Disposition of solar radiation in sea ice and the upper ocean

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Abstract. A one-dimensional sea ice model with an ice thickness distribution is presented to examine the disposition of the incoming surface shortwave radiation within the sea ice and the upper ocean. The sea ice model consists of 15 different ice thickness categories and an open water (leads) category. Ice growth, melting on horizontal and vertical surfaces, meltwater pond growth and drainage, and snow accumulation evolve independently for each of the ice categories. Leads, melt ponds, and thinner ice categories are of particular interest, as these features account for nearly all of the solar energy that is transmitted into the upper ocean beneath the ice pack. Also examined is the surface reflection and absorption, internal ice absorption, and lateral melting for ponded and pond-free ice. Area-averaged results show that 69% of the total annual solar energy is reflected, 15% is absorbed by the snow cover, 12% is absorbed by the ice, and 4% is transmitted to the ocean mixed layer through thin ice and leads.

1. Introduction

The average equilibrium thickness of the ice in the Arctic Ocean is about 3 m. However, the Arctic Ocean sea ice cover is heterogeneous on horizontal scales of tens of meters, with variations in sea ice thickness ranging from zero in leads ("cracks" in the sea ice) to as much as 20 m in some pressure ridges; these irregular properties are a result of the dynamical behavior of the ice pack. Associated with these thickness variations are variations in ice age, snow cover, albedo, and many other parameters. When this heterogeneous surface is exposed to sunlight in summer, the response depends on the aggregate response of the local ice characteristics. Once melting begins, the snow cover quickly disappears, leaving a surface covered by a mixture of bare ice and melt ponds. The relative proportions of bare ice, melt ponds, and leads change continually throughout the melt cycle.

An understanding of the shortwave (SW) radiative interactions between the atmosphere and the sea ice is critical for addressing issues related to the stability of the Arctic Ocean ice pack and cloud-radiative feedback processes. Depending on the nature of the surface and the cloud cover, only a portion of the net shortwave input to the ice goes immediately to surface melting, the remainder being stored as latent heat in melt ponds and internal brine pockets or transmitted to the ocean. Radiation absorbed within the ice changes the ice structure and its optical properties. Substantial differences in brine volume, extinction coefficient, air bubble distribution, and ice thickness exist between first-year and multiyear ice and ponded and pond-free ice, with implications for the surface albedo and mass balance.

The upper Arctic Ocean appears to be thermodynamically decoupled from the deeper ocean in much of the basin, so that most of the oceanic heat flux to the bottom of the ice must be derived from shortwave radiation transmitted through leads and thin ice. Additionally, biological activity within the ice and the upper ocean depends critically on the light levels [e.g., Perovich et al., 1993]. An algal bloom occurs during the months of April and May, resulting in high algae concentrations in the 2-cm thick bottom skeletal layer of both seasonal and multiyear ice [Maykut and Grenfell, 1975]. This growth is more vigorous near the coastal regions and ends abruptly at the onset of bottom ablation.

The penetration of radiation into the ocean has been investigated by Maykut and Grenfell [1975] and Perovich and Maykut [1990]. Maykut and Grenfell [1975] measured the spectral distribution of the downwelling irradiance for first-year ice near Point Barrow. The following four ice types were considered: snow-covered ice, melting white ice above the local water table, blue ice saturated with water, and ice covered by melt ponds. Substantial differences in transmission were found beneath melt ponds, white ice, and snow-covered ice, with no differences detected beneath saturated blue ice and the ice beneath shallow (5-10 cm) melt ponds. For ice with a thickness of about 1.85 m, the ice covered by melt ponds transmitted about 3.3 times more SW than white ice, and the white ice transmitted 10 times more SW than the ice covered with 25 cm of snow. Perovich and Maykut [1990] found that the incoming SW flux, combined with lack of ice movement, input of freshwater flux, and weak water circulation, split the water column into the following three layers: a well-mixed, fresh surface layer; a stable halocline; and a layer of constant salinity. In this stable halocline or pycnocline is a temperature maximum that effectively traps incoming SW radiation, reducing ice growth in the fall.

