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Key Points:

- Hollow glass microspheres (HGMs) darken surfaces of albedo >0.61
- Spreading HGMs over all Arctic sea-ice types would warm the climate and speed the loss of sea ice
- A layer of hypothetical non-absorbing HGMs applied during May could cool the Arctic by 3 Wm^{-2}

Correspondence to:

M. A. Webster,
mwebster3@alaska.edu

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Regional Geoengineering Using Tiny Glass Bubbles Would Accelerate the Loss of Arctic Sea Ice

Melinda A. Webster¹  and Stephen G. Warren² ¹Geophysical Institute, University of Alaska Fairbanks, Fairbanks, AK, USA, ²Department of Atmospheric Sciences, University of Washington, Seattle, WA, USA

Abstract Arctic sea ice might be preserved if its albedo could be increased. To this end, it has been proposed to spread hollow glass microspheres (HGMs) over the ice. We assess the radiative forcing that would result, by considering the areal coverages and spectral albedos of eight representative surface types, as well as the incident solar radiation, cloud properties, and spectral radiative properties of HGMs. HGMs can raise the albedo of new ice, but new ice occurs in autumn and winter when there is little sunlight. In spring the ice is covered by high-albedo, thick snow. In summer the sunlight is intense, and the snow melts, so a substantial area is covered by dark ponds of meltwater, which could be an attractive target for attempted brightening. However, prior studies show that wind blows HGMs to the pond edges. A thin layer of HGMs has about 10% absorptance for solar radiation, so HGMs would darken any surfaces with albedo >0.61 , such as snow-covered ice. The net result is the opposite of what was intended: spreading HGMs would warm the Arctic climate and speed sea-ice loss. If non-absorbing HGMs could be manufactured, and if they could be transported and distributed without contamination by dark substances, they could cool the climate. The maximum benefit would be achieved by distribution during the month of May, resulting in an annual average radiative forcing for the Arctic Ocean of -3 Wm^{-2} if 360 megatons of HGMs were spread onto the ice annually.

Plain Language Summary A proposal was made to halt Arctic sea-ice loss by spreading glass powder (hollow glass microspheres, HGMs) over the ice to brighten its surface. We calculate the changes of solar energy that would result by considering the areas and reflectivities of different surfaces, as well as sunlight, clouds, and the interaction of HGMs with sunlight. HGMs can make thin ice more reflective, but thin ice occurs in autumn and winter when there is little sunlight. In spring the ice is covered by thick snow. In summer the sunlight is intense, and the snow melts, so large areas are covered by meltwater ponds, which could be a target for brightening. However, prior studies show that winds blow HGMs to the pond edges. Because HGMs absorb some sunlight, spreading them onto sea ice would darken bright surfaces such as snow-covered ice. The result is the opposite of what was intended: spreading HGMs would warm the Arctic climate and speed sea-ice loss. If *non-absorbing* HGMs could be manufactured, they could cool the climate. The maximum benefit would be achieved by distribution in May, cooling the Arctic by 3 Wm^{-2} if 360 megatons of HGMs were spread annually onto the ice.

1. Introduction

The recent rapid decline of Arctic sea ice has inspired proposals for regional geoengineering intended to slow or reverse the decline. In contrast to the global geoengineering proposals under current study, such as stratospheric aerosol injection and marine-cloud brightening, the regional geoengineering schemes propose interventions directed only at the Arctic. They can be classified into two categories: increasing the ice growth rate in winter, or reducing the melt rate in summer.

The first option is based on the recognition that, in nature, the way that sea ice thickens in winter is by freezing seawater to the base of existing ice. The latent heat of freezing must be conducted upward through the ice, so the growth rate diminishes as the ice thickens. To bypass this resistance, it has been proposed to pump seawater from below the ice in winter, and spray it onto the upper surface, where it can freeze, losing heat rapidly by emission of thermal radiation to the atmosphere (Desch et al., 2017). This proposal has been critiqued by others (Pauling & Bitz, 2021; Zampieri & Goessling, 2019); we do not consider it further.

To reduce sea-ice melt in summer, it has been proposed to increase the surface albedo by spreading a layer of hollow glass microspheres (HGMs) over the sea ice (Field et al., 2018); this is the proposal we evaluate here.

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It goes by the names “Ice911 Research” and “Arctic Ice Project.” It is reminiscent of an earlier idea for ships to inject hydrosols (microbubbles) into surface seawater, to create whitewater (Seitz, 2011).

The HGMs used by Field et al. are the K1 product developed by the 3M company (https://www.3m.com/3M/en_US/p/d/b40064615/); they have a median diameter of 65 μm and a wall-thickness of 1 μm . Their bulk spectral radiative properties were measured in a laboratory experiment (Figure 1 of Field et al., 2018). Averaged over broadband wavelengths, a 331- μm layer has reflectance 0.43, transmittance 0.47, and therefore absorptance 0.10. In a field experiment on a frozen pond in Minnesota, a 300- μm layer of this glass powder raised the albedo of thin bare ice by ~ 0.2 .

The remainder of the Field et al. (2018) paper was an exercise in climate modeling. The authors arbitrarily increased the albedo of all Arctic and Antarctic sea ice, regardless of surface type, by ~ 0.07 in the National Center for Atmospheric Research's Community Earth System Model (CESM) Version 1.2, and examined the consequences for sea-ice area and volume over a 40-year time-span. A prior modeling study by Cvijanovic et al. (2015), also using the CESM, had similarly examined the climatic impacts of an arbitrary increase in albedo. We think the response of the CESM does not deserve critical attention because the applied forcing was unrealistic. So rather than discussing the model response in Field et al. (2018), here we instead critique their specified forcing. As shown below, since the HGMs significantly absorb sunlight, applying them would actually *darken* high-albedo surfaces such as snow, giving a forcing opposite to what was intended. The albedo of thin ice with no snow cover, which was studied by Field et al. (2018), is low and can be raised by HGMs, but that surface type occurs on the Arctic Ocean mostly during times of the year when sunlight is minimal and the albedo is therefore irrelevant.

To estimate the radiative forcing (RF) achievable by spreading HGMs over the Arctic Ocean, we must consider the various surface types that are largely representative of Arctic sea ice, with their spectral albedos, and how their seasonal cycles relate to the seasonal cycle of sunlight. That is the subject of this paper.

2. Data

2.1. Surfaces of the Arctic Ocean

There are several surface types, each of which has a different seasonal cycle, as summarized in Table 1. They also differ greatly in their albedos, summarized in Table 2. In this section we discuss each of the types as we proceed through the seasons.

In autumn, when the surface water of the ocean cools to the freezing temperature of -1.8°C , ice begins to form. New ice (“nilas”) is dark, and its albedo would increase with HGM application. But there is very little sunlight, so the albedo has minimal relevance. Furthermore, this new ice usually experiences snow accumulation within a few days after formation (Figure 1a), and just a few millimeters of snow dramatically raise the albedo (Figure 2 of Brandt et al., 2005), so only a small fraction of the ice area would remain dark.

Proceeding from autumn to winter, the snow depth increases rapidly (Webster et al., 2014). Ice motion opens small leads, which quickly freeze over to form nilas. There is little sunlight; the average incident solar radiation at the ocean surface over latitudes $70\text{--}90^{\circ}\text{N}$ is only 8.5 Wm^{-2} for the six months October–March.

In spring and summer, the surface insolation becomes substantial, averaging 177 Wm^{-2} . In spring, the ice is mostly snow-covered, with a high albedo >0.8 , so adding HGMs cannot increase it, but in fact slightly *darkens* the snow, as shown below. In late spring and early summer, the snow melts; the meltwater collects in low parts of the floating sea-ice cover as “melt ponds” (Figure 1b). The ponds occupy troughs between areas of drained ice, which develops a surface scattering layer (SSL). The SSL has albedo ~ 0.62 (Light et al., 2008, 2015), and we show below that it will be essentially unchanged if covered with HGMs.

Melt ponds grow in late May and early June from the melting of snow and sea ice; then they partially drain both vertically and laterally. It is on the melt ponds that HGMs might seem to provide a significant benefit, because melt ponds satisfy three criteria: they have low albedo, they can cover a large area, and they occur in summer when the incident sunlight is intense. However, we expect that the ponds will not remain covered; wind will blow the glass bubbles to the pond edges where they will clump, as was observed by Field et al. (2018) in their Minnesota pond experiment after the ice melted.

