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Key Points:

- Albedo of cold laboratory-grown sea ice was observed to be significantly enhanced by the precipitation of a salt crust
- The salt crust is brighter than snow, and a positive salt-albedo feedback could promote sea-ice advance through enhanced albedo
- Salt-albedo feedback could also promote ice retreat through the formation of saline surface melt puddles at temperature as low as −23°C

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The spectral albedo of sea ice and salt crusts on the tropical ocean of Snowball Earth: 1. Laboratory measurements

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Abstract The ice-albedo feedback mechanism likely contributed to global glaciation during the Snowball Earth events of the Neoproterozoic era (1 Ga to 544 Ma). This feedback results from the albedo contrast between sea ice and open ocean. Little is known about the optical properties of some of the possible surface types that may have been present, including sea ice that is both snow-free and cold enough for salts to precipitate within brine inclusions. A proxy surface for such ice was grown in a freezer laboratory using the single salt NaCl and kept below the eutectic temperature (−21.2°C) of the NaCl-H_2O binary system. The resulting ice cover was composed of ice and precipitated hydrohalite crystals ($NaCl \cdot 2H_2O$). As the cold ice sublimated, a thin lag-deposit of salt formed on the surface. To hasten its growth in the laboratory, the deposit was augmented by addition of a salt-enriched surface crust. Measurements of the spectral albedo of this surface were carried out over 90 days as the hydrohalite crust thickened due to sublimation of ice, and subsequently over several hours as the crust warmed and dissolved, finally resulting in a surface with puddled liquid brine. The all-wave solar albedo of the subeutectic crust is 0.93 (in contrast to 0.83 for fresh snow and 0.67 for melting bare sea ice). Incorporation of these processes into a climate model of Snowball Earth will result in a positive salt-albedo feedback operating between −21°C and −36°C.

1. Introduction

The albedo of earth's polar oceans exhibits a sharp transition between its liquid phase, the open ocean, where its broadband albedo is generally less than 0.1 [*Pegau and Paulson*, 2001; *Brandt et al.*, 2005], and its frozen phase (sea ice, with or without snow) where the albedo is typically larger than 0.4, and often close to 0.8 [e.g., *Wiscombe and Warren*, 1980; *Perovich et al.*, 2002; *Brandt et al.*, 2005; *Zatko and Warren*, 2015]. This strong contrast in the amount of light backscattered to the atmosphere, and the resulting heat absorption at the surface, is the basis for the well-known positive ice-albedo feedback whereby large changes in sea ice coverage are driven by relatively small changes in temperature. This ice-albedo feedback operates at temperatures close to 0°C. Such sensitivities involving surface albedo appear capable of amplifying changes in global climate [see e.g., *Serreze and Barry*, 2011] and they could be particularly influential if operating on the large expanses of Earth's surface at low latitude where the top-of-atmosphere solar insolation is large.

The global glaciations of the Neoproterozoic era (1 Ga to 544 Ma) are described by various aspects of the Snowball Earth hypothesis. Whether glaciation extended all the way to the equatorial regions, as in a "hard snowball," or whether some tropical regions remained ice free on an otherwise glaciated planet, (a "waterbelt" [*Pierrehumbert et al.*, 2011]) is not certain, even though the geologic evidence for major Neoproterozoic glacial activity is well recognized [see *Hoffman et al.*, 1998; *Hoffman and Schrag*, 2002; *Fairchild and Kennedy*, 2007]. Regardless of the exact details of Earth's surface condition, ice-albedo feedback was likely key to the inception of low-latitude glaciation. However, much of what we know about the albedo of the frozen sea surface results from observations of modern Earth. Such sea ice is typically snow covered in the autumn, winter, and spring, and, as a result, rarely experiences temperatures much lower than −20°C for extended periods of time [*Perovich and Elder*, 2001]. Model simulations of Snowball Earth suggest that, at low latitudes, snow cover may have been sparse, with large areas where evaporation exceeded precipitation [*Pollard and Kasting*, 2004; *Abbot and Pierrehumbert*, 2010; *Le Hir et al.*, 2010; *Pierrehumbert et al.*, 2011]. Without a pervasive snow cover, the albedo of the tropical oceans would have been determined by the optical properties of bare sea ice with temperatures consistently lower than −23°C [*Pollard and Kasting*, 2004].

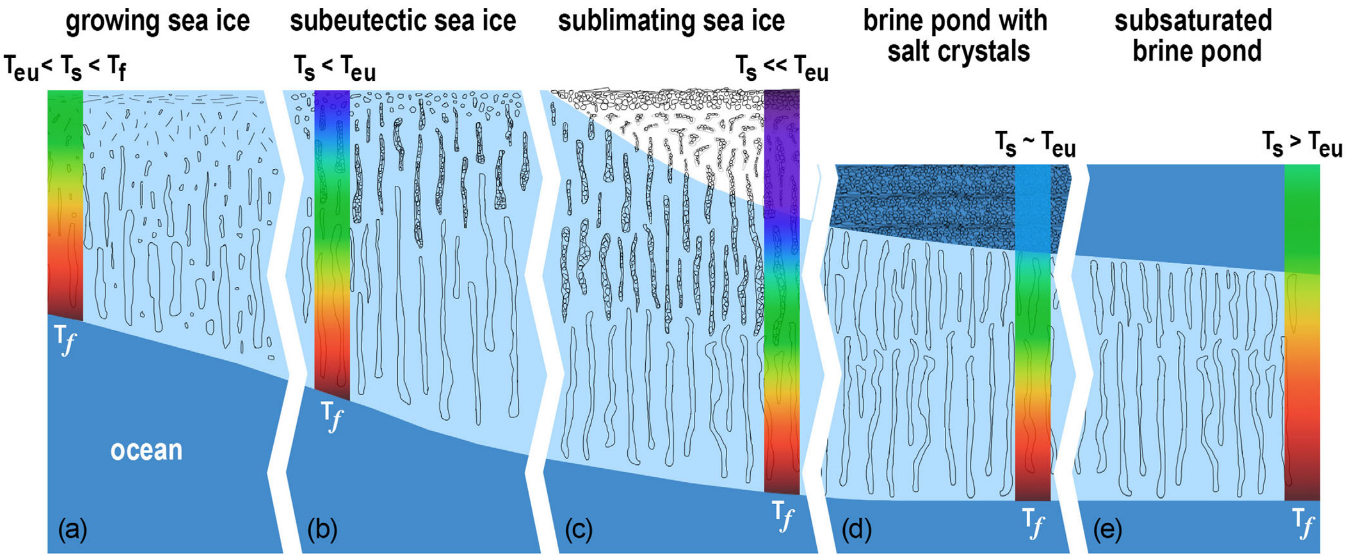

Figure 1. Schematic of hypothesized lifecycle of a Snowball Earth ocean surface at low latitudes. Sublimation in excess of precipitation would keep the surface snow free. (a) Cross section shows growing ice. When the ice becomes thick enough, (b) the surface temperature drops and the ice becomes subeutectic. At this point, crystals of saturated salt precipitate within the brine inclusions. (c) The onset of ice sublimation then leaves a lag deposit of salt crystals. (d) Increase in temperature to near the eutectic temperature would cause the salt crystals to begin to melt the surface ice, resulting in a brine pond saturated with salt. (e) Further warming would result in a subsaturated brine pond.