Detailed modeling studies of radiative transfer in sea ice have been conducted by Grenfell [1979] and Jin et al. [1994]. Using a two-stream model, Grenfell [1979] examined the energy absorption in blue and white ice. It was found that energy absorption at the surface is greater for white ice owing to its high
extinction in the surface layer; more radiation penetrates blue ice, since it has no scattering layer. The absorption of radiation at the surface is independent of ice thickness, but below the surface layer the amount of transmitted radiation deviates significantly from Beer’s law. Jin et al. [1994] found that over 50% of the net solar flux is absorbed in the top 10 cm of multiyear ice and is a strong function of wavelength. Only visible radiation penetrates deeper into the ice and into the ocean. The presence of clouds or a snow cover drastically reduces the absorption within the ice. Air bubbles trapped in the ice scatter SW radiation, allowing less transmission to the ocean. The absorption of solar energy is also a function of ice thickness, with the absorption coefficient decreasing as ice thickness increases, due to the decrease in air bubble content with age.

In this paper we use a one-dimensional thermodynamic sea ice model with an interactive ice thickness distribution to investigate the disposition of solar radiation among surface reflection, surface absorption, and internal absorption and transmission in sea ice for a single summer season. Differences between ponded and pond-free ice are highlighted.

2. Model Description

A one-dimensional sea ice model is employed, having a sophisticated surface albedo parameterization that is highly sensitive to the surface state and allows complex spectral radiative interactions to occur between the sea ice and the atmosphere. A "slab" version of this model has been described by Ebert and Curry [1993] and Curry and Ebert [1992] that includes the following features: specified ice divergence that allows for ice export, a lead parameterization that includes lateral ablation and accretion, ocean heat flux that is tied to warming of the ocean mixed layer by the penetration of solar radiation through the ice and in leads, a melt pond parameterization that allows for pond runoff and a variable pond area and depth, and an interactive surface albedo parameterization. Five surface types are included in the model as follows: new snow, melting snow, bare ice, meltwater ponds, and open water. The albedo parameterization includes spectral discretization into four spectral intervals and dependence on the solar zenith angle, ratio of direct to diffuse radiation, and cloud properties [Ebert and Curry, 1993]. The four shortwave spectral intervals treated by the model are (1) 0.25-0.69 μm, (2) 0.69-1.19 μm, (3) 1.19-2.38 μm, and (4) 2.38-4.00 μm. The seasonal cycle of incoming radiative fluxes are specified in this study using the results of the slab model described by Curry and Ebert [1992].

The slab model was modified to include an ice thickness distribution after Björk [1992], where the sea ice is resolved into a limited number of ice classes that are characterized by their thicknesses and areas. The basic idea behind the ice thickness distribution model is that the sea ice is resolved into a limited number of ice classes, i = 0, n, that are characterized by their thicknesses H_i and fractional areas g_i. Fifteen ice thickness categories are used in this study, including a category for open water. The sum of the fractional ice areas and the fraction of open water must always equal unity. Each ice category has its own pond fraction p_i, thus the ice area can be separated into ponded and unp Ponded: Ice:

\[ a = \sum_{i=1}^{n} g_i p_i + g_i (1 - p_i) \]  

where a is the area of open water. A certain fraction of the ice is exported, maintaining a minimum open water area. New ice classes are continuously created by freezing of water in leads, with classes of similar thicknesses then merged to the prevent the number of ice classes from becoming excessive. Thin ice categories participate in the ridging process, transferring some ice to the thickest category; otherwise, each class of ice evolves thermodynamically, independently from the others. The sea ice is further divided into first-year (FY) and multiyear (MY) ice. An ice category is described by surface characteristics (area, ice thickness, snow cover, pond fraction and depth, surface temperature, and age) and interior ice properties (ice temperature, salt content, and brine pockets). The spatial mean thickness \( \bar{H} \) of the ice sheet is given by

\[ \bar{H} = \sum_{i=1}^{n} H_i g_i \]  

The spatial average of other quantities is similarly weighted by the ice thickness distribution \( g_i \). The model configuration is described schematically in Figure 1.

The sea ice thermodynamics included in the slab model (see Ebert and Curry, 1993 for more details) were modified for the ice thickness distribution model using data described by Grenfell and Maykut [1977] and Grenfell [1983]. For the purpose of specifying the optical properties, nonmelting FY ice is assumed to be composed of blue ice and MY ice is assumed to be composed of white ice. The top 5 cm of bare MY ice is assumed to consist of a granular layer in which all of the surface melting takes place. The discussion below focuses on aspects of the model that directly relate to the disposition of shortwave radiation in the sea ice and upper ocean and that differ from the treatment described by Ebert and Curry [1993].