In this study, we determine the radiative effect of applying HGMs on the eight major surface types of Arctic sea ice: (a) thick sea ice with a cold, thick snow cover, (b) nilas with no snow cover, (c) nilas with a cold, thin

Table 1
Monthly Fractional Areas of Different Surface Types of Arctic Sea Ice, as a Percent of the Total Ice-Ocean Area Over 70–90°N

	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec
Sea ice, cold thick snow	81 <i>2</i>	81 <i>2</i>	82 <i>2</i>	83 <i>2</i>	40 <i>20</i>	-	-	-	-	-	36 <i>1</i>	59 <i>2</i>
Nilas, no snow	10 <i>2</i>	10 <i>2</i>	9 <i>2</i>	9 <i>2</i>	5 <i>2</i>	-	-	-	3 <i>1</i>	4 <i>1</i>	-	-
Nilas, cold thin snow	10 <i>2</i>	10 <i>2</i>	9 <i>2</i>	9 <i>2</i>	5 <i>2</i>	-	-	-	29 <i>1</i>	33 <i>1</i>	28 <i>1</i>	21 <i>2</i>
Sea ice, melting thick snow	-	-	-	-	40 <i>20</i>	35 <i>10</i>	-	-	-	-	-	-
Melting sea ice, no snow	-	-	-	-	-	30 <i>10</i>	44 <i>5</i>	31 <i>4</i>	-	-	-	-
Melt ponds	-	-	-	-	-	5 <i>5</i>	11 <i>5</i>	8 <i>4</i>	3 <i>2</i>	-	-	-
Refrozen melt ponds	-	-	-	-	-	-	-	-	3 <i>2</i>	-	-	-
Leads and open water	-	-	-	-	11 <i>1</i>	30 <i>1</i>	45 <i>1</i>	61 <i>1</i>	33 <i>1</i>	12 <i>1</i>	-	-
Sea ice, cold thin snow	-	-	-	-	-	-	-	-	30 <i>1</i>	51 <i>1</i>	36 <i>1</i>	20 <i>2</i>

Note. In each box, the second number (in italics) is the uncertainty.

snow cover, (d) thick sea ice with a melting thick snow cover, (e) melting sea ice with a SSL, (f) melt ponds, (g) refrozen melt ponds, and (h) thick sea ice with cold, *thin* snow. We estimate the monthly coverage of each surface type using passive microwave (PM) sea-ice concentrations (Meier et al., 2021), observed monthly coverage of melt ponds (Fetterer & Untersteiner, 1998; Nazintsev, 1964; Perovich et al., 2002; Webster et al., 2015, 2022), and the seasonal evolution of sea-ice surface conditions. Table 1 gives the monthly fractional area of each surface type for 70–90°N, which has a total ice-ocean area of $\sim 11 \times 10^6$ km². For monthly coverage, we make the following empirically based assumptions, designed to give HGMs the benefit of the doubt:

Table 2
Albedos of Surface Types for Visible (0.3–0.7 μ m), Near-Infrared (0.7–2.5 μ m), and Broadband (0.3–2.5 μ m) Based on Observations

Season	Surface type	Visible	NIR (clear)	NIR (cloudy)	Broadband (clear)	Broadband (cloudy)
Spring (pre-melt conditions)	Sea ice with cold thick snow	<i>0.96</i>	<i>0.65</i>	<i>0.72</i>	0.81	0.87
	Nilas, no snow	<i>0.18</i>	<i>0.11</i>	<i>0.12</i>	0.15	0.16
Summer (melting conditions)	Sea ice with melting thick snow	<i>0.91</i>	<i>0.56</i>	<i>0.64</i>	0.74	0.81
	Melting sea ice, no snow	0.74	0.40	0.46	0.58	0.63
	Melt ponds	0.27	0.08	0.09	0.18	0.20
	Leads/open water	<i>0.07</i>	<i>0.06</i>	<i>0.06</i>	0.07	0.07
	Sea ice with cold thin snow	<i>0.94</i>	<i>0.65</i>	<i>0.72</i>	0.80	0.86
Autumn (freezing conditions)	Refrozen melt ponds	0.31	0.14	0.16	0.23	0.25
	Nilas with cold thin snow	<i>0.50</i>	<i>0.33</i>	<i>0.36</i>	0.42	0.45

Note. The values for melting sea ice, melt ponds, and refrozen melt ponds are from the Impacts of Climate on the Eco-Systems and Chemistry of the Arctic Pacific Environment (Light et al., 2015) and Multidisciplinary drifting Observatory for the Study of Arctic Climate (Light et al., 2022; Smith et al., 2021) expeditions. The values in italics are from Table 3 of Brandt et al. (2005).

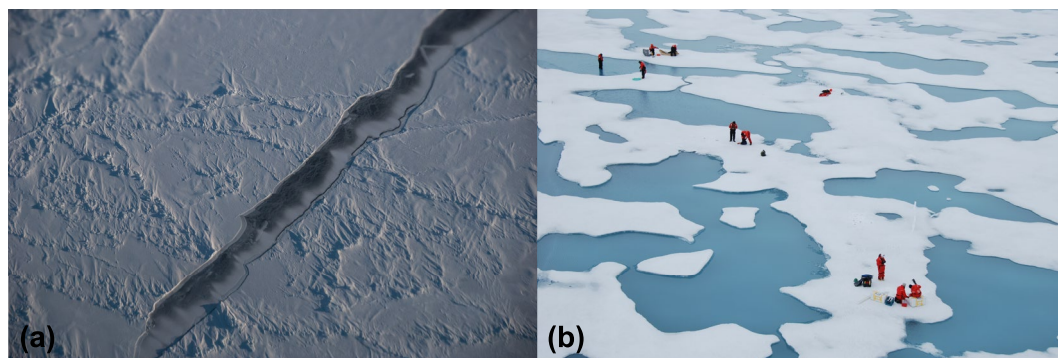


Figure 1. (a) An aerial photograph of a newly frozen lead, in which nilas had formed, and, soon after, became partially snow-covered due to wind-blown snow. The photo was taken on 26 March 2017. (b) Melt ponds on Arctic sea ice, surrounded by drained ice with a surface scattering layer. The photo was taken on 6 July 2011 during the Impacts of Climate on the Eco-Systems and Chemistry of the Arctic Pacific Environment expedition.

- *From January to April*, when most of the Arctic Ocean is covered with snow-covered thick ice, we assume that the open-water area detected from PM consists of nilas, because PM cannot distinguish open water from very thin ice (Ivanova et al., 2015), and the air temperatures in these months are cold, so that newly exposed leads quickly freeze. At any one time, half of the nilas area is snow-free while the other half is covered by a cold, thin snow cover (by “thin” we mean <3 cm). In reality, most nilas would become completely snow-covered after a short period due to snowfall as well as the deposition of wind-blown snow (Figure 1a). Also, extensive open water is normally present in the Barents and Kara seas.
- *In May*, melt is stronger at lower latitudes than at higher latitudes. To account for this, we partition half of the PM open-water area as leads and open water, and the remaining half as nilas. Half of the nilas area is snow-free, while the other half has a cold, thin snow cover. Half of the thick sea ice has a melting, thick snow cover while the other half has a cold, thick snow cover.
- *In June*, the entire PM open-water area is considered open water. Half of the ice is snow-free while the remaining half has a melting snow cover.
- *In July and August*, all sea ice is snow-free. In reality, some snow persists into late summer (Radionov et al., 1997; Warren et al., 1999; Webster et al., 2022).
- *In September*, half of the PM open-water area is leads while the other half is new nilas, with 90% of the nilas being covered in cold, thin snow due to frequent snowfall in autumn (Webster et al., 2014). Half of the melt-pond area is composed of refrozen melt ponds (Perovich et al., 2002; Webster et al., 2022).
- *In October*, 25% of the PM open-water area is leads, while the other 75% is nilas, with 90% of the nilas being covered by cold, thin snow. All other sea ice is covered by cold, thin snow.
- *In November*, half of the thick-ice area has a cold, thin snow cover while the other half has a cold, thick snow cover. The remaining ice-ocean area (28%) is nilas.
- *In December*, 75% of the thick-ice area has a cold, thick snow cover, while the remaining 25% has a cold, thin snow cover. Thereafter, all sea ice apart from nilas is covered by cold, thick snow until May.