The optical properties of snow-free sea ice have been measured during the autumn growing phase [*Warren et al*., 2002; *Brandt et al*., 2005] and during the summer melt phase [*Perovich et al*., 2002]. These conditions, however, may not be representative of sea ice on Snowball Earth. Since it is rarely observed on modern Earth, little is known about the optical properties of bare, cold sea ice, and a quantitative understanding appropriate for building a comprehensive albedo parameterization has yet to be established.

Cold sea ice would have unique properties simply because of its included salt. As sea ice grows, liquid brine is entrained within the ice at the ice-ocean growth interface [see *Weeks and Ackley*, 1986], but the salts remain dissolved only so long as they are not saturated. If the ice is cold enough, the salt within brine inclusions becomes saturated and precipitates in crystal form. The inherent optical properties of sea ice are known to depend strongly on the presence of these crystals [*Perovich and Grenfell*, 1981; *Light et al*., 2004]. The most abundant salt to precipitate is sodium chloride dihydrate, or "hydrohalite" ($NaCl \cdot 2H_2O$). In brine with the same chemistry as the modern oceans, initial hydrohalite precipitation begins at −22.9°C and this precipitation appears not to be complete until about −36.2°C [*Marion et al*., 1999].

If the tropical regions of Snowball Earth favored evaporation over precipitation, then significant ice surface ablation would have been likely at these latitudes. Sublimation of cold saline ice could have led, over time, to ice surfaces covered with accumulations of precipitated salt crystals. Such a cryogenite deposit is a new surface type whose chemistry, physics, and feedbacks have not been extensively studied. *Light et al*. [2009] proposed that this surface type would have albedo significantly higher than that of bare sea ice. Such an enhanced albedo would have implications for Earth's surface energy budget during periods of uniformly low surface temperature. The demise of a salt crust under warming conditions may also influence the surface energy budget of a warming Snowball Earth.

Figure 1 shows five phases in the hypothesized lifecycle of a low-temperature sea-ice cover on Snowball Earth. Initially, the ice grows in the same way it grows on modern Earth. As the surface cools below its eutectic temperature (T_{eu}), dissolved salts within the brine inclusions begin to precipitate. Once the ice surface begins to sublimate, precipitated salt crystals form a lag deposit at the surface. If the surface temperature remains below T_{eu} then the lag deposit continues to deepen. If the surface temperature increases above T_{eu} then ice in contact with the salts will begin to melt, promoting the formation of salt-saturated puddles. Further warming above T_{eu} could result in subsaturated brine ponds on the ice surface.

While numerous studies have been carried out in situ to measure the albedo of various saline ice surfaces [e.g., *Grenfell and Maykut*, 1977; *Perovich et al*., 2002; *Brandt et al*., 2005], bare, cold, sublimating marine ice is

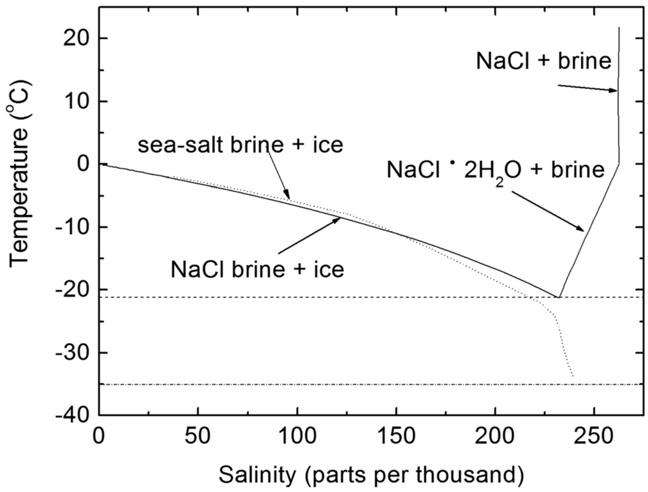

Figure 2. Freezing equilibrium relationships in the NaCl-H2O binary system compared with full seawater system. For the binary system, the eutectic temperature is −21.1°C; for the full seawater system, it is about −36°C.

not commonly found in nature. However, a surrogate surface can be grown and maintained in the laboratory. We stress here that there are limitations to this surrogate. The very long time periods associated with these processes are simply not possible in laboratory simulations and it is difficult to grow laboratory ice with some of the characteristics of natural ice. It was thus necessary to induce conditions we deemed representative; the "shortcuts" we employed are described as the methods are presented below.

In this paper, we report measurements of the evolution of spectral albedo of laboratory-grown bare, cold saline ice with a hydrohalite lag deposit, as a surrogate for similar ice on Snowball Earth. This is a surface type we expect to be widespread only in the initial phase of a snowball event. Later, flowing sea-glaciers [*Goodman and Pierrehumbert*, 2003] would displace the sea ice and thereafter become the dominant ice type. Sea-glacier albedo was studied by *Dadic et al*. [2013]. The sea ice and salt crust we are studying in this paper, although not widespread, could have been crucial in the first catastrophic freezing of the tropics because of its extremely high albedo and could also have contributed to the ultimate demise of the ice cover through a lowering of surface albedo due to the warming, dissolving, and ponding of salt crusts. In a companion paper [*Carns et al*., 2016], we attempt to explain the measured spectral albedo quantitatively using radiative transfer modeling.

2. Experimental Design

A prototype for this plausible Snowball Earth surface was previously created in our laboratory [*Light et al*., 2009]. That sample was grown in an insulated bucket and its horizontal extent (0.28 m) was inadequate for making representative optical measurements. The sample grown for this study was much larger, and like the prototype, was restricted to the binary NaCl-H_2O system. By using a single-salt system, it was possible to create subeutectic saline ice at temperatures achievable in our laboratory ($T \gtrsim -30$°C). The interpretation of the inherent optical properties of the ice was simplified by only having a single cryogenite species, but also by eliminating the presence of liquid brine in equilibrium with the salt crystals [see *Light et al*., 2004]. Figure 2 shows the threshold temperatures and concentrations for the freezing equilibrium relationships of the binary system in comparison with the full seawater system. The eutectic temperature (T_{eu}) of the binary system is −21.2°C. Below this temperature, only ice and hydrohalite exist; there is no liquid brine. A lag deposit was expected to form only at temperatures below T_{eu}. In isolation, hydrohalite melts at 0.11°C [*Marion and Grant*, 1994]. Hydrohalite in contact with ice, however, would be expected to form liquid brine at temperatures above −21.2°C, thus allowing for the possibility of salt-saturated liquid melt-pond formation at temperatures well below 0°C.

For this study, the ice sample was grown in an insulated tank of capacity 988 L, within a walk-in freezer laboratory at the Applied Physics Laboratory, University of Washington. The horizontal dimensions of the tank were 1.22 m × 1.12 m (height 1.24 m; http://www.bonarplastics.com/Products/Polar-Insulated-Containers/Bag-in-the-Box-Insulated-Liquid-Tote). The walls and floor were insulated with 5 cm thick polyurethane foam. A fitted plastic liner (Grayling Industries) was used to improve leak resistance and allow for easier cleanup. Surface and internal temperatures were recorded throughout the experiment using a thermistor string with 5 cm vertical resolution mounted in one corner of the tank prior to freezing.