2.1 Surface Albedo

We adopt the surface albedo parameterization described by Ebert and Curry [1993], with modifications for bare ice (Table 1). Separate spectral albedos are determined for FY and MY ice, following Grenfell and Maykut [1977].

As pointed out by Ebert and Curry [1993], melt ponds have a substantial influence on the surface albedo and thus on the modeled sea ice characteristics. Several changes to the melt pond parameterization have been made in the ice thickness distribution model. Ponds form when the ice surface temperature is greater than the melting temperature of sea ice \( T_m \), which depends on salinity [Neumann and Pierson, 1966]. Salinity is parameterized following Ebert and Curry [1993], to be constant with depth in the ice, the layer salinity value is determined by the depth (and thus age) of the ice. As in the slab model, the pond fraction \( p \) is parameterized to have a maximum value \( p_{max} \) at the beginning of the ice melt season and decrease to its minimum value \( p_{min} \) over the course of 30 days. The maximum and minimum pond fractions are 0.25 and 0.10, respectively, for MY ice, which are consistent with observations described by Maykut [1986]. We introduce values of maximum and minimum pond fractions of 0.90 and 0.50, respectively, for the smoother FY ice, based on the observations of Perovich and Maykut [1990] and J. Maslanik (personal communication, 1994). The pond depth is limited to half of the ice thickness; any further meltwater is allowed to run off.

The albedo of pond-covered ice is a maximum in the visible region (band 1). Since water is mostly transparent in this region, the albedo of the pond depends mainly on the characteristics of the underlying ice. Modeling the albedo of ponded ice is an extremely complex problem. Multiple reflections occur between the water and the underlying ice surface and melt
pond morphology indicate that three-dimensional radiative transfer effects are likely to be important. Therefore, following Ebert and Curry [1993], we determine spectral albedos over ponded ice using the observations of Grenfell and Maykut [1977]. Since Ebert and Curry [1993] did not consider ponds over first-year or blue ice, we have parameterized the albedo in the first three spectral bands over ponded first-year ice to be half the albedo value determined for the multiyear ice; the albedo in the fourth spectral band is assumed to be equal for both first-year and multiyear ice (Table 1).

2.2 Shortwave Attenuation

The model includes parameterizations for the disposition of the incoming surface shortwave radiation within the ice using data described by Grenfell and Maykut [1977] and Grenfell [1983]. The extinction coefficient is parameterized as a function of surface characteristics and ice age. The extinction of solar radiation in water, snow, and ice increases with wavelength for the four spectral intervals treated by the model: (1) 0.25-0.69 μm, (2) 0.69-1.19 μm, (3) 1.19-2.38 μm, and (4) 2.38-4.00 μm. Spectral extinction coefficients $k_i$ are specified in Table 2 for the various surface and ice types. At the longer wavelengths (bands 3 and 4) the extinction coefficient is so great (>500 m$^{-1}$) that absorption is assumed to take place immediately at the surface. This parameterization includes absorption by snow and ice and includes enhanced absorption by the 10-cm surface layer of MY ice. This surface layer is composed of an upper 5-cm granular layer overlying a 5-cm transition layer, which separates the granular layer from the ice interior [Grenfell and Maykut, 1977]. The extinction coefficients for these layers are shown in Table 2. Spectral extinction coefficients $k_i$ for the ice are functions of ice age (FY or MY) and the amount of brine trapped in the sea ice, since increasing the brine volume makes the ice more transparent.

