https://doi.org/10.5194/egusphere-2022-552
Preprint. Discussion started: 11 October 2022

Modelled variations of the inherent optical properties of summer Arctic ice and their effects on the radiation budget: A case based on ice cores from CHINARE 2008–2016

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Abstract. Variations in Arctic sea ice are not only apparent in its extent and thickness but also in its internal properties under global warming. The microstructure of summer Arctic sea ice changes due to varying external forcing, ice age, and extended melting seasons, which affect its optical properties. Sea ice cores sampled in the Pacific sector of the Arctic obtained by the Chinese National Arctic Research Expeditions (CHINARE) during the summers of 2008 to 2016 were used to estimate the variations in the microstructures and inherent optical properties (IOPs) of ice and determine the radiation budget of sea ice based on a radiative transfer model. Compared with 2008, the volume fraction of gas bubbles in the top layer of sea ice in 2016 increased by 7.5%, and decreased by 50.3% in the interior layer. Meanwhile, the volume fraction of brine pockets increased clearly in the study years. The changing microstructure resulted in an increase in the scattering coefficient in the top ice layers by 9.3% from 2008 to 2016, while an opposite situation occurred in the interior layer. These estimated ice IOPs fell within the range of other observations and their variations were related to increasing air temperature and decreasing ice ages. At the Arctic basin scale, the changing IOPs of ice greatly changed the amount of solar radiation transmitted to the upper ocean even when a constant ice thickness is assumed, especially in marginal ice zones, implying the presence of different sea ice bottom melt processes. These findings revealed the important role of the changing IOPs of ice in affecting the radiation transfer of Arctic sea ice.

1 Introduction

The recent rise in air temperature in the Arctic is almost twice the global average, known as Arctic amplification (Dai et al., 2019), which has been seen in the retreat of sea ice, especially in summer. The extent of sea ice in summer has decreased (Comiso et al., 2008; Parkinson & Comiso, 2013; Petty et al., 2018), and summer ice is thinner (Kwok, 2018), younger (Stroeve and Notz, 2018), and warmer (Wang et al., 2020) than before. These changes have affected the transfer of sunlight into the Arctic Ocean, and the optical properties of sea ice are changing the solar radiation budget in the area.

Variations of Arctic sea ice cover are related not only to the macroscale properties described above but also to the ice microstructure. Sea ice is a multiphase medium consisting of pure ice, gas bubbles, brine pockets, salt crystals, and sediments (Hunke et al., 2011). In the last decades, the length of the Arctic ice melt season has shown a significant positive trend (Markus et al., 2009), and the Arctic ice cover has experienced a transition from predominantely old ice to primarily first-year ice (Stroeve and Notz, 2018; Tschudi et al., 2020). At the same time, in melting ice gas bubbles and brine pockets tend to become larger (Light et al., 2003), and phase changes due to brine salinity and temperature result in variations in the volume of gas and brine (Weeks and Ackley, 1986; Crabeck et al., 2019). Therefore, the physical properties of ice have changed and in the past 10 years the bulk density of summer Arctic sea ice has been lower than reported in the 1990s due to increased ice porosity (Wang et al., 2020).

Gas bubbles and brine pockets, as dominant optical scatterers, directly influence the inherent optical properties (IOPs) of sea ice (Grenfell, 1991; Perovich, 2003). IOPs include scattering and absorption coefficients and information about the phase function of the domain. The varying IOPs of ice have attracted attention due to their important role in the ice–albedo feedback process. Light et al. (2008) and Katlein et al. (2019; 2021) demonstrated clear IOPs variations during the melting season, and difference in the IOPs between first-year ice and multiyear ice have been ascertained in many observations (e.g., Light et al., 2015; Grenfell et al., 2006). There are also differences in the IOPs of first-year ice because of different stages of melting (Veyssière et al., 2022). Yu et al. (2022) carried out a study to investigate the sensitivity of ice IOPs on its microstructures, but they did not show how the ice IOPs change in the real world. Changes in ice microstructure and IOPs are especially important for the Arctic under the general warming climate and decreasing ice age. Even in the latest studies and sea ice models, IOPs are set as constants based on previous field observations (e.g., Briegleb & Light, 2007), which is somewhat in contrast to the reality in the Arctic Ocean.

In this study, *in situ* observations of the physical properties of summer Arctic sea ice during the Chinese National Arctic Research Expeditions (CHINARE) from 2008 to 2016 were employed as input data. Variations of the microstructure and IOPs of Arctic sea ice are presented. Also shown are their quantitative effects on the radiation budget in the area. Applying these varying IOPs to satellite-observed sea ice conditions has allowed us to obtain Arctic-wide estimates of role of ice's microstructure in the radiation budget of the region.

2 Data and method

2.1 Arctic sea ice coring

The Arctic sea ice cores were sampled in the Pacific sector of the Arctic Ocean during summer cruises of the CHINARE

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program from 2008 to 2016 (Figure 1). Each ice core was evenly divided into 10 layers. Detailed volume fractions of the gas bubbles and brine pockets (V_a, V_b) in the ice cores were given by Wang et al. (2020). Almost all cores were sampled in August, when the ice had started to melt.

A typical undeformed sea ice floe consists texturally of three layers due to its growth conditions (Tucker et al., 1992). The first two layers are relatively thin and consist of a granular layer and a transition layer, and the lowest layer generally consists of columnar ice. The ice texture controls the ice microstructure (Crabeck et al., 2016). Thus, the development of gas bubbles, brine pockets, and IOPs in the three ice layers is different. Analogous to the parameterization of the Los Alamos sea ice model (CICE; Briegleb & Light, 2007), the ice was divided into ten vertical layers. The top (1/10) layer of an ice core was defined as the top layer (TL), the second layer (2/10) was the drained layer (DL), and layers 4–10/10 collectively constitute the internal layer (IL). Note that the surface scattering layer (SSL) and part of the DL were mixed in the TL and could not be separated completely. Layer 3/10 was also a mixture of a DL and IL, and is therefore neglected in the following analysis.