The tank was filled with filtered tap water and enough reagent-grade NaCl to bring the salinity to 24 parts per thousand (‰). Typically, natural sea ice grows from seawater with a salinity 30–33‰; the intentional

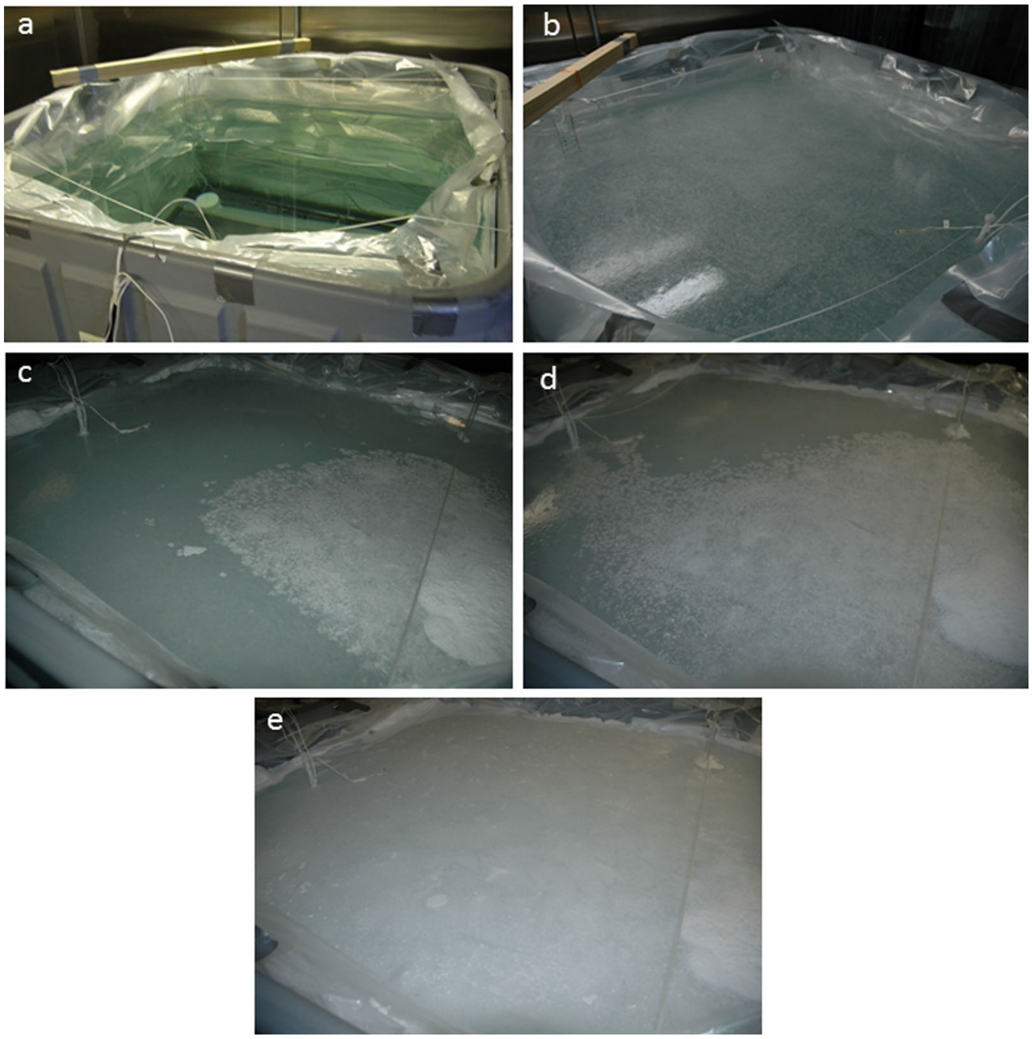

Figure 3. Progression of tank surface during cooling. (a) Liquid water immediately after tank was filled, (b) initial ice growth, (c) initial hydrohalite precipitation, (d) further hydrohalite coverage, and (e) complete hydrohalite coverage. The room temperature was set to −30°C during this sequence.

reduction in tank salinity was a shortcut necessitated by the limited volume of the laboratory tank. (When growing ice in an isolated tank, the residual melt concentrates quickly as salt is partially excluded at the ice-melt interface. Excessive salt in the melt causes laboratory ice to have excessively high brine volume and low structural integrity, at least until the entire sample is cooled below T_{eu}.) Furthermore, the paleoclimatic ocean salinity is not well constrained and the single most relevant attributes of the ice grown for this work are that it contain salt, and that salt be distributed within the ice in a manner representative of naturally grown congelation ice. The tank was not filled to the rim given the expected expansion during freezing, so the water level was brought to approximately 15 cm below the top of the tank.

To encourage the concentration of salt within the surface layers of the ice, the ice was grown as quickly as possible [*Weeks and Ackley*, 1986] by setting the room temperature to its lowest setting, −30°C. This was done in attempt to concentrate salt in the surface layers of the laboratory ice. This shortcut was motivated by the need to approximate the effects of many years of ice sublimation, a time scale not practical for laboratory studies. Additionally, this study was intentionally focused on simulation of the initial sea ice cover, and no attempt was made to maintain a thick ice cover in steady state. The temperature of the room was then kept at −30°C throughout the crust development phase of the experiment. The surface progression during ice growth is illustrated in the series of photographs shown in Figure 3. The progression shows the tank containing liquid water, an ice skim, thicker ice coverage and the onset of hydrohalite precipitation,

Table 1. Summary of Albedo Experiments

Experiment	Description	Timing	Notes
1	Solid ice block, hydrohalite precipitated within brine inclusions, but no measureable surface sublimation	30 days ice growth	Average $T_{ice} = -26°C$; isothermal block
2	Sublimating crust	1–42 days after application of concentrated NaCl in H_2O	Average $T_{ice} = -26°C$; sublimating crust
3	Maximum crust thickness	42 days	
4	Dissolving lag deposit	8 h	
5	Induced ponding	2 days	

more extensive hydrohalite precipitation, and full surface precipitation. After freezing, approximately 5 cm of the gray tank wall was exposed between the ice surface and the tank rim.

Initial experiments attempted to form a salt crust by allowing the ice to sublimate passively. Despite lack of humidity control in the freezer lab, a salt crust did develop slowly over a period of several months, and some preliminary measurements were acquired. However, mechanical failures in the cold room refrigeration repeatedly destroyed the fragile crust by allowing temperatures to rise above the eutectic temperature. While working with prototype surfaces in smaller buckets, we were able to hasten a lag deposit by blowing cold, desiccated air across the ice surface of smaller samples [*Light et al.*, 2009]. Such an approach was not practical for this larger tank. To promote the formation of a surface lag deposit, we therefore took a more proactive approach. A nearly saturated (233‰) NaCl solution was mixed (see Figure 2), chilled to just above its eutectic temperature, and sprayed onto the cold ($-26°C$) ice surface using a trigger-type pump spray bottle. This near-eutectic mixture was sprayed in thin, even layers. Each new application was allowed to solidify for between 5 and 24 h before the next was added. After the application of 11 layers (approximately 2 mm average total thickness), dark-colored markers that had been placed on the ice surface before the spraying became obscured, so the surface was judged to be optically thick, at least at visible wavelengths. While this "shortcut" may have lessened the fidelity of our surrogate material, its similarities to an actual salt crust built up over time should be more notable than its differences for purposes of exploring its optical properties. Of course one difference that is entirely unaccounted for is the grain size of the salt crystals, which could have implications for light scattering, and this will be addressed below (see section 4.1).