### Table 1. Spectral Albedos for Ice Types

<table>
<thead>
<tr>
<th>Surface</th>
<th>Band 1, 0.25 - 0.69 μm</th>
<th>Band 2, 0.69 - 1.19 μm</th>
<th>Band 3, 1.19 - 2.38 μm</th>
<th>Band 4, 2.38 - 4.00 μm</th>
</tr>
</thead>
<tbody>
<tr>
<td>First year (FY) ice</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>$H_i &lt; 1$ m</td>
<td>0.760 + 0.140 ln $H_i$</td>
<td>0.247 + 0.029 ln $H_i$</td>
<td>0.055</td>
<td>0.036</td>
</tr>
<tr>
<td>$H_i \geq 1$ m</td>
<td>0.760</td>
<td>0.247</td>
<td>0.055</td>
<td>0.036</td>
</tr>
<tr>
<td>Multiyear (MY) ice</td>
<td>0.778</td>
<td>0.443</td>
<td>0.055</td>
<td>0.036</td>
</tr>
<tr>
<td>FY ice under pond</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>MY ice under pond</td>
<td>0.15 + exp(-8.1$h_p$-0.47)</td>
<td>0.05 + exp(-31.8$h_p$-0.94)</td>
<td>0.033 + exp(-2.6$h_p$-3.82)</td>
<td>0.030</td>
</tr>
</tbody>
</table>

$H_i$ is ice depth in meters. The band 1, 2, and 3 values for FY ice under pond are half of the values of MY ice under pond.
Table 2. Spectral Extinction Coefficients for Snow and Ice

<table>
<thead>
<tr>
<th>Surface Type</th>
<th>Band 1, 0.25 - 0.69 µm</th>
<th>Band 2, 0.69 - 1.19 µm</th>
</tr>
</thead>
<tbody>
<tr>
<td>Dry snow</td>
<td>19.6</td>
<td>196</td>
</tr>
<tr>
<td>Melting snow</td>
<td>10.7</td>
<td>118</td>
</tr>
<tr>
<td>Blue ice (FY)</td>
<td>1.8</td>
<td>19.6</td>
</tr>
<tr>
<td>White ice (MY) 0-5 cm</td>
<td>4.67</td>
<td>32.0</td>
</tr>
<tr>
<td>White ice (MY) 5-10 cm</td>
<td>2.0</td>
<td>14.0</td>
</tr>
<tr>
<td>White ice (MY) &gt;10 cm</td>
<td>1.4</td>
<td>17.6</td>
</tr>
</tbody>
</table>

Coefficient units are m\(^{-1}\). Values are estimated from Grenfell and Maykut [1977].

To examine the disposition of the incoming solar flux \(F_{SW}\), we calculate an incoming solar flux \(F\) for each spectral interval \(j\), whereby \(F_j = \omega_j F_{SW}\) and \(\omega_j\) is the relative spectral weight for the interval. The total amount of reflected radiation for a given ice thickness category is therefore

\[
F_{\text{refl}} = \sum_{j=1}^{4} F_j \alpha_j,
\]

where \(\alpha_j\) is the spectral albedo associated with the ice and surface type. The amount of reflected radiation averaged over the ice thickness distribution can be determined using (2).

The amount of shortwave absorbed in the snow cover, \(F_{\text{abs}}\), for each ice thickness category is calculated according to

\[
F_{\text{abs}} = \sum_{j=1}^{4} F_j (1 - \alpha_j) [1 - \exp(-\kappa_j h_j)]
\]

where \(\alpha_j\) is the spectral albedo of snow, \(\kappa_j\) is the spectral extinction coefficient of snow, and \(h_j\) is the snow depth. For snow-covered FY or MY ice the fraction of radiation transmitted through the ice to the mixed layer is

\[
F_p = \sum_{j=1}^{4} F_j (1 - \alpha_j) \exp(-\kappa_j h_j) \exp(-\kappa_j h_i)
\]

where \(\kappa_j\) is the spectral extinction coefficient of the ice and \(h_i\) is the ice depth. The remainder of the shortwave stays in the ice interior \(F_{\text{ice}}\):

\[
F_{\text{ice}} = F_{SW} - F_{\text{abs}} - F_{\text{refl}} - F_p.
\]