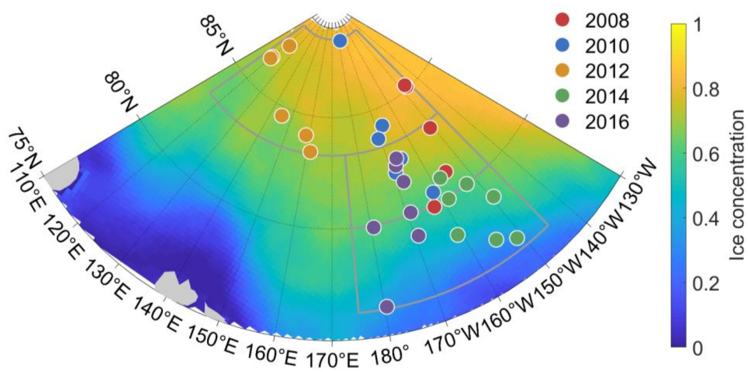

Figure 1. Locations of the sampled ice cores during CHINARE cruises. The ice cores were assorted into three parts according to latitude and ice concentration. Their quantities were nearly the same in each zone. The ice concentration in the base map was the mean in August from 2008 to 2016.

2.2 Sea ice optics modeling

The IOPs of sea ice, including the scattering coefficient, σ, absorption coefficient, κ, and asymmetry parameter, g, can be determined directly from the icemicrostructure. Following the theory of Grenfell (1991), scattering in ice is caused by gas bubbles and brine pockets, and absorption is caused by brine pockets and pure ice. This parameterization has been proved by extensive observations (e.g., Light et al., 2004; Smedley et al., 2020). The IOPs of sea ice can be obtained from the sum of the scatterers weighted by their relative volumes as:

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$$\sigma = \sigma_\mathrm{a} + \sigma_\mathrm{b} = \int_{r_\mathrm{min}}^{r_\mathrm{max}} \pi r_\mathrm{a}^2 Q_\mathrm{a}^\mathrm{sca} N_\mathrm{a}(r) dr + \int_{l_\mathrm{min}}^{l_\mathrm{max}} \pi r_\mathrm{b}^2 Q_\mathrm{b}^\mathrm{sca} N_\mathrm{b}(l) dl \quad (1)$$

$$\kappa = \kappa_\mathrm{i} + \kappa_\mathrm{b} = k_\mathrm{i} V_i + \int_{l_\mathrm{min}}^{l_\mathrm{max}} \pi r_\mathrm{b}^2 Q_\mathrm{b}^\mathrm{abs} N_\mathrm{b}(l) dl \quad (2)$$

$$g = \frac{g_\mathrm{a}\sigma_\mathrm{a} + g_\mathrm{b}\sigma_\mathrm{b}}{\sigma} \quad (3)$$

In these equations, the subscripts a and b represent gas bubbles and brine pockets, respectively, r is their radius (or equivalent radius), and l is the length of the brine pockets. Q^sca and Q^abs are the scattering and absorption efficiencies, respectively, which can be calculated using Mie theory. N is the size distribution function, subscript i represents pure ice, and $V_\mathrm{i} = 1 - V_\mathrm{a} - V_\mathrm{b}$ is its volume fraction. The values of these parameters are summarized in Table 1. Brine pockets longer than 0.03 mm are modelled as cylinders rather than spheres (Light et al. 2003). The conversion function from Grenfell & Warren (1999) is employed to represent hexagon columns as spheres with the same optical properties. Besides, Q^abs and Q^sca in the required size range are obtained using their effective radii, which are calculated according to Hansen & Travis (1974).

Table 1. Parameters used in the radiation transfer model in Arctic summer and their sources

Parameter	Reference(s)
refractive index of gas bubbles	Light et al. (2004)
refractive index of brine pocket	Smith and Baker (1981)
N_a, N_b	Light et al. (2003)
k_i	Grenfell and Perovich (1981)
g_a, g_b	Light et al. (2004)
$r_\mathrm{min} = 0.5$ mm, $r_\mathrm{max} = 2$ mm	Grenfell (1983); Frantz et al. (2019)
$l_\mathrm{min} = 1$ mm, $l_\mathrm{max} = 20$ mm	Light et al. (2003); Frantz et al. (2019)

The Delta-Eddington multiple scattering model, where the constant IOPs from Briegleb & Light (2007) were replaced by the modeled IOPs, was employed to estimate the apparent optical properties (AOPs: albedo α_λ, tansmittance T_λ, and absorptivity A_λ) of the ice at the sampling sites (Yu et al., 2022). The broadband albedo (α_B), transmittance (T_B), and absorptivity (A_B) were calculated by integrating the spectral values over band of the incident solar radiation, F_0 as:

$$X_B = \frac{\int_{\lambda_1}^{\lambda_2} X_\lambda F_0(\lambda) d\lambda}{\int_{\lambda_1}^{\lambda_2} F_0(\lambda) d\lambda}, X = \alpha, T, A, \quad (4)$$

In the following sections, the broadband absorption coefficient, κ_B, was also derived by this equation, following CICE (Briegleb & Light, 2007). Considering the generally cloudy weather in Arctic summer, the incident solar irradiance under an overcast sky in August from Grenfell & Perovich (2008) was chosen as the default value for F_0. The studied wavelength band was set as the photosynthetically active band, i.e. $\lambda_1 = 400$ nm and $\lambda_2 = 700$ nm.

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2.3 Arctic-wide up-scaling

To conduct an up-scaling analysis of the radiative budget of the Arctic sea ice cover based on observations of the ice microstructure in the Pacific sector, we used representative basin-scale sea ice data to estimate the variations in the distribution of radiation fluxes in summer during 2008-2016. The sea ice concentration (C) was provided by the National Snow and Ice Data Center (NSIDC) (Cavalieri et al., 1996), the sea ice thickness was based on CryoSat-2/SMOS data fusion (Ricker et al., 2017), and the incident shortwave radiation flux at the surface (E_d) was obtained from the European Centre for Medium-Range Weather Forecasts (ECMWF). The latter two datasets were interpolated to a 25 km NSIDC Polar Stereographic grid. Then, the mean radiation fluxes and ice concentrations from July to September from 2008 to 2016 were set as the representative values in summer. Due to the limitation of satellite remote-sensing data of summer ice thickness, the representative thickness was estimated according to the mean value in October from 2011 to 2016, together with the growth rate estimated by Kwok and Cunningham (2016). Thus, the reflected, absorbed, and transmitted radiation flux by Arctic sea ice are $E_r = E_d \cdot C \cdot \alpha_B$, $E_a = E_d \cdot C \cdot A_B$, and $E_T = E_d \cdot C \cdot T_B$, respectively.