Spectral albedo measurements were made for five distinct surface types: (i) cold, subeutectic bare ice, (ii) sublimating ice, (iii) optically thick surface lag deposit, (iv) dissolving hydrohalite crust, and (v) puddled surface with standing liquid. Table 1 summarizes the five experiments and their parameters. After the application of the concentrated solution, the surface was monitored for a total of 90 days. The first 42 days of the experiment were used to monitor the evolving albedo of the sublimating crust. Little to no change occurred in the crust thereafter, until day 90 when the crust was intentionally dissolved. During days when albedo was not being measured, the measurement apparatus was elevated on blocks placed between it and the edge of the tank, and a small electric fan was positioned to pull ambient air through the gap to help speed the sublimation process.

Albedo measurements were carried out using the "albedo dome" technique described by *Light et al.* [2015]. The central element of this technique is a rigid acrylic hemispherical dome that is placed over the entire top surface of the tank. The interior surface of the dome is treated with a diffusely reflective coating; broadband light from a lamp mounted exterior to the dome enters through a small chimney in the dome and is reflected from a small Lambertian plate mounted at the dome origin to illuminate the dome interior. Irradiance estimates are made of the downwelling light field as it leaves the dome wall and of the upwelling light field as it backscatters from the ice using a fiber optic probe coupled to a spectroradiometer. Each albedo is computed from a set of 3–10 individually measured upwelling/downwelling irradiance pairs. Since measurement of albedo involves only relative quantities, and since our measurements were spectrally resolved, it was not necessary to use a source that could simulate spectral solar output. Rather, a tungsten-halogen source was used, and ratios of the two irradiance fields were computed at each wavelength. For these measurements, a 750 W lamp was used. Concerns for the potential of heat generated by the lamp to thermally alter the surface were allayed by careful monitoring of the ice surface temperature, which never exceeded $-26°C$. The lamp was typically turned on for no more than 45 min at a time, and, even though

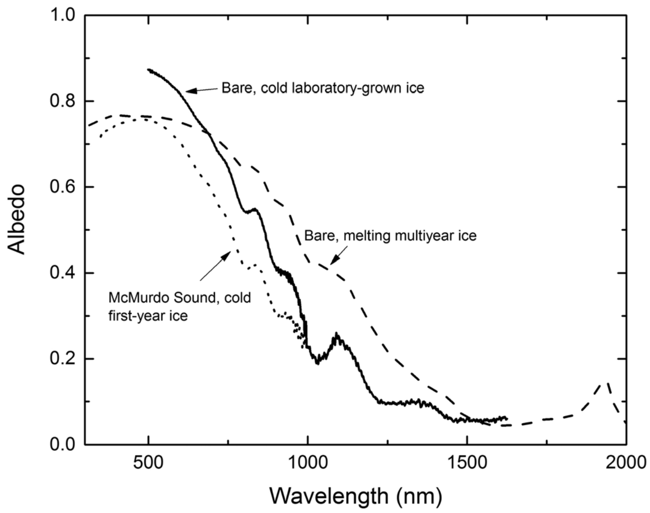

Figure 4. Measured spectral albedos for (1) bare subeutectic laboratory-grown sea ice (solid line, this study); (2) bare melting multiyear sea ice (measured in the field, dashed [*Perovich et al*., 2002]); (3) cold first-year sea ice in McMurdo Sound, Antarctica (also in situ *Carns et al*. [2015]) dotted.

the lamp power was 750 W, we estimate that after losses are accounted for due to lamp geometry, collimation optics, and diffuse scattering from the plate and dome interior that the irradiance at the ice surface was order 100 W m^{-2}. The more serious concern of extraneous heat from the 750 W lamp actually came from the exterior of the dome, which was in close proximity to the lamp and collimator. For this reason, the exterior was masked with reflective shielding, so as to protect the dome walls from warming from the outside and radiating onto the ice surface.

The albedo measurement technique also relies on the assumption that the "optical footprint" on the ice surface, which is the target area measured, is representative of the entire surface. The total measurement time for an albedo estimate was between 5 and 30 min, depending on the number of upwelling/downwelling pairs measured. During the crust-formation phase, the surface changed slowly, so 10 pairs were typically measured during this phase. During the crust demise phase, however, the character of the surface changed quickly and was therefore more sensitive to the time needed for a complete a set of measurements. Three pairs were typically measured during this phase.

3. Measured Spectral Albedos

3.1. Hydrohalite Within Brine Inclusions

Measurements began once the entire tank had solidified (approximately 1 m thick ice), prior to significant surface sublimation. The block was nearly isothermal with average temperature −26°C. There was visible evidence of hydrohalite precipitation within the ice, but no obvious crystal accumulation on the ice surface. Such a bare, subeutectic surface is optically important, since it would likely be a precursor to the formation of a salt crust, as well as the underlying ice type once a lag deposit formed. Figure 3b shows a photograph of the ice surface at this stage, which is clearly not homogeneous. The patchy areas of bright white surface are areas of higher-density hydrohalite. Once measurements began, the ice temperature was entirely below T_{eu} for the NaCl-H_2O system, so we assumed that the entire depth had precipitated hydrohalite, although the precipitation at depth was not confirmed.

The measured spectral albedo for this surface is shown in Figure 4. For comparison, the average spectral albedo measured on natural bare, melting multiyear ice [*Perovich et al*., 2002] in the Arctic basin is shown, along with bare, subeutectic ice in McMurdo Sound [*Carns et al*., 2015]. All of these ice types have high albedo at visible wavelengths, decreasing at near-infrared wavelengths, consistent with the absorption properties of pure ice and liquid brine. Variations in their relative magnitudes are attributable to the presence or absence of a surface scattering layer, which strongly increases the near-infrared albedo for the melting ice. Differences between the bare, cold laboratory-grown ice, and the cold first-year ice at McMurdo Sound may be due to several factors that could maintain a lower natural albedo. The laboratory ice was subeutectic all the way down to the base of the tank (approximately 1 m), whereas the natural ice had a thinner subeutectic layer, as the −22.9°C horizon was at 10 cm depth. Also, as *Carns et al*. [2015] demonstrated, it is likely that many of the inclusions in the natural ice failed to precipitate hydrohalite right at the temperature of initial precipitation. Furthermore, the ice in the tank was composed of the NaCl-H_2O binary, whereas the natural sea ice would have contained the full complement of sea salts, in which hydrohalite precipitation

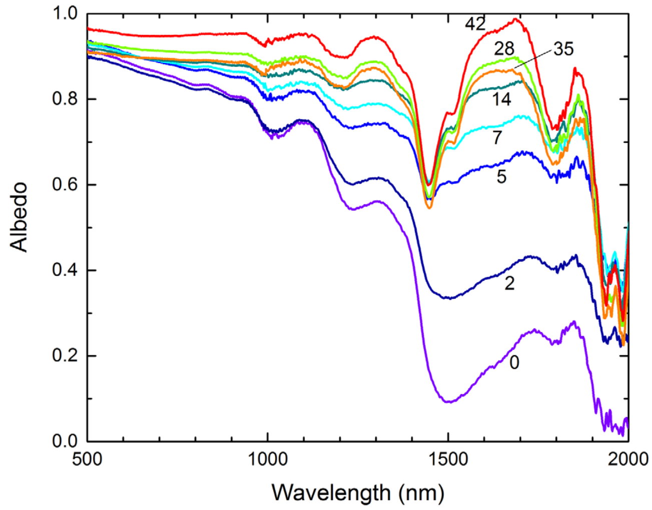

Figure 5. Measured spectral albedos for laboratory-grown sea ice with a sublimating crust. The number of days since the onset of sublimation is indicated for each curve.

does not culminate until the ice temperature drops below −30°C, and liquid brine remains until all the salt has precipitated.