The monthly coverages of different ice types in Table 1 are designed to give HGMs the benefit of the doubt within their uncertainty range. The largest uncertainties occur for sea ice with thick snow in May–June due to the high variability in the date of melt onset (Bliss & Anderson, 2018). Melt-pond coverage is also highly variable from year to year and throughout the melt season (Rösel & Kaleschke, 2012), but, as shown below, HGMs have negligible effect on melt ponds. Other ice types, such as nilas, refrozen ponds, and leads, have small (1%–2%) uncertainties in their monthly coverage. The uncertainties associated with lead and sea-ice coverages are based on uncertainties in the PM retrievals (Meier & Stewart, 2019). We incorporate these uncertainties, as well as other uncertainties associated with albedo, cloud cover, and solar radiation, into the calculations below.

2.2. Albedos of Arctic Sea-Ice Surfaces

Table 2 gives, for each surface type, the albedos for three wavelength-bands: visible, near-infrared (NIR), and broadband (a weighted average of visible and NIR). Some of these values come from previously published work (Brandt et al., 2005; Zatko & Warren, 2015). Others were calculated using measurements of spectral albedo

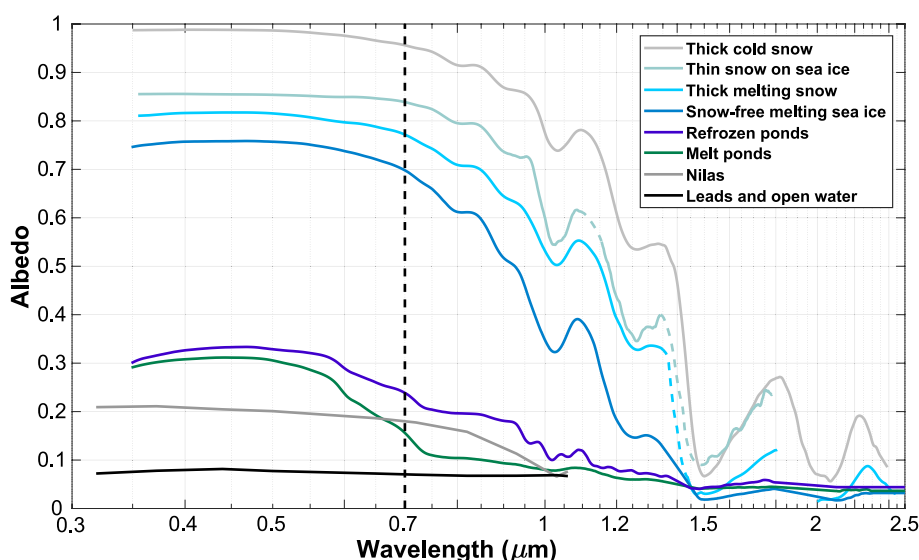


Figure 2. Spectral albedos from in-situ measurements. The values for melting sea ice, melt ponds, and refrozen melt ponds are from the Impacts of Climate on the Eco-Systems and Chemistry of the Arctic Pacific Environment (Light et al., 2015) and Multidisciplinary drifting Observatory for the Study of Arctic Climate (Light et al., 2022; Smith et al., 2021) expeditions. The values for thick cold snow are from Hudson et al. (2006); the remaining values are from Brandt et al. (2005) and Zatko and Warren (2015). The vertical dashed black line at $\lambda = 0.7 \mu\text{m}$ separates the visible and near-infrared bands. The other dashed lines represent interpolated values.

shown in Figure 2, some of which were made on two Arctic expeditions: (a) the Multidisciplinary drifting Observatory for the Study of Arctic Climate (Light et al., 2022; Smith et al., 2021), and (b) NASA's Impacts of Climate on the Eco-Systems and Chemistry of the Arctic Pacific Environment (Light et al., 2015). In these calculations, the visible (0.3–0.7 μm) and NIR (0.7–2.5 μm) band albedos were first determined and then weighted using the normalized solar spectra for clear and overcast sky conditions (Grenfell & Perovich, 2008) to produce the broadband (0.3–2.5 μm) albedo.

Observations from the Arctic Climate Atlas (Environmental Working Group, 2000) show that the cloud cover for the ocean part of the Arctic is largely constant at 80% across all months for 70–90°N; the NIR albedos were weighted accordingly for cloudy (0.8) and clear-sky (0.2) conditions. The cloud climatology of Eastman and Warren (2010) has smaller cloud cover in spring, about 70% averaged over latitude. Using that value would cause our computation of top-of-atmosphere RF by HGMs to be smaller by about 3%. This difference is small compared to other sources of uncertainty in our calculation. The uncertainty in monthly average total cloud cover we use in our calculations is based on the reported interannual variability over high-latitude oceans, about 5% (Figure 2 of Eastman et al., 2011).

Here, we briefly summarize the broadband albedos (weighted averages of clear and cloudy), from the darkest to most reflective surfaces. Open water has albedo 0.06–0.07, which is independent of wavelength across the solar spectrum (Pegau & Paulson, 2001; Zatko & Warren, 2015). Nilas with no snow is the second darkest surface, with albedo 0.15. Melt ponds have albedo ~ 0.20 ; it is higher than that of open water and nilas because scattering of solar radiation by the ice below the pond contributes to the albedo (Grenfell & Perovich, 1984; Langleben, 1971; Light et al., 2022; Perovich & Polashenski, 2012). Refrozen melt ponds, often having a scant layer of snow, have an albedo of 0.25, only slightly higher than that of open melt ponds. Nilas with a thin snow cover has an albedo of 0.44. Snow-free melting sea ice has a substantially higher albedo (0.62) due to the presence of the SSL (Light et al., 2008, 2015, 2022).

Clouds absorb NIR radiation, thus weighting the broadband albedo of a surface below the cloud more toward its visible albedo. The visible and NIR albedos differ more for snow than for the other surface types, so consequently the clear and cloudy broadband values differ more notably for snow than for the other ice types. Under clear sky, the albedo of cold thick snow is 0.81, and for thick melting snow is 0.74. The values are much higher under clouds, giving broadband albedos 0.86 and 0.79, respectively, for the average of 20% clear and 80% cloudy (Table 2).

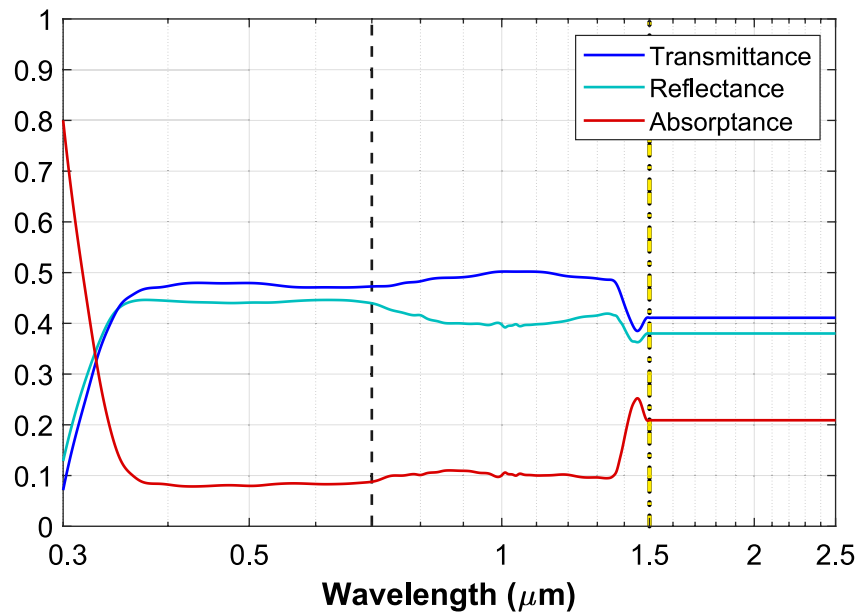


Figure 3. Reflectance (r), transmittance (t), and absorptance (a) as functions of wavelength (λ), for a 331- μm layer of hollow glass microspheres. The values of r and t from 0.3 to 1.5 μm are from Figure 1 of Field et al. (2018). The absorptance at each wavelength is obtained by requiring $r(\lambda) + t(\lambda) + a(\lambda) = 1$. The black vertical dashed line at $\lambda = 0.7 \mu\text{m}$ separates the visible and near-infrared regions. The yellow vertical dash-dot line at $\lambda = 1.5 \mu\text{m}$ marks the longwave limit of the data in Field et al. (2018). For 1.5–2.5 μm we assume that r , t , and a are constant at their values for 1.5 μm . Under clear sky, the fraction of the incident solar energy at $\lambda > 1.5 \mu\text{m}$ is only 8% (or 4% under cloudy sky).