3 Results

3.1 Microstructure of the ice cores

There were different variation trends in the volume fraction of gas bubbles and brine pockets (V_a, V_b) as a function of ice core depth (Figure 2). The upper granular ice was typically bubbly, associated with the drainage of brines, and the interior columnar ice is usually depleted in gas bubbles (Cole et al., 2004). Thus, a clear decreasing trend along depth could be seen in V_a. The mean V_a of the TL, DL, and IL for all ice cores was 23.7 ± 7.8%, 18.4 ± 7.2%, and 12.1 ± 7.8%, respectively. These values are similar to the observations made by Eicken et al. (1995) where V_a decreased from > 20 % at the top to < 5 % at the bottom for summer Arctic sea ice.

The drainage of brine resulted in a relatively small V_b of TL, with a mean of 3.5 ± 3.3%, while it was 4.7 ± 4.3% and 13.1 ± 8.4% in the other two layers, respectively (Figure 2a). V_b = 5% is usually chosen as a threshold where discrete brine inclusions start to connect and the columnar ice is permeable enough to enable drainage (Carnat et al., 2013). Thus, the ice cores in the present study have been melting for some time, agreeing with the sampling season during CHINARE. Most V_b profiles had a maximum in the middle depth, except for the ice cores in 2012 (Figure 2d). This can be explained by the later sampling date in 2012 relative to the other years by about 10 days, which resulted in enhanced brine drainage. Furthermore, the shape of the V_b profile was also associated with ice age (Notz and Worster, 2009). Compared with the ice cores in 2010, although the ice cores in 2016 had similar sampling dates (one day difference), the maximum position of V_b in 2016 was

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lower than in 2010 (Figure 2c, f). This was because all ice cores in 2010 were sampled from first-year ice, and the ice cores in 2016 were comprised of first-year ice and multiyear ice (Wang et al., 2020).

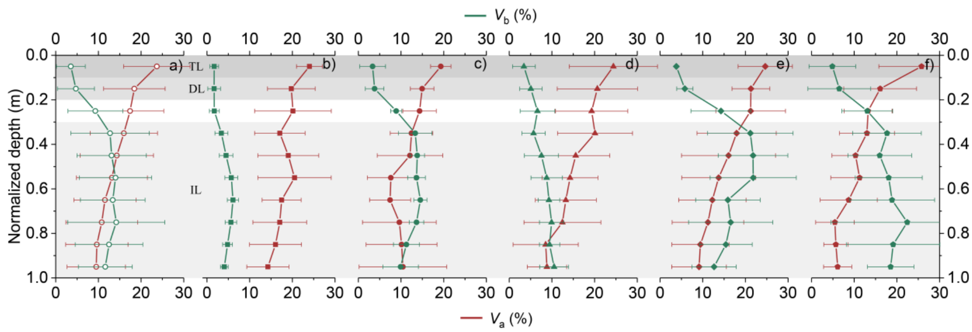

Figure 2. Profiles of V_a and V_b against normalized depth in (a) the whole study period, (b) 2008, (c) 2010, (d) 2012, (e) 2014, and (f) 2016. The error bars showing the standard deviation from the mean of the results. The shady areas represent the ice layer structure.

In addition to the different variations in V_a and V_b with depth, the annual variations in each layer were also different (Figure 3a). V_a was relatively small in the TL of 2010 because all ice cores were sampled from first-year ice (Wang et al., 2020). While the quantities of first-year ice cores were similar to the amount of multiyear ice cores in the other years, the overall trends were still clear. Compared with 2008, the V_a of TL in 2016 increased by 7.5%. This indicated that the melting process of the ice surfaces of the cores in 2016 was more dramatic than in 2008. Contrary to the TL, the V_a in the IL tended to decrease from 2008 to 2016. The corresponding variation ratio was -50.3%. The V_a values of DL were relatively stable, which did not show a clear increase or decrease in the study period.

Things were different for V_b and ice porosity. There were clear increases in the V_b of all three ice layers (Figure 3b), which implied dramatic variations in the permeability of summer sea ice. Furthermore, increases of V_b in the DL and IL were clearer than in the TL. From 2008 to 2016, the increase in the IL was clearest. Meanwhile, the increases in V_b of the TL and DL were similar. Simultaneously, the ice salinity of the IL decreased (Figure S1), which agreed well with the observed and modeled results with warming conditions (e.g., Vancoppenolle et al., 2009). From the combined effects of changing V_a and V_b, it was clear that the porosity of ice was on the increase (Figure 3c). Furthermore, no obvious differences in the developments of porosity were seen in the three layers.

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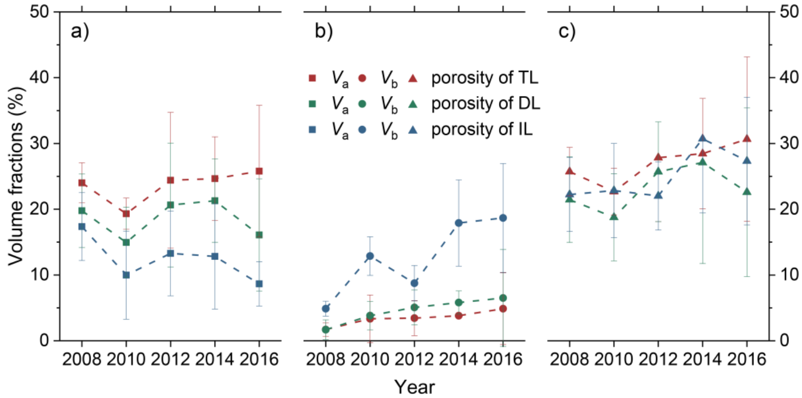

Figure 3. Variations in (a) V_a, (b) V_b, and (c) the porosity of the TL, DL, and IL of the ice cores during 2008 to 2016.

3.2 Variations in the IOPs of the ice cores

The mean scattering coefficient, σ, of the TL, DL, and IL for all ice cores was 272.8 ± 25.8 m^{-1}, 217.3 ± 28.7 m^{-1}, and 162.0 ± 27.4 m^{-1}, respectively (Figure 4a). There was a clear decreasing tendency along with depth in the mean σ of all ice cores, associated with a decreasing volume of gas bubbles (Figure 2). Although the V_b values of the ice cores increased clearly with depth, they did not enhance the scattering capacity of ice. The reason for this was that the refractive indices of brine pockets and pure ice are close (Smith and Baker, 1981; Grenfell and Perovich, 1981).