3.2. Development of Salt Crust

Following the measurement of bare, subeutectic ice, a hydrohalite crust was produced using the spray technique discussed previously. Figure 5 shows the evolution of the observed spectral albedo as the crust desiccated over a 42 day period, as measured every 2–7 days. The cold room thermostat remained at −30°C throughout this time and the ice temperature remained approximately isothermal at −26°C. The albedo measured on day 42 was the highest observed at virtually all wavelengths, with remarkably high near-infrared values. While the geometric thickness of the spray crust was estimated to be about 2 mm, it was difficult to monitor the movement of the sublimation front as it progressed downward, creating the lag deposit at the surface. Albedo changes appeared to be largest at wavelengths longer than the limit of the visible spectrum. For this reason, changes in the visual appearance of the crust were not readily apparent.

3.3. Warming, Melting, and Dissolution

Once the albedo stopped increasing, it was assumed that the surface had undergone sufficient sublimation that the crust was optically semi-infinite. We then subjected the surface to melting. We anticipated that the melt evolution would happen quickly, so we monitored the albedo frequently during a single day. In preparation for surface melt, the temperature of the laboratory was increased to just below T_{eu} for several days. By warming the ice block up to just below its eutectic temperature, we anticipated that the precipitated hydrohalite throughout the block would dissolve readily upon further warming. This was done in an attempt to keep the physical properties of the block as vertically uniform as possible so that results could be more easily interpreted.

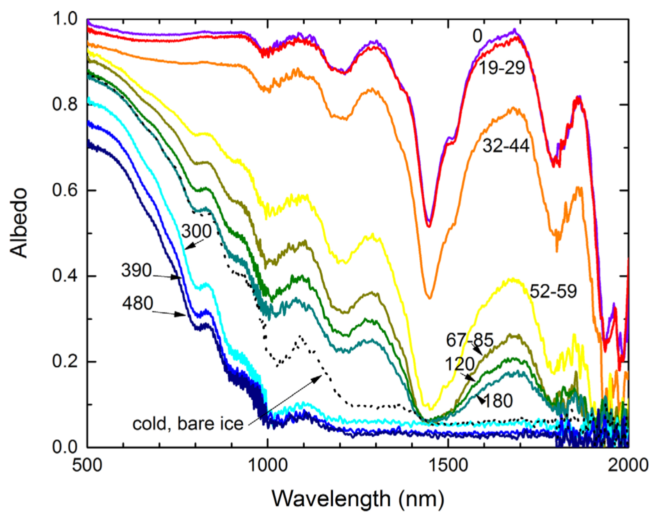

Figure 6. Spectral albedo during the warming experiment. The number of minutes since the room thermostat was increased from −22 to 0°C is indicated for each curve.

The objective was to create a puddle of standing liquid brine on the ice surface. We were, however, concerned that the meltwater from the warming surface would run off the surface of the block, since its center was slightly domed from the pressure created during freezing in a rigid-walled tank. We constructed barriers on the edge of the tank by placing strips of foam on the ice surface and slowly saturating them with fresh water, creating a fresh-ice barrier while disrupting the surface as little as possible. No foam was installed within or proximal to the optical footprint, and tests confirmed that the measured albedo of the

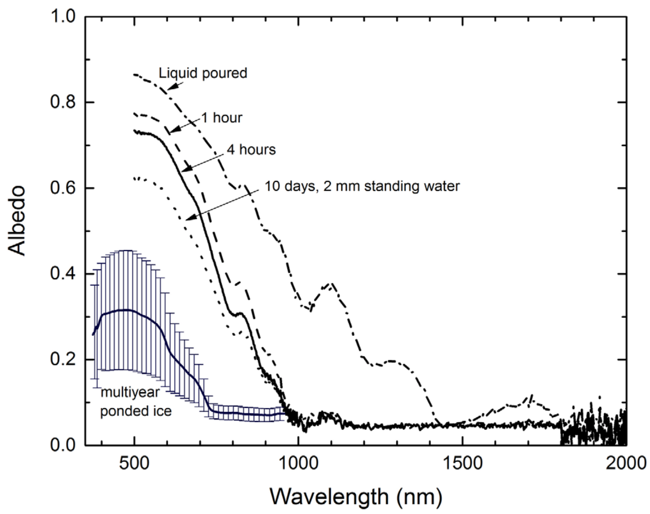

Figure 7. Various phases of spectral albedo during our ponding experiment compared to average spectral albedo for ponded multiyear ice in nature [*Perovich et al.*, 2002].

surface appeared unchanged after the construction of the barriers.

In preparation for melt observations, the temperature of the room was increased from approximately −22°C to slightly below 0°C. The albedo was then measured approximately every 20 min during the first 2 h, and at 1–2 h intervals thereafter. Three separate measurements of albedo were made at each interval; measurement of each set of three incident and backscattered irradiance pairs took 7–12 min. Figure 6 shows the spectral albedo observed during the demise of the surface lag deposit. The albedo decreased rapidly once the air temperature was raised above T_{eu}.

Our attempt at producing a ponded surface resulted in a very thin transient layer of liquid brine. It was not entirely clear whether puddled melt drained laterally as a result of the slightly domed surface, or whether the ice surface became permeable and some of the puddled liquid brine percolated downward through the ice. In either case, we would expect natural ice under Snowball conditions to be capable of retaining surface melt, as domed ice would not likely be found in nature and ice deformation processes would simply lead to catchment depressions, as happens on the modern sea ice cover in summer, and because the permeability threshold for congelation sea ice with realistic salinity tends to occur at significantly higher temperatures (−5°C) [*Golden et al.*, 2007]. In an effort to enhance liquid retention at the surface, as we would expect in nature given the widespread occurrence of melt ponds on Arctic sea ice, we attempted to rapidly reduce the permeability of the surface of the ice block. To do this, NaCl crystals were sprinkled on the wet surface and the temperature of the room was returned to −30°C. Subeutectic ice with large salt content tends to be very hard and it was hoped that application of such a layer would form an impermeable barrier. In an effort to create a liquid pond, we then gently introduced modest amounts of cold (−20°C), saturated NaCl-H_2O liquid onto the ice surface. The albedo was then measured at four different times: (a) time of pouring, with room thermostat held at −20°C, (b) 1 h after pour, (c) 4 h after pour, and (d) after increasing the room temperature to −15°C for 3 days, during which a puddle formed with 2 mm depth. The salinity of the pond was measured to be about 60‰ and the ice surface temperature was about −3°C. While it is possible this laboratory "shortcut" could have obscured natural processes relevant to sea ice surfaces on Snowball Earth, such as the immediate drainage of surface meltwater, we deemed the possibility of cold, saline puddles present on cold, saline ice potentially relevant and so elected to include them in the set of albedo observations presented by this study. While crust formation is expected to be a very slow process in nature, it appears that liquid puddles may occur on diurnal time scales, much as sometimes occurs in the modern Arctic at the beginning of the melt season. Melting during daytime and refreezing at night would result in lower albedo, because a refrozen melt pond has lower albedo than a salt crust.