For each of these broadband albedos, we estimate an uncertainty. The uncertainty of an individual measurement is small; the major cause of uncertainty in regional averages is instead the spatial and temporal variation of scattering properties within the defined type, such as variation of snow grain size or variation of nilas thickness. Based on reported measurements from different times and locations, we estimate uncertainties of 0.04 for dry snow, 0.02 for nilas, and 0.01 for open water; intermediate albedos for other surface types have intermediate uncertainties.

2.3. HGM Optical Properties

We use the term “reflectance” to mean the property of a layer, and “albedo” the property of a system, which may consist of several layers. Albedo is the ratio of upward to downward solar flux; it may be measured at the surface, at the cloud top, or at the top of the atmosphere. We use the symbol r for both reflectance and albedo.

The reflectance r and transmittance t of HGM layers were measured as functions of wavelength λ in a laboratory experiment by Field et al. (2018). We use the values from their thickest layer, 331 μm . The results presented in their Figure 1 cover the wavelength range 0.3–1.5 μm , and are reproduced here in Figure 3. For the remainder of the solar spectrum beyond 1.5 μm , in our calculations we assume that values of r and t are constant at their values for $\lambda = 1.5 \mu\text{m}$ ($r = 0.44$; $t = 0.45$). Since the fraction of incident solar energy at $\lambda > 1.5 \mu\text{m}$ is only 8% under clear sky (or 4% under cloudy sky), this assumption has a minimal impact on the broadband reflectance and transmittance. Also shown in Figure 3 is the spectral absorptance a , determined by the requirement $r(\lambda) + t(\lambda) + a(\lambda) = 1$. Similar to the methods for sea ice in Section 2.2, we first determined the weighted visible-band and NIR-band reflectance and transmittance from the spectral values, and then computed the broadband values weighted by cloudy and clear-sky conditions (Table 3). The broadband reflectance for cloudy sky is nearly identical to that for clear sky.

2.4. Seasonal Solar Irradiance at the Top of the Atmosphere (TOA)

Seasonal solar irradiance as a function of latitude and month is given in many publications; we have taken values from Figure 2.6 of Hartmann (2016), and plot them in Figure 4a. We computed a 70–90°N average of the solar irradiance by weighting by latitude, shown in Figure 4c. The solar constant is $1,360.5 \pm 0.5 \text{ W m}^{-2}$ (Kopp and

Table 3

Optical Properties of a 331- μm Layer of Hollow Glass Microspheres for Visible (0.3–0.7 μm), NIR (0.7–2.5 μm), and Broadband (0.3–2.5 μm), Based on Laboratory Measurements of Spectral Reflectance and Transmittance (Figure 1 of Field et al., 2018), Plotted Here in Figure 3

	Visible	NIR (clear)	NIR (cloudy)	Broadband (clear)	Broadband (cloudy)
Reflectance	0.440	0.407	0.409	0.424	0.428
Transmittance	0.469	0.473	0.479	0.471	0.473
Absorptance	0.091	0.120	0.112	0.105	0.099

Note. NIR, Near-Infrared.

Lean (2011), so the relative uncertainty in each of the TOA solar irradiances is very small, just 0.04%.

2.5. Seasonal Solar Irradiance at the Arctic Surface

In summer, the Arctic Ocean is 80% cloud-covered, mostly by thin stratus clouds (Eastman & Warren, 2010). The clouds reflect sunlight, so that the ocean surface below receives only about half of the TOA solar flux. The climatology of solar radiation flux, composed from measurements at drifting ice stations of the Soviet Union and the United States, as well as coastal stations on land, was compiled by the Environmental Working Group (2000), edited by Fetterer and Radionov, and is available from the National Snow and Ice Data Center (<https://doi.org/10.7265/N5MS3QNJ>). Figure 4b gives monthly values of solar irradiance at the ocean surface for latitudes 70, 75,

80, 85, and 90°N, and shows that in summer the surface solar radiation does not vary much with latitude. The values at 75°N are just 5% lower than at 80°N in June–July, and 3% higher in August. The uncertainty in these values, caused mainly by differences in cloud cover and cloud thickness, is $\sim 15\%$, based on differences among three climatologies of the Arctic (Serreze et al., 1998).

3. Methods

We estimate the RF caused by spreading a 331- μm layer of HGMs over representative Arctic sea-ice surfaces using: (a) monthly coverages of the different surface types (Table 1), (b) albedos of the different surface types (Table 2), (c) optical properties of HGMs (Table 3), (d) solar irradiance at the top of the atmosphere (Figure 4a), (e) solar irradiance at the surface (Figure 4b), (f) observed cloud cover over the Arctic Ocean (Arctic Climate Atlas), (g) a two-layer radiation model (Section 3.1), and (h) albedo and transmittance values of Arctic stratus clouds (Section 3.2). Section 3.3 demonstrates how we combine this information for determining the change in albedo caused by spreading HGMs over a surface, using the example of melting thick snow over thick sea ice in June.

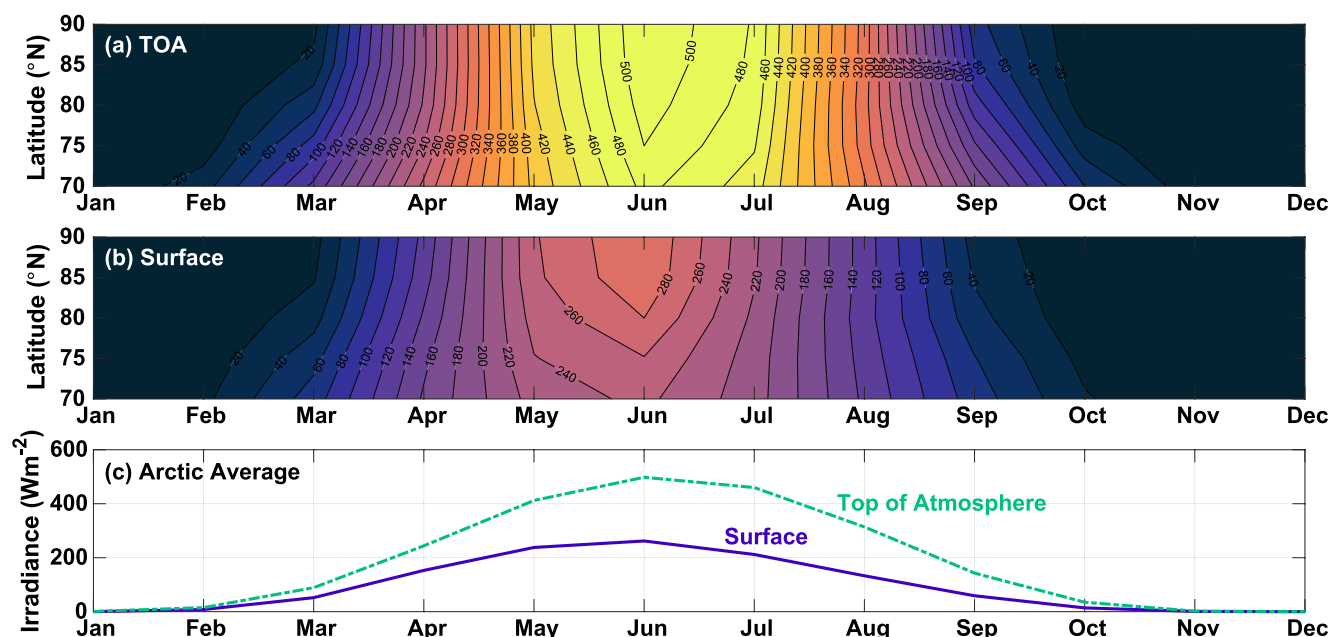


Figure 4. Incident solar radiation in the Arctic (Wm^{-2}) based on Figure 2.6 of Hartmann (2016) for top of the atmosphere, and the Arctic Climate Atlas (Environmental Working Group, 2000) for the surface. (a) Top of atmosphere; (b) Surface; (c) Arctic average.