The vertical variations in κ_B and g were not clear as seen for σ because they depend on V_i and V_b/V_a, respectively. Due to the low porosity of the ice ($V_a + V_b$), κ_B showed a slightly increasing trend with depth, which varied in the range 0.09–0.1 m^{-1}. The mean value of g was 0.93 except in 2008 (which was g = 0.89), and it slightly increase with depth. This value is smaller than the commonly used one; for example, the previous typical range of g was from 0.86 to 0.99 (e.g., Ehn et al., 2008), and 0.94 was often adopted for computational efficiency in models (Light et al., 2008). However, we note that particulate matter and inclusion shape were not considered here, which may be a possible reason for the different values of g found here.

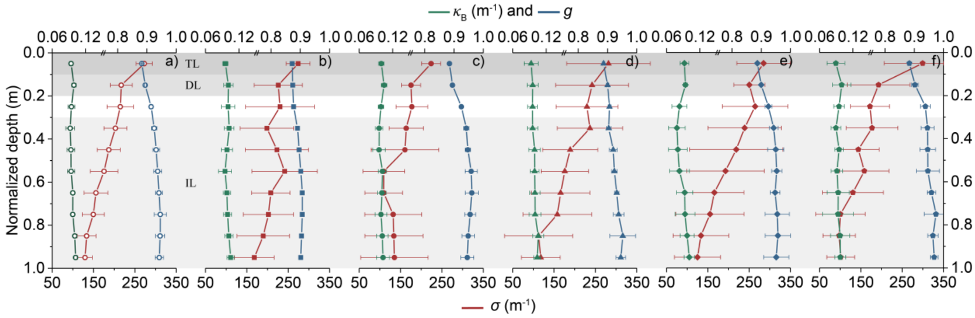

Figure 4. IOP profiles of ice cores against normalized depth in (a) the whole study period, (b) 2008, (c) 2010, (d) 2012, (e) 2014, and (f)

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2016.

The annual IOPs of the TL, DL, and IL of the ice cores are shown in Figure 5. As shown in Figure 5a, the variations in σ of the TL, DL, and IL were different. A generally increasing σ could be identified in the TL. Compared with 2008, the σ of the TL in 2016 increased by 9.3% due to the increased V_a of the ice. A similar phenomenon was observed by Light et al. (2008), where the σ of the ice surface layer increased as melting progressed. The variations of the σ for the DL were not as clear as for the TL due to ongoing drainage. Overall variations in the σ of the IL were contrary to those seen for the TL. Compared with 2008, the σ of the IL in 2016 decreased by 36.4% due to the decreased V_a (Figure 3).

The broadband absorption coefficient, κ_B, of the TL and IL in 2016 decreased by about 6% relative to their values in 2008 (Figure 5b). Similar variations were not clear in the κ_B values of the DL. Because the absorption in sea ice is mainly caused by pure ice and brine pockets, the absorption efficiency is weaker in brine than in pure ice; thus, an increasing porosity results in decreasing κ_B. As shown in Figure 5c, the values of g of the TL and DL were nearly constant. Because their values of V_b were sufficiently small due to drainage (Figure 3b), their values of g are mainly attributed to gas bubbles. In contrast, the values of g of the IL increased by 5% with increasing V_b in the study years (Figure 3b).

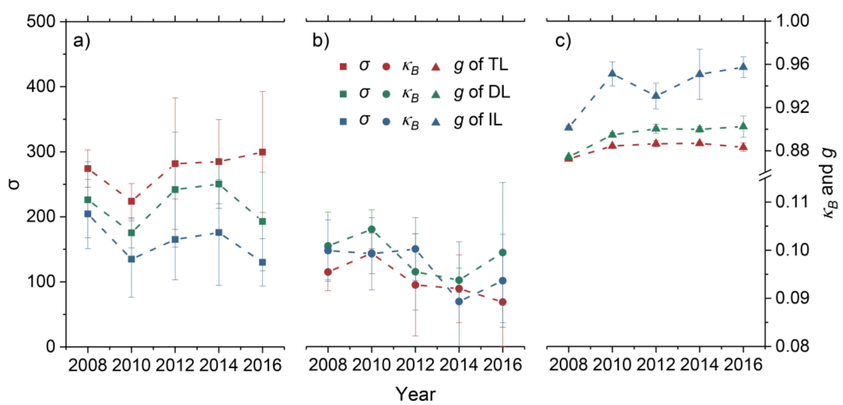

Figure 5. Annual (a) σ, (b) κ_B, and (c) g for the TL, DL, and IL of the ice cores from 2008 to 2016.

3.3 Variations in the AOPs of the ice cores

Having seen that the IOPs of the sea ice were not constant in the different years (Figure 5), a more important question is how these changes affected the AOPs. The AOPs of the sampling sites are shown in Figure 6. Note that the AOPs here were

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based on level ice. Surface properties, such as a snow layer or melt ponds, were not considered here, because the focus was on the effects of the ice microstructure on their AOPs. The results obtained with the same IOPs profiles but for a constant ice thickness (1 m) are also presented to quantify the contributions from the ice microstructure and thickness separately.

The values of α_B changed because of the effects of the ice IOPs and thickness (Figure 6b). The variations in α_B were similar to those in the σ of the TL and DL; i.e., the IOPs of the upper-most ice largely controlled the albedo of bare ice, which was also observed in field observations (e.g., Light et al., 2015). Besides, the α_B values did not present clear decreases. This was different from the remote-sensing results (-0.05 per decade for 1982 to 2009) of Lei et al. (2016). The reason for this was the direct factor that reduces the annual ice albedo is not the ice microstructure but rather the surface conditions. Eicken et al. (2004) and Landy et al. (2015) reported that the evolution of melt ponds on the ice surface could explain 85% of the variance in the summer ice albedo.

Different from α_B, T_B (A_B) tended to increase (decrease) clearly (Figure 6c). The value of T_B in 2016 was over treble of that in 2008. Meanwhile, A_B decreased by about 20.9% from 2008 to 2016. Furthermore, the change of A_B in the study years was lower than the actual change in the ice thickness (-35.3%). Thus the difference, 22.3% ($\frac{1-20.9\%}{1-35.3\%} - 1$), was attributed to an increase in the absorbed solar energy per unit volume of sea ice. The increasing absorbed radiation due to the changing IOPs and thickness of the ice may enhance internal melting.

