and middle columns of (a) are identical to the second and fourth rows of Fig. 4.

column). The time evolution of anomalous sea ice thickness averaged over the Atlantic sector (cf. the red and black curves in Fig. 5c) indicates that the rapid decrease in sea ice thickness during the early stage of the downward IR events (until day +2) is dictated by ice drift. The ice drift-induced thickness change then rapidly diminishes at about day +3 as the southerly winds weaken. However, sea ice thickness continues to decrease. Based on the results presented in Fig. 3a, it is likely that beyond day +3 the majority of the subsequent sea ice decline is driven by the increased downward IR. Over most of the Arctic Ocean, the total

sea ice thickness steadily declines through to day +16. This contrasts with the ice drift-induced thickness change, which exhibits a dipole anomaly pattern. These differences are clearly seen when the total and ice drift-induced thickness changes are averaged over the Arctic Ocean (Fig. 5d). While sea ice thickness continuously decreases until day +16 (black curve), by this time the ice drift effect on sea ice thickness is small (red curve). These findings indicate that the ice drift is effective in decreasing sea ice thickness regionally, such as over the Atlantic sector. However, for the entire Arctic Ocean, during the downward IR events,

FIG. 6. (left) Composites of the anomalous outgoing longwave radiation (W m^{-2}) for (from top to bottom) lag -14, -10, -6, and -2 days. (right) Anomalous 250-hPa streamfunction ($10^6 \text{ m}^2 \text{ s}^{-1}$) for (from top to bottom) lag -11, -7, -3, and +1 days. Statistically significant (Student's t test; $p < 0.05$) values are hatched.

the effect of thickness changes due to ice drift appears to be more limited.

The above findings suggest that an increase in the poleward moisture flux during the winter, realized through an enhanced moisture flux convergence and downward IR on intraseasonal time scales, leads to a substantial reduction in Arctic sea ice thickness. Recent studies consistently indicate that Arctic warming is preceded by poleward propagating planetary-scale Rossby waves that are often triggered by tropical convection (Lee et al. 2011; Yoo et al. 2012b; Ding et al. 2014). Indeed, as Fig. 6 shows, the downward IR events are preceded several days earlier by enhanced convection [as indicated by negative anomalies in OLR] over the tropical Indian–western Pacific Ocean and a few days later by a Rossby wave train over the

North Pacific and North America. This result is consistent with a modeling study (Yoo et al. 2012b) that shows a similar arching wave train in the transient solution of a global circulation model, when initialized with warm pool convective heating superimposed on a climatological DJF background flow. By lag day -3, a cyclonic circulation is firmly established over the Canadian Arctic Archipelago and Greenland. With the accompanying anticyclonic circulation to its east, the circulation can bring warm, moist air poleward toward the Barents and Kara Seas, as shown in Fig. 1. A recent observational study (Henderson et al. 2014) also found a linkage between tropical convection (associated with the MJO) and Arctic sea ice, which they attributed to wind-driven sea ice motion.

4. Discussion

As the thickness of the Arctic sea ice declines over the long term, (1) indicates that the downward IR forcing can be effective at thinning the sea ice. In future climate change, Arctic sea ice is projected to become even thinner (Holland et al. 2006). Therefore, the winter weather perturbations and the associated downward IR are likely to be even more influential in changing sea ice thickness over wider areas of the Arctic Ocean. Also downward IR events are accompanied by wind-driven sea ice motion and, as the thinning continues, this effect on sea ice variability is also expected to become more potent. In addition, since long-term changes in the frequency of intraseasonal time scale processes have been shown to account for interdecadal variability in the atmospheric circulation (Johnson et al. 2008; Lee and Feldstein 2013), it is possible that an enhanced understanding of processes such as those investigated in this study can help to improve our understanding of the interdecadal changes in sea ice.

Acknowledgments. We thank the editor and anonymous reviewers for their constructive comments. HSP is supported by the Basic Research Project of the Korea Institute of Geoscience and Mineral Resources (KIGAM) funded by the Ministry of Knowledge Economy of Korea. SWS's work is supported by the Basic Science Research Program through the National Research Foundation of Korea (NRF) funded by the Ministry of Education (2013R1A1A1006530). SL is supported by Seoul National University, South Korea. SBF is supported by NSF Grants AGS-1036858 and AGS-1401220, USA. YK is supported by the Japanese Ministry of Environment through the Environment Research and Technology Development Fund A-1201 and by the Japan Society for the Promotion of Science Grant-in-Aid for Scientific Research 25287120.

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