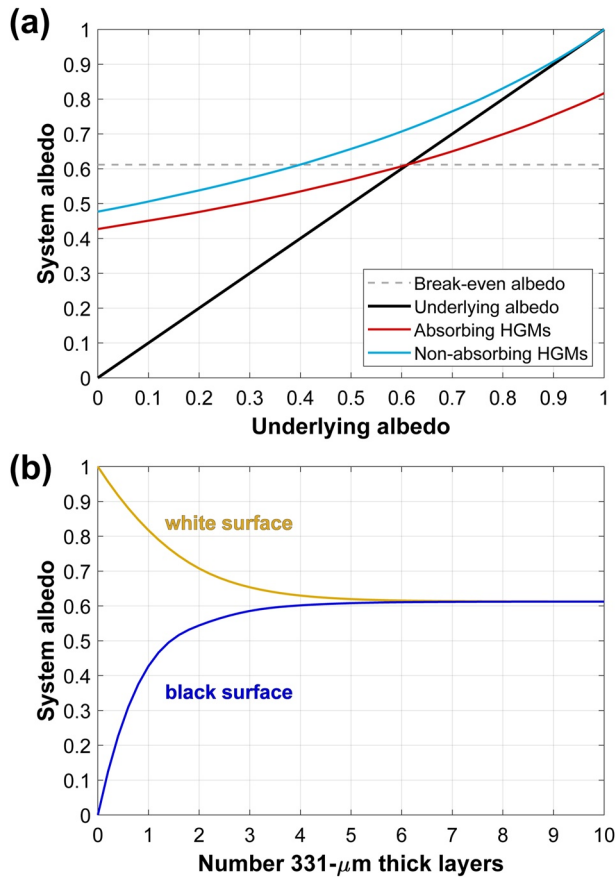


Figure 5. (a) The broadband system albedo resulting from the application of a 331- μm layer of hollow glass microspheres (HGMs), using the two-layer model (Equation 1). Note that for an underlying albedo >0.61 , the absorbing HGMs decrease, rather than increase, the system albedo. The horizontal gray dashed line represents the “break-even” albedo at which no change in the system albedo occurs. (b) Evolution of the system albedo as an increasing number of HGM layers are applied on top of a surface with albedo 1.0 (orange) or 0.0 (blue). Each layer has a thickness of 331 μm . The asymptote occurs at a system albedo of 0.61, consistent with the “break-even” albedo in (a). For this figure we use broadband values of HGM radiative properties from Table 3, taking a weighted average of 20% clear sky and 80% cloudy sky to represent the average climate.

3.1. Computing the Change in Albedo Caused by a Layer of HGMs

We use a simple two-layer model, considering multiple reflections between the 331- μm HGM layer and the opaque layer below (which may be snow, ice, or water). This multiple-reflection model is in common use in college classes on atmospheric physics. A textbook derivation is given by Thomas and Stamnes (1999, Section 6.11). The albedo r of the combined system is

$$r = r_g + \frac{r_i t_g^2}{1 - r_i r_g} \quad (1)$$

where r_i is the albedo of the underlying surface (subscript i for ice); r_g and t_g are the reflectance and transmittance of the HGM layer (subscript g for glass). Equation 1 is equivalent to Equation 6.76 in Thomas and Stamnes (1999).

To demonstrate the effect of HGMs on surface albedo with Equation 1, as an example we use the broadband reflectance (0.4272) and transmittance (0.4726) values for an HGM layer under an average sky (20% clear, 80% cloudy), from Table 3. Figure 5a shows the albedo r of the combined system for various values of underlying albedo r_i . The red curve shows that the application of HGMs will raise the albedo of dark surfaces ($r_i < 0.61$) but will lower the albedo of bright surfaces ($r_i > 0.61$). Figure 5a also shows (as the blue curve) the result of applying a layer of hypothetical non-absorbing HGMs (with reflectance $r_g = 0.48$, transmittance $t_g = 0.52$). In the case of non-absorbing HGMs, there is no “break-even albedo” at which the HGMs will have no effect. Figure 5b shows how the albedo of the system increases with increasing thickness of HGMs, applying up to 10 layers, each of thickness 331 μm , over a black surface ($r_i = 0$). The glass powder gradually hides the underlying surface, reaching an asymptotic albedo of 0.61, consistent with the “break-even” albedo in Figure 5a. Also shown is the case of an underlying white surface ($r_i = 1$), showing the same asymptotic albedo.

3.2. Accounting for Transmittance and Reflectance of Clear Sky and Cloudy Sky

To propagate a change of surface albedo to a change of TOA albedo, we need cloud properties, and we need to consider multiple reflections between the surface and the cloud base, using the form of Equation 1. We assume that the cloud layer in isolation (i.e., if underlain by a black surface) has a broadband

albedo of 0.4 and a transmittance of 0.5. These estimates are based on properties of Arctic stratus clouds reported by Wiscombe (1975), Tsay and Jayaweera (1984), and Curry and Ebert (1992). Using Equation 1 again with these values, a change of surface albedo from 0.25 to 0.50 results in a change of TOA albedo from 0.47 to 0.56. This shows that, because clouds hide the surface, an imposed albedo change at the surface results in a damped change at the TOA.

In 20% of the Arctic sky in summer, the solar beam is not blocked by clouds. A clear sky also absorbs and reflects some sunlight, but less than a cloudy sky does. We take the albedo of a clear atmosphere (over a black surface) as 0.08 for a solar zenith angle of 53° (Table 5.1 of Paltridge & Platt, 1976). The clear-sky atmospheric transmittance for sub-Arctic summer is 0.76 (Warren and Wiscombe, unpublished; quoted by Charlson et al., 1991). Using Equation 1 again, a change of surface albedo from 0.25 to 0.50 results in a change of the TOA albedo from 0.23 to 0.38.

Table 4

An Example of the Calculations Using Equations 1–7 to Determine the Change in Albedo and Radiative Forcing Caused by the Application of a Hollow Glass Microspheres Layer, for the Case of Thick Melting Snow on Sea Ice, in the Month of June

Surface type: thick melting snow on sea ice in June	Value	Uncertainty
Sea ice area (10^6 km^2), average for the years 2010–2020	7.8	0.1
Fractional coverage of melting thick snow on sea ice within the total sea-ice area for this month	50%	14%
Area of melting thick snow on sea ice A_M (10^6 km^2)	3.9	1.1
Incident solar flux at the TOA, weighted average over 70–90°N (F_T , Wm^{-2})	498	0.2
Incident solar flux at the surface, weighted average over 70–90°N (F_S , Wm^{-2})	262	39
Apparent atmospheric transmittance (F_S/F_T)	52.6%	7.9%
Change of albedo caused by a cover of HGMs, Δr_s	−0.098	0.007
(a) Local radiative forcing at surface (ΔF_{SL} , Wm^{-2}) = $-\Delta r_s F_S$	25.8	4.3
Fraction of Arctic Ocean covered by melting thick snow (A_M/A_o), where $A_o = 11.1 \times 10^6 \text{ km}^2$	35%	10%
(b) Surface radiative forcing for the Arctic Ocean, (ΔF_{SA} , Wm^{-2}) = $-(A_M/A_o)\Delta r_s F_S$	9.0	3.0
(c) Albedo change at TOA over melting thick snow, Δr_T	−0.053	0.005
Albedo change at TOA averaged over the Arctic Ocean ($\Delta r_T A_M/A_o$)	−0.018	0.005
(d) TOA radiative forcing over the Arctic Ocean, (ΔF_{TA} , Wm^{-2}) = $-(\Delta r_T A_M/A_o)F_T$	9.17	2.73
<i>Annual average TOA RF over the Arctic Ocean</i>	<i>1.48*</i>	<i>0.43*</i>
(e) <i>TOA annual radiative forcing for the entire Earth, due to HGM application on melting thick snow on Arctic sea ice</i>	<i>0.03*</i>	<i>0.01*</i>

Note. In the bottom two rows, the italicized descriptions and values with asterisks pertain to annual averages, which include calculations for both months when thick melting snow is present (May and June). The bold letters in parentheses correspond to the steps described in the text (Section 3.3).

3.3. Computing the Radiative Forcing

The calculation for determining the RF caused by spreading HGMs onto each surface type uses the equations below. Here we give an example using melting thick snow on thick sea ice in June, a surface type that covers a large area of the Arctic Ocean in late spring. Table 4 shows the results using the following steps:

(a) The local RF at the surface:

$$\Delta F_{SL} = -\Delta r_s F_S \text{ (Wm}^{-2}, \text{ monthly)} \quad (2)$$

where F_S is the downward solar flux (irradiance) at the surface; Δr_s is the change in the surface albedo r_s .