To make a direct comparison with the above variations, we considered a constant ice thickness, finding that the changes in the AOPs were different but the overall trends were similar (dashed lines in Figure 6). T_B increased from 0.03 to 0.06 from 2008 to 2016, accounting for about 32.9% of the real change with changing thickness. Thus, the changing microstructure of the melting ice resulted in an increased transmittance that was independent of the ice thickness. A similar result was observed in the laboratory, where the changing ice microstructure during the warming process (no decrease in thickness) increased the ice transmittance (Light et al., 2004). Different from T_B, whether the thickness was taken into account or not, the overall trends in α_B and A_B were hardly affected. This demonstrated that the present variations in ice thickness had more effects on the ice T_B than α_B and A_B.

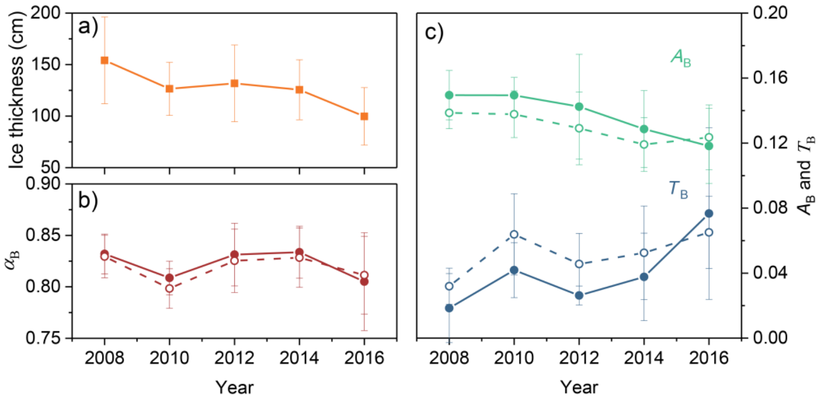

Figure 6. (a) Thickness and (b, c) estimated AOPs of the ice cores from 2008 to 2016. Also shown as dashed lines are the AOPs with the same IOPs and constant thickness (1 m).

3.4 Arctic-wide analysis

In this section, we expand the variations of the ice cores (Figure 5) to an Arctic-wide scale under the following assumptions. First, the IOPs of Arctic ice were based on the results derived from the ice cores. This follows the common approach used in current models, i.e. they are taken as constant and seasonal and spatial differences are ignored (e.g., Briegleb and Light, 2007). Second, a decreasing trend of -5.8 cm yr^{-1} in ice thickness according to Lindsay and Schweiger (2015) was adopted to get a general view of the contributions of the changing ice thickness on the radiation budget. The representative basin-scale sea ice and radaition data in summer were used here to estimate the variations in the distribution of radiation fluxes.

With the combined effects of the changing microstructure and thickness of ice, Arctic-wide variations in the mean α_B, T_B, and A_B (Figure 7a) were clearer than those in Figure 6, especially the overall trends of the mean T_B (r = 0.97, p < 0.01) and A_B (r = -0.97, p < 0.01) of ice. Although the mean α_B decreased during 2008 to 2016, there was not much change in E_r, only about 51.1 W m^{-2} during the study years (Figure 7b). This was because the decreasing α_B was largely provided by marginal ice zones. The decreasing rate of α_B in regions with ice thicknesses < 1 m (equivalent to 16.4% of the entire ice area) was over 1.6 times the rate of the entire ice cover (Figure S2). With the retreat of sea ice, the reflected flux of the marginal zone contributes less and less to the reflected flux of the entire ice cover.

The mean transmitted solar radiation, E_T, increased from 1.8 W m^{-2} to 9.7 W m^{-2} from 2008 to 2016 (Figure 7b). Most of the increase in E_T is ascribed to thin ice in marginal ice zones (ice thicknesses < 1 m), which contributed 49.2% of the increasing E_T from 2008 to 2016 (Figure 8a–e). Meanwhile, E_a decreased from 15 W m^{-2} in 2008 to 13.8 W m^{-2} in 2016. As the decrease in ice volume from 2008 to 2016 was 32.2%, the solar energy absorbed by a unit volume of sea ice increased by 35.7% on the Arctic scale.

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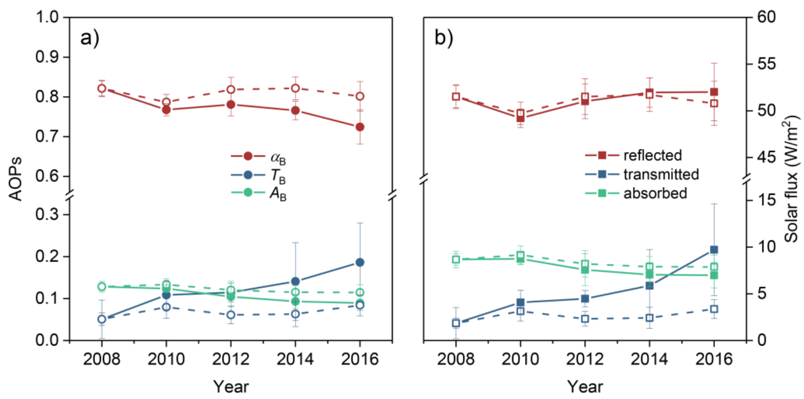

Figure 7. Arctic-wide variations in the mean (a) AOPs of ice and (b) solar flux distribution during 2008 to 2016. Also shown as dashed lines are the AOPs and fluxes with the same IOPs and constant thickness field.

When the ice thickness was set as a constant, variations in the mean AOPs were different, which resulted in differences in the solar flux (dashed lines in Figure 7b). Among them, differences in the reflected flux were relatively small. Meanwhile, the mean E_T increased from 1.8 W m^{-2} in 2008 to 3.4 W m^{-2} in 2016. E_a decreased from 8.7 W m^{-2} to 7.9 W m^{-2} in the same period. These changes corresponded to 19.3% and 46.8% of the combined effects of the ice IOPs and thickness, respectively, from 2008 to 2016. Furthermore, marginal ice zones with ice thicknesses < 1 m still contributed 37.1% of the increasing E_T from 2008 to 2016 (Figure 8f-j). This value was about 75.4% of the rate of the combined effects of the changing IOPs and thickness of ice. In other words, the same changes in the ice microstructure had more effects on the T_B of thin sea ice, and these effects were clearer than those resulting from general decreasing ice thickness.