Figure 7 shows the measured spectral albedo progression for the three stages of forced ponding. Upon further warming, the ice became porous, drained, and the puddled liquid drained entirely, leaving a pitted ice surface with no standing water. The measured albedos are shown in comparison with field albedos for ponded multiyear ice [*Perovich et al.*, 2002]. The ponded field albedo is much lower, partly because the natural pond was much deeper, but also because the ice beneath the pond was warm, rotting, and had lost most of its air bubbles.

4. Discussion

The albedo of sublimating, cold, snow-free sea ice has implications for the advance of sea ice during cold climatic conditions. The possibility of a runaway ice-albedo feedback was outlined by *Budyko* [1969], but

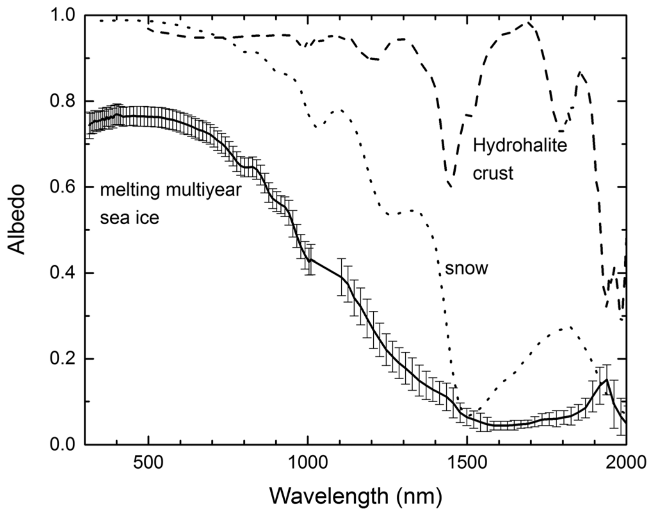

Figure 8. Highest spectral albedo measured for laboratory lag deposit, compared with spectral albedo of Antarctic snow [*Hudson et al.*, 2006, Figure 6], and bare, thick melting multiyear sea ice with a drained layer of granular ice on its surface [*Perovich et al.*, 2002].

the exact magnitude of the albedo of the ice is subject to environmental conditions. At low temperature, the precipitation of hydrohalite was observed to strongly enhance the ice albedo. Such an enhancement could amplify a positive ice-albedo feedback. Likewise, the albedo for warming ice with a hydrohalite lag deposit appears to rapidly decrease as salt crystals dissolve and puddles form, even at temperatures lower than $-20°C$. Such a rapidly decreasing albedo could also amplify a positive feedback whereby decreased ice albedo promotes the absorption of additional heat and the subsequent promotion of surface melt.

4.1. Albedo Spectra of the Crust

Figure 5 shows the albedo observations made as a deposit of hydrohalite and ice was created on the surface of cold, subeutectic ice, and then as the ice in this surface layer slowly sublimated, leaving a hydrohalite lag deposit. The albedo increases sharply as the ice sublimates because hydrohalite is less absorptive than ice. The albedo increase is most pronounced at near-infrared wavelengths, especially between 1500 and 1900 nm. Visible-wavelength albedo is relatively high for the initial salt-rich crust, so increases there are more subtle. The sublimating crust was observed in our laboratory tank for 90 days, but measurements made after day 42 showed no increases in albedo, and in fact even slight decreases. We assumed the albedo stopped increasing either because the crust had become optically thick or because the sublimation front hit the bottom of the hydrohalite-enriched layer, e.g., the interface with the lower-salinity ice surface, and further ice sublimation revealed only small additional amounts of hydrohalite. This would cause the increase in albedo to slow considerably even if the hydrohalite layer was not yet optically thick. However, given that the albedo had not measurably increased after 38 additional days, it seems likely that the hydrohalite crust was effectively optically semi-infinite. The slight albedo decreases observed at the end of the cold phase may have resulted from accumulation of fine dust on the surface with time, or frost deposition onto the surface, or metamorphism of the hydrohalite crystals themselves. None of these possibilities were quantified during this study.

The growing hydrohalite crust developed a strong peak in albedo at $\lambda = 1.7$ μm, as shown in Figure 5. Figure 8 contrasts the spectral albedo of the thickest crust to that of snow. The albedo of the hydrohalite crust measured in this study is higher than that of fresh snow at all solar wavelengths longer than about 750 nm. In fact, at wavelengths longer than 1200 nm, it is more than 50% higher than snow. In the region 1.5–2.0 μm, snow shows only one peak, at 1.84 μm, whereas the hydrohalite crust shows a broad peak at 1.6–1.7 μm and a weaker peak at 1.86 μm. NaCl has no absorption bands in the near-infrared, but hydrohalite has two bound water molecules, so it is plausible that hydrohalite could retain some features of liquid water in its absorption spectrum. We should additionally expect to see absorption by ice crystals within and beneath the crust. The absorption spectrum of liquid water resembles that of ice, but shifted slightly to shorter wavelengths [e.g., *Dozier*, 1989, Figure 1]. Water absorption [*Querry et al.*, 1991] peaks at 1.45 μm, where the hydrohalite crust albedo shows a sharp minimum, and ice absorption [*Warren and Brandt*, 2008] peaks at 1.5 μm, where the hydrohalite crust albedo shows a shoulder. Water absorption peaks again at 1.93 μm, and ice at 2.0 μm; these strong absorptions are seen at the right-hand end of Figure 8. The minimum albedo at 1.80 μm is more difficult to explain [see *Carns et al.*, 2016]. Interestingly, it appears that the very low absorption of the salt and the relative absence of ice in the crust are primarily responsible for the high albedo of the crust. Given the manner in which the crust was formed in this study, there is certainly no guarantee that the salt grain size is representative of natural salt grains left behind by sublimation of a naturally grown ice cover. The direct relationship between grain size and scattering cross section make this result potentially sensitive to grain size variation, although the lack of absorbing material at the surface of such a deposit is likely to be more important than the details of grain size, as long as the crust is thick enough to be opaque, as in the case of an optically thick snow pack [see *Warren*, 1982].