(b) The surface RF for the Arctic Ocean:

$$\Delta F_{SA} = -(A_m/A_o) \Delta r_s F_S \text{ (Wm}^{-2}, \text{ monthly)}, \quad (3)$$

where A_m is the area of a given surface type; A_o is the ice-ocean area over 70–90°N, $11.1 \times 10^6 \text{ km}^2$. The total area of a surface type, A_m , is obtained as the product of fractional coverage (Table 1) and the sea-ice area, averaged for each month over 2010–2020 using PM sea-ice concentration retrievals (Meier et al., 2021) and field observations (Fetterer & Untersteiner, 1998; Nazintsev, 1964; Perovich et al., 2002; Webster et al., 2015, 2022).

(c) Albedo at the TOA (r_T), taking into account multiple reflections between the surface and cloud base:

$$r_T \text{ (cloudy)} = r_c + (r_s t_c^2) / (1 - r_c r_s). \quad (4)$$

Here, we again use the form of Equation 1, but where $r_c = 0.4$ is the cloud albedo and $t_c = 0.5$ is the cloud transmittance.

For clear sky, Equation 4 is altered as follows: We replace r_c by the atmospheric albedo, $r_a = 0.08$, and replace t_c by the atmospheric transmittance, $t_a = 0.76$. For the cloudy fraction 0.8 and clear-sky fraction 0.2, the weighted average change in TOA albedo (Δr_T) is:

$$\Delta r_T = 0.8 \times \Delta r_T \text{ (cloudy)} + 0.2 \times \Delta r_T \text{ (clear)}, \quad (5)$$

(d) The TOA RF over the Arctic Ocean (analogous to Equation 3):

$$\Delta F_{TA} = -(A_m/A_o) \Delta r_T F_T \text{ (Wm}^{-2}, \text{ monthly)}, \quad (6)$$

where F_T is the incident solar flux at the TOA.

(e) The annual TOA RF over the entire Earth:

$$\Delta F_{TE} = \frac{A_o}{A_E} \frac{1}{12} \sum_{m=1}^{12} \Delta F_{TA,m} \quad (7)$$

where the sum is over 12 months; A_E is the area of the Earth, $512 \times 10^6 \text{ km}^2$.

(f) The results of Tables 1–3 and Equations 1–7 for individual surface types are added to obtain the monthly and annual forcing for all types combined, and are presented in the next section.

4. Results

Figure 6 and Table 5 show the monthly RF for each surface type caused by spreading HGMs onto its surface. Table 5 also gives the annual averages. In this idealized calculation, the HGMs are assumed to be renewed every month to maintain a 331- μm thick layer, so that they are not buried by new snowfall. Figure 6 shows that the application of a 331- μm layer of HGMs causes a positive RF (i.e., warming) over three surface types: (a) sea ice with a cold, thick snow cover (prominent in March–May); (b) sea ice with a melting, thick snow cover (prominent in May–June); and (c) sea ice with a cold, thin snow cover (in September–October). The annual average surface forcings for these three types are +2.9, +1.5, and +0.3 Wm^{-2} , respectively (Table 5). The TOA forcings are similar. The effect of this positive forcing is greatest from early spring to early summer and causes a net annual positive RF at both TOA and surface. Thus, the monthly application of a 331- μm layer of HGMs on Arctic sea ice in these spring and summer months would increase solar absorption and accelerate the melting of Arctic sea ice.

For melting sea ice with no snow cover, and for refrozen melt ponds, spreading HGMs onto their surfaces causes a negligible RF. In the case of snow-free melting sea ice, which is the dominant ice type in summer, this is due to an insufficient increase in albedo with the application of HGMs (the albedo of this surface type is close to the “break-even” albedo in Figure 4). For refrozen melt ponds, their areal coverage is small, and irradiance in autumn, when ponds freeze over, is also small. Both factors contribute to the negligible change in surface and TOA RF over refrozen ponds.

HGMs do cause a negative RF at the surface for a handful of surface types: nilas with no snow cover (-0.9 Wm^{-2}), nilas with a cold, thin snow cover (-0.5 Wm^{-2}), and open melt ponds (-1.1 Wm^{-2}). However, in reality the melt ponds will not remain covered by HGMs due to wind, so the effect of HGMs on the RF over open melt ponds will be negligible. Winds will blow the HGMs to the pond edges where they will clump, as happened on the pond in Minnesota after the ice melted (Figure 5 of Field et al., 2018; Leslie Field, personal communication, 2021) and similarly in situations where plastic spheres were placed in wastewater ponds containing tailings and oil (Ramirez, 2010). Moreover, melt ponds have lateral drainage channels that connect to leads and open water (Webster et al., 2022), in which the glass bubbles would be transported out of ponds.

We therefore compute the *net* RF of HGMs for all surface types combined under the assumption that HGMs deposited in melt ponds do not change the melt-pond albedo. Figure 7 shows this net monthly RF (blue bars). The RF is positive from March to June, and very small in other months; the annual average for the Arctic Ocean area is +3.3 Wm^{-2} at the surface and +3.5 Wm^{-2} at TOA.

The surface for which HGMs cause the largest negative forcing is young bare ice (nilas, no snow), with forcings as large as -4 Wm^{-2} in April and May (Figures 6b, Table 5). But in these months, that negative forcing is more than offset by the positive forcing of HGMs on sea ice with a thick snow cover. In fact, the application of HGMs on Arctic sea ice is most detrimental in April–June due to HGMs darkening the surface. In July–September, the application of HGMs has minimal effect on the RF. However, there is an important caveat to consider in the interpretation of this result: snow, which is more reflective than HGMs, is present on some of the sea ice in July and August (Radionov et al., 1997; Warren et al., 1999; Webster et al., 2022), and our results do not account for