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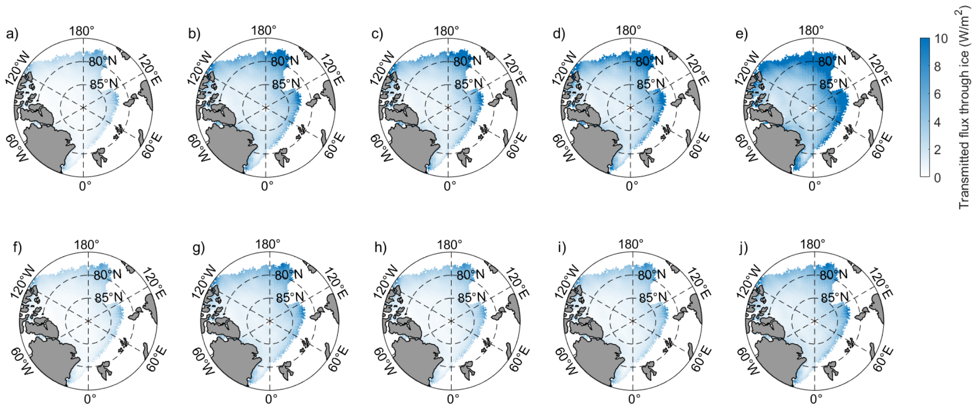

Figure 8. Distribution of transmitted solar radiation through sea ice in the summers of 2008 to 2016 when the sea ice thickness was set (a–e) to decrease and (f–j) to a constant value. Only flux that penetrated through the sea ice is considered in this map.

4 Discussion

4.1 Comparisons with IOP measurements

In Section 3.2, we estimated the ice IOPs according to the observed ice physics and structural-optical theory. Other methods estimated ice IOPs in previous studies. In this section, we compared the ice scattering coefficient, the most variable value among all IOPs, determined in the present study with previous results (Figure 9). The differences in wavelength bands were ignored in the comparisons because σ was nearly wavelength-independent.

It is clear from Figure 9 that the range of σ of the present study covered the majority of previous results. The derived values of σ for the SSL and DL of melting bare ice in August ranged from 920 to 2,000 m^{-1} and 40 to 150 m^{-1}, respectively (Light et al., 2008). According to the layer structure, wherein the TL was composed of a 5 cm SSL and the others were DLs, the bulk σ of the TL in Light et al. (2008) ranged from 270 to 435 m^{-1}. This result was slightly higher than our results. The results of Mobley et al. (1998) and Perron et al. (2021) agree with our range. The σ of the DL in Perron et al. (2021) was in our range, and the values of Light et al. (2008) were smaller than those in the present study.

Differences in the σ of the IL were clearer than in the TL and DL. The σ values of the IL of most our cores were relatively larger than those of Light et al. (2008, 2015) and Frantz et al (2019). In these results, Light et al. (2008) estimated the σ using the observed ice albedo and a three-layer structure with fixed thicknesses. The results of Light et al. (2015) and Frantz et al. (2019) were obtained in a cold laboratory by simulating the radiative transport in subsections of the sea ice. Meanwhile, the results of Grenfell et al. (2006) and Perron et al. (2021) are close to the minimum of our range. The σ of ice in Grenfell et al., (2006) was calculated from the ice extinction coefficient, and it was measured *in situ* using a diffuse reflectance probe in the

Perron et al. (2021). The values calculated by the same method as used in the present study by Mobley et al. (1998) were close to the maximum of our range. Thus, it was expected that the differences in the IL's σ largely resulted from the different methods used in the myriad studies.

One possible reason for the differences was the uncertainties in the ice microstructure introduced by brine loss during measurement and segmenting. Thus, our V_a values of the IL are greater than the values derived from nondestructive methods (e.g., Perron et al., 2021). As a result, the maximum underestimate of V_b was 15–25% and the maximum overestimate of V_a was 96–160% when taking the uncertainties introduced by the measurements and brine drainage into account (Wang et al., 2020). Taking the mean V_a and V_b of all ice cores as an example, these uncertainties overestimated the σ of the IL by 78 m^{-1} at most. Although brine loss during sampling and measurements introduced uncertainties to V_a and V_b, the methods used for obtaining and measuring the ice cores during the CHINARE cruises were the same. Therefore, the uncertainties introduced by the methodology hardly affected the changes seen in Figure 6 and Figure 7.

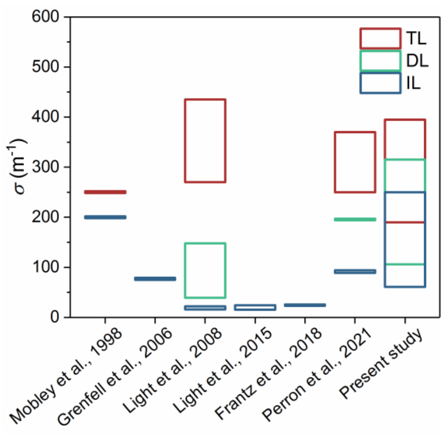

Figure 9. Comparison of the ice scattering coefficient in the present study to the published results for Arctic sea ice using various methods. All comparison results have been scaled to the layer structure used in the current study according to their ice thicknesses.

Another source of difference is the distribution function of gas bubbles employed in the IOP parameterization. Many distributions are obtained in a cold laboratory, where the ice temperature is not consistent with that in the summer Arctic. As the refractive indices of brines and pure ice were similar, the distribution function of brine pockets had a smaller influence on the ice IOPs than gas bubbles (Yu et al., 2022). Here, we tentatively adjusted the exponent of the distribution function of the gas bubbles from its default value of -1.5 to -1, i.e., the fraction of small bubbles decreases, which coincides with warming ice

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(Light et al., 2003). Then, the changed distribution function was used for 1 m thick ice with mean values of V_a and V_b for every ice core. This change resulted in an uncertainty of 8 m^{-1} in the σ of each layer. These uncertainties did not alter the above results and are considered acceptable.