4.2. Can the Surrogate Surface Represent a Natural Process on Snowball Earth?

The intentional application of a hydrohalite-rich layer on the cold ice surface sped the process of accumulating enough hydrohalite to form a perceptible deposit. About 30 times as much ice must sublimate from natural sea ice to leave the same thickness of hydrohalite crust as was formed in the laboratory. The chilled NaCl solution sprayed onto the tank was not observed to melt the ice beneath it, so we assumed that the applied layer of hydrohalite-rich ice contained the same relative amounts of sodium chloride and water molecules as the initial solution. The original 233‰ solution was 36% hydrohalite by weight, or 25% hydrohalite by volume (once frozen) if we assign hydrohalite a density of 1.6 g/cm^3 [*Adams and Gibson*, 1930]. Compared with sea ice with a salinity of 10 parts per thousand, which would only have 1.6% hydrohalite by weight or 0.9% by volume, this imposed layer is salt rich.

Inspection of the crust during the sublimation process did not provide clear indication as to how the sublimation front progressed through the ice. The surface material clearly became progressively less cohesive. The newly frozen hydrohalite-rich ice was hard and required firm scraping to sample, while the developing hydrohalite crust was flaky and easy to remove from the surface.

Although the hydrohalite-rich ice layer was created in the laboratory for reasons of practicality, it is an analog to a situation that could arise naturally on a Snowball planet: the partial or complete dissolution and reformation of an established hydrohalite lag deposit. This situation occurred in our laboratory tank when the freezers malfunctioned during an early run of the experiment. A hydrohalite crust had developed over a period of time below -21.2°C; when the temperature rose above -21.2°C, the crust dissolved. Pure hydrohalite with no access to ice or liquid water is stable at temperatures up to 0.11°C, but in contact with ice above the eutectic temperature it will melt the ice [*Marion and Grant*, 1994], forming a concentrated brine. When the laboratory temperature dropped back below the eutectic point, the brine that formed refroze as a layer of hydrohalite-rich ice. Over time, a salt crust gradually reformed. The same sequence of events—crust formation at subeutectic temperature, dissolution with increasing temperature, refreezing upon subsequent cooling—could be expected to happen at various times and places in the tropics of Snowball Earth.

4.3. Melt Ponds at Temperatures Significantly Below 0°C

The formation of high-salinity melt ponds could be important for understanding how the surface heat balance of a sea ice cover responds to warming events, even at temperatures below -20°C. The observation of surface melt processes along with decreased albedo on warming, at a temperature still below -20°C, confirms the possibility for an important positive "salt-albedo feedback" in sea ice, as hypothesized by *Light et al*. [2009].

Upon warming the surface above T_{eu}, the initial indication of change was that the crust became smooth in appearance, and dark patches appeared to form beneath the crust. The albedo began to drop rapidly around 30 min after the beginning of the experiment. Dark patches were visible on the surface of the crust where liquid brine was beginning to saturate the hydrohalite crystals. By 120 min after the onset of warming, obvious specular reflections showed that a film of liquid brine covered the hydrohalite crystals, and the albedo was only marginally higher than the albedo of the original bare subeutectic ice. Increasing amounts of standing water accumulated in low spots, some estimated to be as large as a few centimeters in diameter, but puddles were never deeper than a skim. By hour 5, the albedo had dropped below that of subeutectic ice and it continued to drop slightly over the next three hours. The albedo decreased below that of the bare subeutectic ice as hydrohalite in the ice below began to dissolve. At this point, ice in the tank was nearly isothermal, so it is difficult to say how deep within the ice the dissolution front progressed, but it is clear that most of the hydrohalite within the ice near the surface had dissolved. The continuing decrease in albedo likely represents a decrease in total scattering as crystals of ice melted and hydrohalite dissolved in the brine. At this point, the ice surface most closely resembles ordinary sea ice during the modern summer melt season. These observed changes in the surface during melt appeared to be relatively uniform across all areas of the tank.

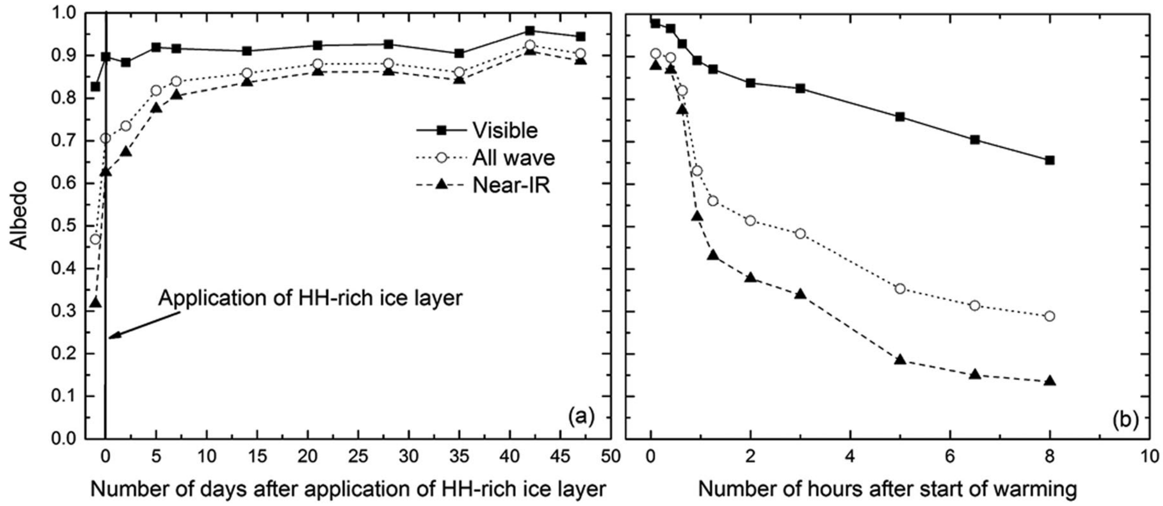

Figure 9. Evolution of broadband albedos for (a) cooling and (b) warming experiments.

4.4. Broadband Albedos

The relative strength of the sea-ice albedo feedback depends on the contrast between the albedo of the open ocean and the albedo of the frozen ocean. The larger the difference, the stronger the positive feedback. In the modern Arctic, the relevant albedo for the frozen part in summer is bare, melting, typically ponded sea ice, and in autumn and spring is snow-covered ice. Regardless of whether the ice is snow covered or the snow has melted, the albedo is generally much lower than the albedo of subeutectic sea ice with a hydrohalite crust. The difference is significant across much of the solar spectrum, but the contrast is most pronounced at near-infrared wavelengths. The largest albedo we measured was for a hydrohalite crust approximately 2 mm thick observed after 42 days of desiccation. The comparison with freshly fallen snow (which is the highest naturally occurring albedo measured in the modern Earth's cryosphere) and with bare, melting multiyear sea ice is shown in Figure 8. While the all-wave albedo of the crust is 0.93, comparable albedo for fresh snow is 0.83, and 0.67 for bare, melting sea ice.

The integrated broadband albedos for the two phases of the experiment are shown in Figure 9. For ease of comparison with other measurements they are calculated for 300–700 and 700–2500 nm. The laboratory measurements are noisy below 500 nm and above 2200 nm, so the calculation assigns to the 300–500 nm region the mean albedo value from 495 to 505 nm, and assigns to the 2200–2500 nm region the mean albedo value from 2150 to 2200 nm. The broadband albedo of the hydrohalite crust itself rises above 0.9 at day 5 and stays there. This albedo is higher than that of sea ice with thick dry snow, which is $\sim$0.84 [*Brandt et al.*, 2005]. Table 2 shows approximate broadband albedos used for modeling typical surfaces on a frozen ocean and albedos derived from the measurements made in this study. Our observations indicate the subeutectic crust has albedo at visible wavelengths comparable to fresh snow, and near-infrared albedos larger than snow and much larger than bare, melting sea ice.