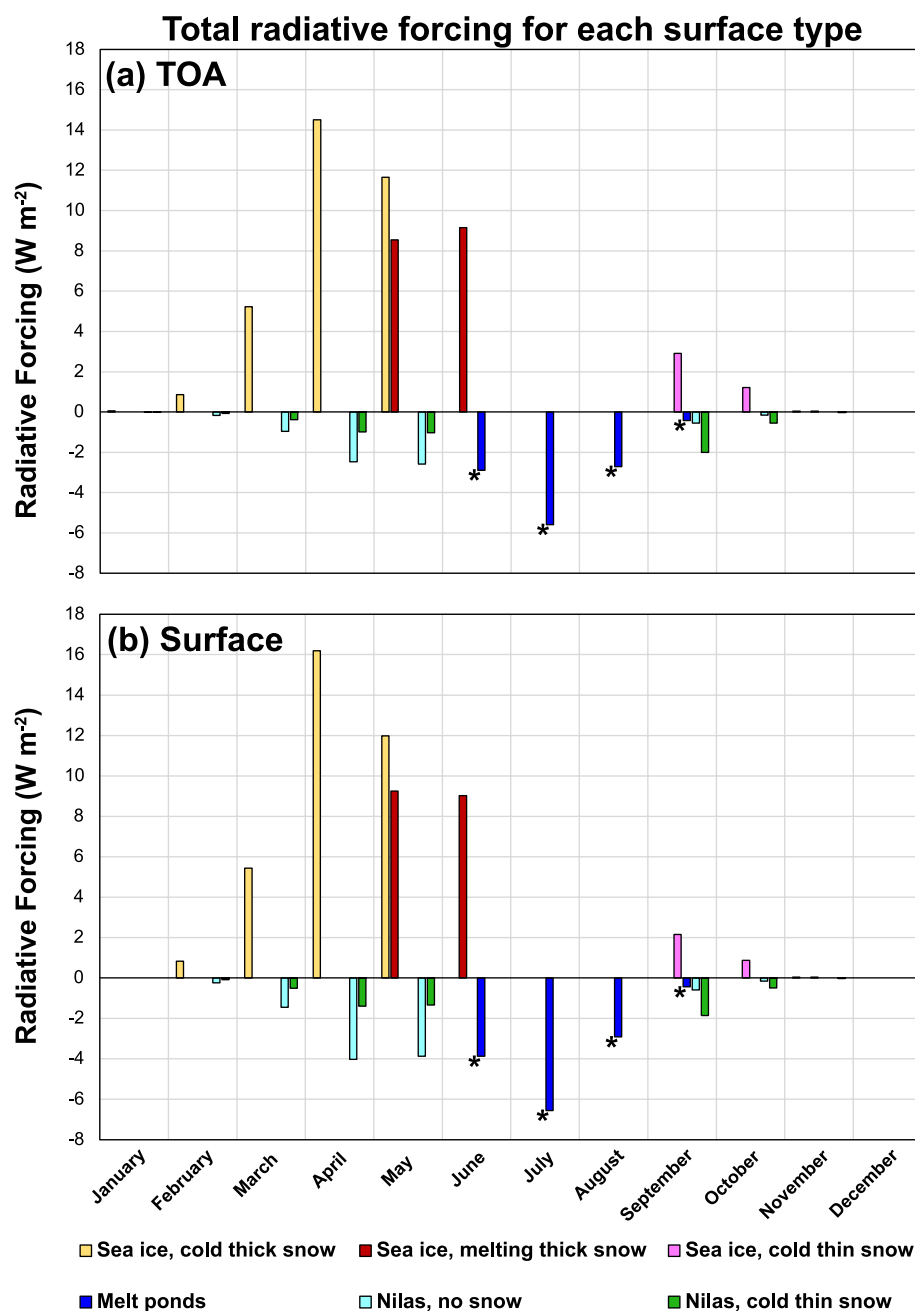


Figure 6. The monthly radiative forcing (RF) over 70–90°N at (a) top of the atmosphere and (b) surface, due to the application of a 331- μm layer of hollow glass microspheres (HGMs) over different surface types of Arctic sea ice. We show only the surface types that experience RF greater than 1 W m^{-2} in at least one month. Numerical values, as well as the annual averages, are given in Table 5 for all eight surface types. Asterisks indicate melt ponds that remain uniformly covered by HGMs. In reality HGMs would not maintain uniform coverage on this surface type due to wind, and accordingly, would cause negligible RF.

summertime snow. If snow persists into these summer months, HGMs would have an even larger detrimental effect on Arctic sea ice.

In short, the application of HGMs on Arctic sea ice in March, April, May, and June would cause a strong positive RF, averaging about $+10 \text{ W m}^{-2}$ during these months, both at the surface and TOA. Application of HGMs during the rest of the year (summer, autumn, winter) would have almost no effect, as shown in Figure 7. The annual average RF over the Arctic at TOA would be $+3.5 \text{ W m}^{-2}$, leading to an accelerated loss of sea ice. The global RF

Table 5

The Monthly and Annual Average Radiative Forcing at the Top of the Atmosphere and Surface Due To the Application of a 331- μm Layer of Hollow Glass Microspheres Over Different Surface Types of Arctic Sea Ice

TOA radiative forcing over the Arctic ocean (Wm^{-2})	Sea ice, cold thick snow	Nilas, no snow	Sea ice, melting thick snow	Melting sea ice, no snow	Melt ponds	Sea ice, cold thin snow	Refrozen ponds	Nilas, cold thin snow	Monthly total
January	0.06 0.01	−0.01 0.00	-	-	-	-	-	−0.01 0.00	0.04 0.01
February	0.87 0.08	−0.17 0.04	-	-	-	-	-	−0.07 0.02	0.63 0.09
March	5.23 0.48	−0.96 0.23	-	-	-	-	-	−0.38 0.09	3.89 0.54
April	14.51 1.34	−2.47 0.63	-	-	-	-	-	−0.99 0.25	11.04 1.5
May	11.66 5.99	−2.58 1.01	8.55 4.39	-	-	-	-	−1.03 0.41	16.60 7.51
June	-	-	9.16 2.73	0.09 0.11	−2.89* 2.77	-	-	-	9.25 2.73
July	-	-	-	0.12 0.14	−5.59* 2.60	-	-	-	0.12 0.14
August	-	-	-	0.06 0.07	−2.71* 1.41	-	-	-	0.06 0.07
September	-	−0.55 0.18	-	-	−0.41* 0.32	2.92 0.28	−0.37 0.28	−2.00 0.20	−0.00 0.48
October	-	−0.15 0.04	-	-	-	1.22 0.11	-	−0.55 0.06	0.52 0.13
November	0.05 0.00	-	-	-	-	0.05 0.00	-	−0.03 0.00	0.07 0.01
December	-	-	-	-	-	-	-	-	-
Annual Average	2.70 0.51	−0.57 0.10	1.48 0.43	0.02 0.02	−0.97* 0.34	0.35 0.02	−0.03 0.02	−0.42 0.04	3.53 0.68
Surface radiative forcing over the Arctic Ocean (Wm^{-2})	Sea ice, cold thick snow	Nilas, no snow	Sea ice, melting thick snow	Melting sea ice, no snow	Melt ponds	Sea ice, cold thin snow	Refrozen ponds	Nilas, cold thin snow	Monthly Total
January	-	-	-	-	-	-	-	-	-
February	0.83 0.14	−0.24 0.06	-	-	-	-	-	−0.08 0.02	0.51 0.16
March	5.43 0.93	−1.45 0.40	-	-	-	-	-	−0.50 0.14	3.48 1.02
April	16.18 2.76	−4.02 1.16	-	-	-	-	-	−1.39 0.40	10.77 3.02
May	11.98 6.39	−3.87 1.61	9.26 4.93	-	-	-	-	−1.34 0.56	16.03 8.26

Table 5
Continued

TOA radiative forcing over the Arctic ocean (Wm^{-2})	Sea ice, cold thick snow	Nilas, no snow	Sea ice, melting thick snow	Melting sea ice, no snow	Melt ponds	Sea ice, cold thin snow	Refrozen ponds	Nilas, cold thin snow	Monthly total
June	-	-	9.02 <i>2.99</i>	0.25 <i>0.13</i>	-3.87* <i>3.74</i>	-	-	-	9.27 <i>2.99</i>
July	-	-	-	0.29 <i>0.12</i>	-6.5* <i>3.18</i>	-	-	-	0.29 <i>0.12</i>
August	-	-	-	0.13 <i>0.05</i>	-2.92* <i>1.58</i>	-	-	-	0.13 <i>0.05</i>
September	-	-0.59 <i>0.21</i>	-	-	-0.44* <i>0.34</i>	2.16 <i>0.37</i>	-0.38 <i>0.30</i>	-1.85 <i>0.32</i>	-0.67 <i>0.61</i>
October	-	-0.16 <i>0.05</i>	-	-	-	0.88 <i>0.15</i>	-	-0.50 <i>0.09</i>	0.22 <i>0.18</i>
November	0.05 <i>0.01</i>	-	-	-	-	0.04 <i>0.01</i>	-	-0.03 <i>0.01</i>	0.06 <i>0.01</i>
December	-	-	-	-	-	-	-	-	-
Annual Average	2.87 <i>0.59</i>	-0.86 <i>0.17</i>	1.52 <i>0.48</i>	0.06 <i>0.02</i>	-1.15* <i>0.43</i>	0.26 <i>0.03</i>	-0.03 <i>0.02</i>	-0.47 <i>0.06</i>	3.35 <i>0.78</i>

Note. The bold numbers are the radiative forcing (RF); the lower numbers in italics are the uncertainties. Asterisks indicate melt ponds that remain uniformly covered by hollow glass microspheres (HGMs). In reality, melt ponds would not remain uniformly covered by HGMs. Therefore, the RF over melt ponds is not included in the monthly total and annual average of all surfaces combined.

would be $+0.076 \text{ Wm}^{-2}$. For context, this is about one-tenth of the global RF that would result from the removal of all ice year-round, $+0.7 \text{ Wm}^{-2}$ (Hudson, 2011).

Our calculations are for the present distribution of surface types on the Arctic Ocean. If the HGMs were to be applied in the future, with less sea-ice coverage as a consequence of global warming, the RF caused by HGMs would be smaller. To a first approximation, as the ice margin shrinks, the area covered by each of the sea-ice types would be smaller in all seasons. Where ice does exist, the forcing by HGMs would be similar to what we have computed, but the average forcing for the Arctic Ocean would be correspondingly smaller.

5. A Non-Absorbing Glass?

The laboratory experiment of Field et al. (2018) showed that the HGMs they are using have significant absorption, as shown in Figure 3. As we have shown, spreading those HGMs over the Arctic Ocean would actually warm the climate, the opposite of what is desired. If HGMs are to be spread over Arctic sea ice to raise its albedo, it will be necessary to manufacture non-absorbing HGMs, avoiding any contamination during the manufacturing process (such as black carbon from graphite lubrication on machinery), and to avoid contamination during transport and dispersal (particularly because the wind that carries the HGMs away from the ship will also carry the plume of the ship's engine-exhaust in the same direction, which would contain black carbon). Tiny amounts of absorptive impurities are effective at lowering the albedo of high-albedo surfaces. For example, just 20 parts per billion of black carbon will reduce the albedo of snow by $\sim 1\%$ (Figure 5a of Dang et al., 2015).