Although brine loss and the difference in the distribution functions of gas bubbles introduced uncertainties in σ, they did not affect the ice AOPs much. Considering a 1 m thick ice layer described by the mean physics of ice cores, the effects of the former factor on the ice AOPs were less than 0.02. The uncertainties in α_B and T_B introduced by the latter factor were 0.005 and 0.002, respectively. Therefore, our estimated α_B range (0.76–0.87) agreed with the observed results of Light et al. (2008, 2015) and Grenfell et al. (2006). Meanwhile, the estimated T_B (0.01–0.1) was also in the corresponding observed ranges.

4.2 On the interannual variations of the IOPs

Extensive measurements of the IOPs of Arctic sea ice have been carried out, and some authors have noticed the seasonal variations of the ice microstructure and IOPs (e.g., Light et al., 2008; Frantz et al., 2019; Katlein et al., 2021). However, interannual variations in sea ice IOPs are still not clear, although such changes in sea ice extent, thickness, and age are evident. A lack of continuous IOP measurements is the primary reason. Compared with previous observations, the ice core data in the present study were more appropriate for interannual analyses of the IOPs of ice because of their long time span and consistencies in the sampling method, seasons, and sea areas. The reasons we could not introduce other ice core data (SHEBA, ICESCAPE, N-ICE, MOSAiC, etc.) into this study was that not only the differences in sampling seasons, sites, and methods increase the dispersion in time and space during such an analysis, but also the lack of information about the ice microstructure or essential physical properties will limit how much we can determine from such a comparison. Considering that sampling ice cores is a commonly used method for *in situ* observations, with more suitable ice core data in the future, large-scale time series of ice IOPs may be obtained.

The ice cores used in the present study were sampled at different ice stations but not at the same floe (Figure 1). Thus, the data did not form a time series in the strictest meaning. Figure 10 illustrates the different IOPs of the ice cores in three latitude zones, which shows that there are spatial differences in the present ice core data. Among the three IOPs, variations in σ are the clearest. The difference in κ_B and g in the different latitude zones were not more than 5% and 2%, respectively (Figure 10b, c). As a transition layer between the TL and IL, variations in the IOPs of the DL were more discrete than in the other two layers. For σ, there were no clear changes in the TL. This demonstrated that the variations of σ in the TL largely resulted from interannual factors. Meanwhile, the differences in the σ of the IL are clear. With an increase of latitude, the σ of the IL tended to increase. The rate of spatial variation of σ in the IL can be up to 16.5% in the different latitude zones. This value was equivalent to 45.8% of the maximum rate of interannual variation seen in Figure 5.

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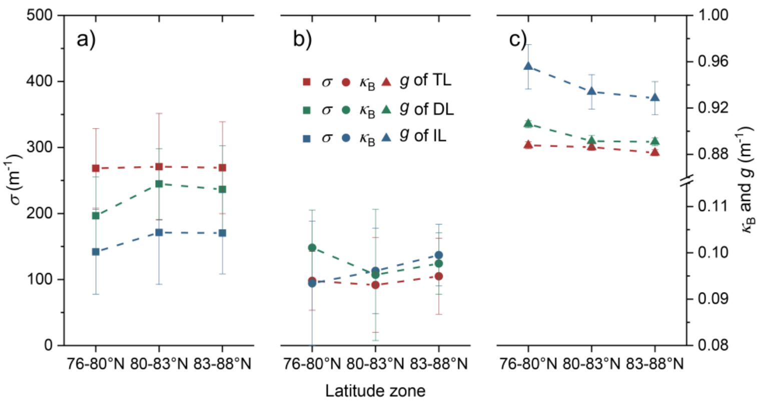

Figure 10. Different values of (a) σ, (b) κ_B, and (c) g for the TL, DL, and IL of the ice cores in the three latitude zones.

The melting days of sampling sites, which were calculated according to the sampling date (Aug. 18 ± 9 days) and melt onset from Markus et al. (2009), were similar (58 ± 7 days). The amount of surface radiation during the study years was also similar (Laliberté et al., 2021). However, the observed air temperature increased continuously (Figure 11a). This clear difference in the temperatures of the sampling date can hardly be explained by spatial variations, but agree more with the interannual variations (e.g., Collow et al., 2020). Previous observations demonstrated that the evolution of the σ of the SSL is highly affected by warming processes (e.g., Light et al., 2008; Smith et al., 2022). Except for the ice cores in 2010, the variations in the σ of the TL in the present study agreed with the increasing air temperature (Figure 11a).

There were clear differences in the σ of the IL (Figure 5a) which can hardly be explained by increasing air temperature because the σ of the IL is relatively constant (Light et al., 2008) and even increases in the melt season (Frantz et al., 2019). Figure 11b shows the correlations among the annual σ of the IL, ice age, and T_B. Note that T_B here is the result under the assumption of a constant ice thickness (dashed line in Figure 6c). The ice ages were obtained according to fieldwork (Wang et al., 2020) and remote-sensing data (Tschudi et al., 2019). Because the ice age of each grid cell in the remote-sensing data is represented as the age of the oldest floe, once an ice core was distinguished as first-year ice in the fieldwork, the corresponding ice age was set as one year regardless of the remote-sensing data. The use of remote-sensing data is acceptable because the ice cores in this study were all sampled in large and thick floes for safe fieldwork. These floes were more likely older than the surrounding ice. Figure 11b demonstrates that the decrease in the σ of the IL is correlated with decreasing ice age. The σ of the IL in the first-year ice was smaller than in multiyear ice, which was also observed by Light et al. (2015). This could partly explain the spatial variations in the σ of the IL (Figure 10a) because ice in high latitude zones was likely older than in the other

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zones (Stroeve and Notz, 2018).

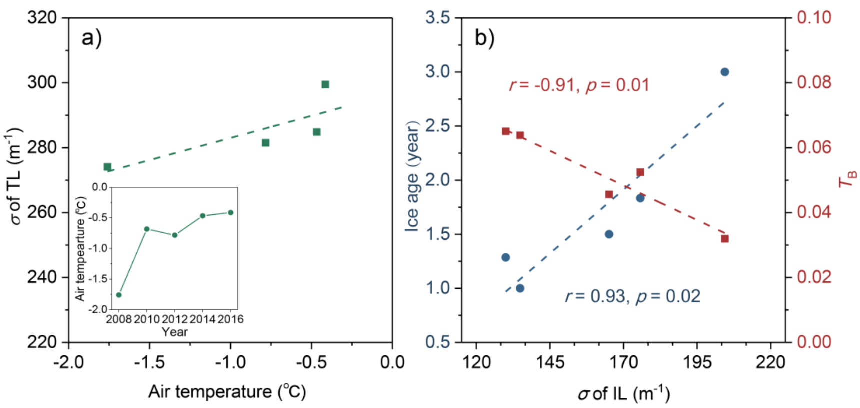

Figure 11. (a) Changing air temperature at the sampling sites and its effects on the σ of the TL. (b) Correlations among the σ of the IL with ice age and T_B.