4.5. Suggestions for Future Work

This experimental treatment of laboratory-grown saline ice does not perfectly imitate the behavior of natural sea ice. The choice to restrict the chemistry to the binary NaCl-H_2O system and to not include the full

Table 2. Wavelength-Averaged Albedos for Snow-Covered Ice, Bare and Ponded Melting Ice, Bare Subeutectic Ice, Hydrohalite Lag Deposit, and Puddled Crust

Ice Type: Spectral Region	Snow (r_{gr} = 100 μm) [*Dang et al.*, 2015]	Bare, Melting Sea Ice [*Perovich et al.*, 2002]	Ponded, Melting Sea Ice [*Perovich et al.*, 2002]	Bare, Subeutectic Sea Ice (This Study)	Hydrohalite Lag Deposit on Sea Ice (This Study)	Puddled Hydrohalite Deposit (This Study)
Visible 0.3–0.7 μm	0.985	0.75 $\pm$ 0.02	0.25 $\pm$ 0.11	0.84	0.95	0.67
Near IR (0.7–2.5 μm)	0.660	0.57 $\pm$ 0.03	0.09 $\pm$ 0.02	0.37	0.90	0.19
All wave	0.834	0.67 $\pm$ 0.03	0.17 $\pm$ 0.11	0.64	0.93	0.47

complement of sea salts resulted in part from the inability to achieve temperatures in the freezer laboratory that would allow the study of subeutectic sea ice with the full complement of salts. This decision was also motivated by the desire to keep the experiment and its analysis simple. Interactions between salts in the multisalt system are complex. Regardless, about 82% (by mass) of the total salt in seawater precipitates as hydrohalite, so this should be a useful proxy. Had the full salt system been used, and had temperatures been low enough, we would expect similar results. Of course, the Neoproterozoic ocean chemistry is not definitively known, but for modern seawater chemistry, the predominant cryogenite would be hydrohalite.

The exact nature of the salt deposit would depend on both the composition of the ocean and the surface temperature. If temperatures remain lower than T_{eu} for the mixture of salts, a dry deposit with very high albedo could form, possibly promoting further cooling of the climate. While this deposit was slow to develop in the laboratory, net-evaporation environments on Snowball Earth would be exposed to strong sunlight and wind that could possibly accelerate the process. If hydrohalite has precipitated but liquid brine is still present, it is unclear what the expected albedo would be. Precipitated salts suspended in liquid brine would have reduced backscattering, relative to their dry crystal counterparts, but this scenario was not tested. If a dry deposit forms while the surface is below the eutectic temperature, but later warms above this temperature, the deposit will rapidly dissolve and reduce the albedo of the surface, introducing the possibility of forming brine puddles on the surface. The dissolution phase of our experiment shows how quickly this process can occur.

Another limitation of this investigation is the laboratory light source. The only radiometric measurements made were relative, so there is no dependence on the spectral output of the source. However, despite the brightness of the 750 W bulb, considerable light is lost to the laboratory and the lamp interior, and light levels within the dome are low compared to the conditions for which the spectrophotometer we used is designed. Measurements are notably noisy in regions of the spectrum where photon counts are low. We compensated by using multiple measurements to reduce noise in the final result. Validation results suggest that the low counts do not introduce any wavelength-dependent bias [see *Light et al.*, 2015]. The exception is wavelengths shorter than 500 nm, where reported albedos are consistently too high; the effect (which is a result of low output at these wavelengths from the light source) is too unpredictable to fix using a correction, so we simply do not report albedos at shorter wavelengths in this paper. The signal-to-noise ratio at wavelengths shorter than 500 nm and longer than 2200 nm was too low to make use of those data.

5. Conclusions

Under low-temperature conditions, the presence of salt in sea ice introduces the possibility of a positive salt-albedo feedback which has not been previously considered. Bare sea ice exposed to low temperatures for long periods of time in a net-sublimation environment is hypothesized to develop a lag deposit of precipitated salt crystals. The large increase in observed albedo suggests this mechanism could be a powerful positive feedback operating at temperatures below the temperature of initial hydrohalite precipitation. Upon warming, our laboratory observations showed the demise of the lag deposit and inception of high-salinity surface melt puddles composed of concentrated brine. These melt ponds occurred at a temperature just above the eutectic temperature of the ice ($-21.2°C$ in our study of the NaCl-water binary). These ponds significantly and rapidly decreased the surface albedo, suggesting this low-temperature positive feedback could promote further melting and absorption of solar radiation. The measurements of ice albedo during the evolution of the ice surface as it cooled and as it warmed have implications for parameterizations used for climate simulation of Snowball Earth, including a potentially strong positive albedo-salt feedback capable of operating during ice growth as well as during ice demise.

Planetary climate states are tightly tied to surface albedo. Transition into and out of a Snowball state would have been facilitated by, and probably required, surface albedo feedbacks. The details of the surface energy balance of course can be treated in numerical climate simulation, but only if the models are fed appropriate parameterizations for the temperature-dependent ice albedo. Presently, models for Snowball Earth use a surface albedo scheme that is generally restricted to four ice types: snow covered (cold), bare (cold), melting bare (warm), and ponded (warm). Even if hydrohalite crusts were not pervasive on a cold sea-ice surface, the ice would likely have contained precipitated salts. Our studies demonstrate that even this cold bare ice, without a salt lag deposit, has higher visible albedo than the bare ice albedo for higher (but still subfreezing) temperatures.

Time constraints precluded the ability to create a thick lag deposit on the ice surface in our laboratory. Our solution to this constraint was to apply additional concentrated salt solution to the cold ice surface. It is possible that this proxy illustrates additional processes that need to be considered. What if the slow accumulation of salt at the surface resulted in ample time for that salt to become redistributed by wind? Perhaps the crystals would never accumulate. Or, it is possible, that like snow on modern Earth, the crystals could redistribute, but still accumulate, perhaps even drifting in ways analogous to snow? Alternatively, a lag deposit in nature could be considerably thicker than what was formed in the lab, so it is possible that the albedo of such a deposit would be even larger, at some wavelengths, than what we observed.

Another subject for further study is the optical properties of the various phases of natural sea ice containing the full complement of salts. One likely scenario includes the simultaneous presence of both salt crystals and liquid brine. Such a scenario would result from sea ice at temperature higher than the multisalt eutectic, but low enough to have precipitated salt within surface puddles. The cycle of cooling, formation of a lag deposit, and subsequent melting could occur many times over; the results of such a cycle would be worth investigating. Additional questions relevant to this study involve a better understanding of the permeability of cold sea ice to high-salinity brine. In the laboratory, crust formation took place over a period of many days. The crust demise, however, took only hours. These laboratory time scales may not be directly applicable to nature, but they should be indicative of the necessity for long periods required for the development of a high-scattering cryogenite lag deposit, relative to its rapid demise.

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