Now we assume that all these difficulties can be overcome, and also that the glass spheres will not experience breakage with shear (as they are susceptible to, according to the manufacturer), and we compute the RF by HGMs that are completely non-absorptive. We partition the HGM broadband absorption of 0.10 (Table 3) between the reflectance and transmittance, increasing each by 0.05 to yield an adjusted broadband reflectance of 0.48 and transmittance of 0.52. The resulting RF is shown in yellow in Figure 7. With non-absorbing HGMs, there is substantial negative forcing. Maintaining a $331\text{-}\mu\text{m}$ thick layer of non-absorbing HGMs every month results in an annual average RF of -4.6 Wm^{-2} at the surface and -4.4 Wm^{-2} at the TOA.

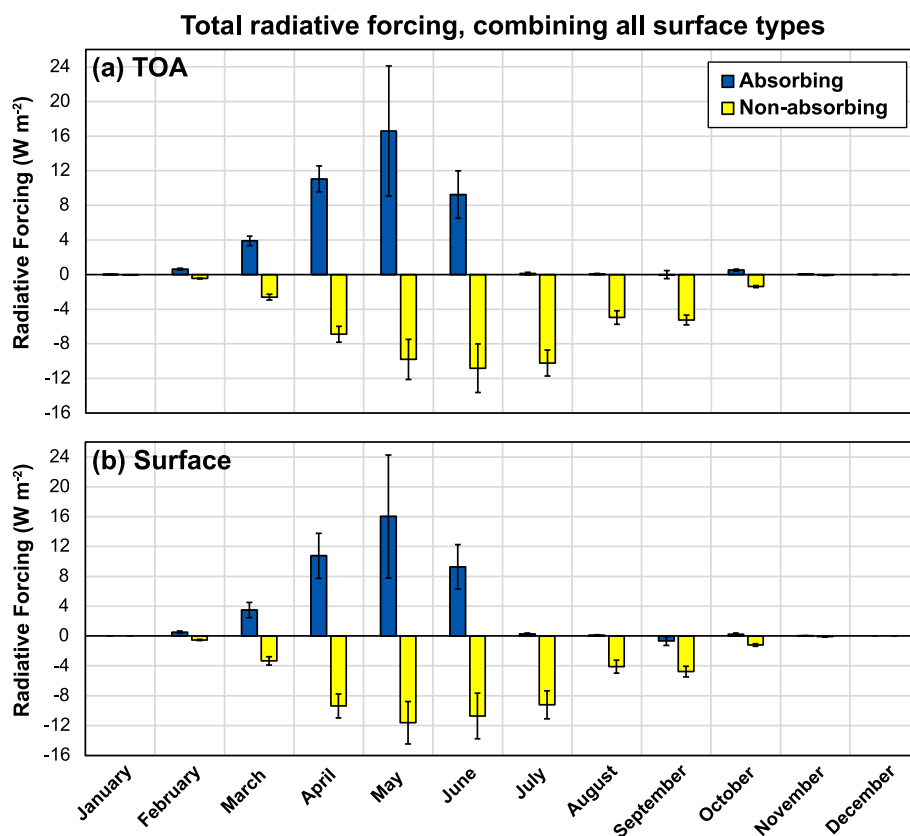


Figure 7. The monthly total radiative forcing (sum over all surface types) over 70–90°N at (a) top of the atmosphere and (b) surface, due to the application of a 331- μm layer of absorbing hollow glass microspheres (HGMs) (blue), as in Field et al. (2018), or hypothetical non-absorbing HGMs (yellow), as described in Section 5. In both cases we assume melt ponds are unaffected because they cannot remain covered by HGMs. The error bars are the uncertainties.

Assuming the non-absorbing HGMs would be applied only once a year, at a time when they would not be immediately covered by new snowfall, they should be applied during May, and will then be effective for the four months May–August. The average RF for these four months is -8.9 W m^{-2} at the surface and -9.0 W m^{-2} at TOA. Assuming only these four months would be affected, the annual RF would be -3 W m^{-2} . The Arctic Ocean area is 2.2% of the Earth's area, so the *global annual* RF at TOA from spreading non-absorbing HGMs on Arctic sea ice in May would be -0.06 W m^{-2} .

Next we compute the mass of HGMs to be applied in May, so as to obtain the RF “efficiency,” that is, the RF per unit mass. The bulk density of HGMs is 0.125 g cm^{-3} , so a 331- μm layer will weigh 41 g m^{-2} . Application of a 331- μm layer in May to the entire sea-ice area, $8.8 \times 10^6 \text{ km}^2$ in May, requires $3.63 \times 10^{11} \text{ kg}$ total. The annual average TOA RF efficiency for the Arctic Ocean for non-absorbing HGMs is then $-3.0 \text{ W m}^{-2} / 3.63 \times 10^{11} \text{ kg} = -8.26 \text{ W m}^{-2} \text{ GT}^{-1}$. The *global annual* TOA RF efficiency for non-absorbing HGMs would be $-0.18 \text{ W m}^{-2} \text{ GT}^{-1}$.

We might compare this last value to the global shortwave efficiency for stratospheric sulfur injection, which is between -0.2 and -1 W m^{-2} per teragram of sulfur injected per year (Laasko et al., 2022), that is, a factor of 1,000–4,000 larger than the HGM efficiency. Admittedly, this is not an apples-to-apples comparison, nor even an apples-to-oranges comparison, but rather a glass-to-sulfur comparison! It is also overly simplified, as it does not consider the cost of deploying, nor the possible harm to Arctic ecosystems, of either method.

6. Conclusions

The problem with HGMs is that they exhibit significant absorption at solar wavelengths; spreading them across the Arctic Ocean could brighten surfaces of albedo < 0.61 , but they would darken the many Arctic surfaces with

albedo >0.61 . Considering the radiative properties of the eight major surface types on the Arctic Ocean, and their seasonal cycles, as well as the radiative properties of HGMs, we conclude that spreading this glass powder would cause a warming of the Arctic Ocean, by an annual average of $3.3\text{--}3.5\text{ Wm}^{-2}$, and would thereby accelerate sea-ice melt. The RF for the Arctic Ocean area would peak in May with $+16.0\text{ Wm}^{-2}$ at the surface and $+16.6\text{ Wm}^{-2}$ at TOA. Although clouds substantially hide the HGM-covered surface, the incident solar flux at TOA is about twice that at the surface, so the TOA forcing does not differ greatly from the surface forcing. September is the only month in which HGMs cause a net negative RF at the Arctic surface (-0.665 Wm^{-2}) and TOA (-0.004 Wm^{-2}). However, the decrease caused by HGMs in September is more than offset by changes in other months. In particular, the HGMs have the most detrimental effect on surfaces with thick snow because HGMs darken high-albedo surfaces.

If a method could be devised to manufacture completely *non-absorbing* HGMs, and if they could be transported and distributed over the Arctic without contamination, they would be able to cool the climate. If HGMs are to be distributed once a year, the maximum benefit would be achieved by distributing them in early May, resulting in an annual RF at TOA of -3.0 Wm^{-2} for the Arctic Ocean. This compares to the positive RF of $+21\text{ Wm}^{-2}$ that would result from loss of all Arctic sea ice during the sunlit season, relative to a 1971 baseline (Pistone et al., 2019). The annual global TOA RF from non-absorbing HGMs would be -0.06 Wm^{-2} , or -0.18 Wm^{-2} per GT of HGMs applied.

Data Availability Statement

Stephen Hudson provided digital values for the spectral albedo of cold deep snow, available in the Supporting Information of <https://agupubs.onlinelibrary.wiley.com/doi/full/10.1029/2006JD007290>. Bonnie Light and Maddie Smith provided spectral albedos from impacts of climate on the eco-systems and chemistry of the Arctic pacific environment (<https://seabass.gsfc.nasa.gov/cruise/icescape2010>, <https://seabass.gsfc.nasa.gov/cruise/icescape2011>) and multidisciplinary drifting observatory for the study of Arctic climate (<https://arcticdata.io/catalog/view/doi%3A10.18739%2FA2FT8DK8Z>).

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