In summary, the differences in the IOPs of the ice cores were related to interannual variations in the air temperature and ice age. The changing ice age largely manifested in the ice microstructure in the IL. In other words, the changing ice age may be partly responsible for the modeled results shown in Figure 7, even without any decrease in the ice thickness. To our knowledge, this is the first study to link ice microstructure and IOPs at interannual scales. Although these ice core data are not a time series in the strictest meaning, they are still helpful for understanding the general effects of the scenario where the Arctic is warming and the ice ages are decreasing. Our results suggested that in this scenario, the σ values of the TL and IL of summer ice tended to be greater and smaller than before, respectively. It is expected leading to interannual trends of the ice microstructure and IOPs. This process needs to be confirmed by future observations and more simulations.

4.3 Implications for the future Arctic

Previous studies have reported that surface properties (snow, ponds, etc.) largely control the variations in the ice albedo (Landy et al., 2015). The present results also asserted that interannual variations in the ice's microstructure or IOPs had little effect on the albedo of bare ice (< 2%), but they do play an important role in ice transmittance (Figure 6). With continued Arctic warming, the summer ice age is on the decrease, and the ice microstructure and IOPs change accordingly, leading to an overall higher ice transmittance. Furthermore, the transmitted solar energy affects the temperature of the upper ocean and results in further melting of the bottom of sea ice (Timmermans, 2015). Along with the melting of ice, gas bubbles and brine

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pockets change simultaneously (Light et al., 2004), which affect the IOPs of ice in turn. Consequently, the sea ice is expected to become thinner and more porous than before. This process has been seldom considered in previous studies. Related studies generally regarded the surface properties and thickness of the ice as predictors for light transmittance (e.g., Katlein et al., 2015; Perovich et al., 2020). The microstructure and morphological parameters of sea ice (e.g., thickness, extent, etc.) may together influence the melting processes of Arctic sea ice.

For safe field observations, the ice core data used in this study were all sampled in large and thick floes. Variations in the microstructure of the ice in marginal zones or under melt ponds cannot be addressed by this study. Light et al. (2015) reported that the differences in the σ between the IL of ponded first-year ice and multiyear ice were larger than the those between bare first-year ice and multiyear ice. The changes in the IOPs of the marginal ice zone were expected to be more obvious than found in the present results because ice in marginal zones is more likely young and ponded (Rigor and Wallace, 2004; Zhang et al., 2018). Furthermore, the same changes in the ice microstructure have more effects on the T_B of thin sea ice (Section 3.4). Marginal ice zones, comprising 16.4% of the entire ice area, contributed 37.1% of the extra-transmitted solar energy due to the ice changing microstructure from 2008 to 2016 (Figure 8). Both processes promote an increase of transmitted flux through sea ice and ice bottom melting in marginal ice zones. Arndt & Nicolaus (2014) quantified light transmittance through sea ice into the ocean for all seasons as a function of variable sea ice types. The mean annual trend was 1.5% per year, which mainly depended on the timing of melt onset. If the variations in the microstructure of bare and ponded ice are taken into consideration, this trend is expected to increase. We suggest that future ice observations and models should pay more attention to variations in the ice age, microstructure, and their effects, especially in marginal ice zones.

5 Conclusions

Based on ice cores sampled during the CHINARE expeditions (2008–2016), interannual variations in the IOPs of Arctic sea ice in summer due to the changing microstructure of ice were modeled according to structural-optical theory. Variations in the AOPs and solar flux distribution due to the changing IOPs in the summer Arctic were also estimated. Clear variations in the microstructure and IOPs of each year (Figure 5) enabled us to construct a quantitative view of changes that the Arctic sea ice interior underwent in these years.

As a result of our study, it was found that interannual variations in V_a, V_b, and the IOPs of three ice layers were different, and the overall trends of V_a in TL and IL were nearly the opposite. Meanwhile, increases in V_b could be seen in each layer, and the increase in IL was the clearest. These changes in microstructure were related to the increasing air temperature and decreasing ice ages. The ice σ were highly affected by V_a, which results in the different interannual changes in the σ of the

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different layers. The bare ice albedo was mainly controlled by the σ of the upper layers (TL and DL), and the transmittance is highly correlated with the σ of the IL. With the combined effects of changing microstructure in each ice layer, the interannual variations in the transmitted radiation through the ice were clearer than in the albedo of bare ice, especially in marginal ice zones (Figure 8).

Previous studies paid more attention to changing transmittance due to declining ice thickness. The present findings demonstrated that the changing IOPs derived from the ice microstructure could also alter the partitioning of solar radiation in sea ice by itself. With continued Arctic warming, summer ice will become younger and more porous than before, leading to more light reaching the upper ocean. This reminds us to pay more attention to the variations in the IOPs of interior ice, especially ice with different ages.

Acknowledgments. We are grateful to the NSIDC, Alfred Wegener Institute, and ECMWF for providing the sea ice and radiation data. This work was financially supported by the National Key Research and Development Program of China (Grant number 2018YFA0605901), the National Natural Science Foundation of China (Grant numbers 41922045, 41906198, 41976219, and 41876213), and the Academy of Finland (Grant numbers 333889, 325363, and 317999). We also wish to acknowledge the crews of the R/V Xuelong for their fieldwork during CHINARE.

Author contributions. MY carried out the estimations and wrote the paper. RL, BL, and QW provided the ice core data. All coauthors discussed the results and edited the manuscript.

Data Availability Statement. The sea ice and radiation data are available at https://nsidc.org/data; https://data.meereisportal.de/gallery/index_new.php?lang=en_US&ice-type=extent&active-tab1=measurement&active-tab2=thickness; https://cds.climate.copernicus.eu/cdsapp#!/home.The ice cores data applied in this work can be accessed by the link: https://doi.org/10.5281/zenodo.6693981.

Competing interests. The authors declare that they have no conflict of interest

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