For bare ice the equations are similar, but there is enhanced spectral absorption in a 10-cm surface layer atop bare MY ice, producing enhanced melting near the surface. The bulk extinction coefficient decreases sharply in the top 5 cm of MY ice (the granular layer) and then smoothly in the 5 cm transition zone to slowly decreasing interior ice values [Grenfell and Maykut, 1977]. We use a spectrally varying extinction coefficient in the granular and transition layers, as shown in Table 2. The shortwave absorbed in the 10-cm surface layer is represented by the sum of the SW absorbed in the upper 5-cm granular layer and the SW absorbed in the transition layer

\[
F_{10,\text{cm}} = \sum_{j=1}^{4} F_j (1 - \alpha_j) [1 - \exp(-\kappa_{jz} z_j) \exp(-\kappa_{jz} z_i)]
\]

where \(\alpha_j\) is the spectral albedo of the ice; \(\kappa_{jz}\) and \(k_{jz}\) are the spectral extinction coefficients of the granular and transition layers, respectively, and the thickness of the granular and transition layers are represented by \(z_j\) and \(z_i\) respectively. This allows the calculation of a spectral value of \(i_s\), where \(i_s\) is the ratio of the unattenuated shortwave at 10 cm to the incident shortwave to the surface of the MY ice

\[
i_s = \exp(-\kappa_{jz} z_j) \exp(-\kappa_{jz} z_i)
\]

The radiation transmitted to the ocean mixed layer for MY ice can then be calculated as

\[
F_{\text{ul}} = \sum_{j=1}^{4} F_j (1 - \alpha_j) i_s \exp\{-\kappa_{jz} (H_i + z_j + z_i)\}.
\]

The remainder of the radiation is absorbed in the ice interior:

\[
F_{\text{ice}} = F_{SW} - F_{\text{abs}} - F_{\text{refl}} - F_p.
\]

The disposition of shortwave radiation in ponded ice is complex. The absorption of shortwave radiation in the melt ponds is parameterized using the model results described by Jin et al. [1994]. The amount of radiation absorbed in the pond under clear-sky conditions \(a_{\text{clair}}\) is parameterized using the model results of Jin et al. [1994] to be

\[
a_{\text{clair}} = -0.001 + 0.963 h_p^{0.13}
\]

and the corresponding value, \(a_{\text{clad}}\), for overcast conditions is

\[
a_{\text{clad}} = 0.012 + 0.932 h_p^{0.23}.
\]

For a given cloud fraction \(N\) the pond absorption is determined to be

\[
a = (1 - N) a_{\text{clair}} + N a_{\text{clad}}
\]

It is assumed that all incident (nonreflected) radiation in bands 2-4 is absorbed within the pond; only radiation in the first band penetrates beneath the pond. Using this assumption, the pond absorptivity in band 1 \(a_1\) is estimated to be

\[
a_1 = (a - 0.48)/0.52
\]

The amount of radiation transmitted through the pond and into the ice \(F_{up}\) is given by

\[
F_{up} = F_{1} (1 - \alpha_1 - a_1)
\]

where \(\alpha_1\) is the spectral albedo of ponded ice for band 1. The SW penetrating to the ice interior under ponded ice is

\[
F_p = F_{SW} - F_{\text{refl}} - F_{\text{abs}}
\]

where the reflected SW, \(F_{\text{refl}}\), is calculated according to (3) using the spectral pond albedo.

The radiation that reaches the ice interior is absorbed exponentially with depth in each of the model layers according to Beer's law. The SW absorbed in the top layer of FY and MY ice contributes to increasing the surface temperature. The portion of \(F_{ic}\) or \(F_i\) absorbed in each layer of unpended and ponded ice, \(\Delta F_{ic}^{\text{ic}}\) and \(\Delta F_i^{\text{ic}}\) are used to calculate the change in interior ice temperature in each layer

\[
\frac{dT}{dt} = \sum_{i=1}^{4} \left[ (1 - \rho_i) \frac{\Delta F_i^{\text{ic}}}{(\rho_i)_{ic}} + \rho_i \frac{\Delta F_i^{\text{ic}}}{(\rho_i)_{ic}} \right]
\]

Here \((\rho_i)_{ic}\) is the specific heat of sea ice, which is the sum of the heat required to warm the ice and the brine plus that needed for melting in the brine pockets [e.g., Maykut and Untersteiner, 1971]. If this change in temperature causes the interior ice temperature in any layer to go above freezing, this excess heat \(\Delta H_u = (\rho_i)_{ic} (T_i - T_F) dw\) goes into the latent heat of the brine pockets and the interior ice temperature remains at the freezing temperature \(T_F\). The latent heat released when the melt ponds freeze over is also added to this reservoir. In the fall, when the ice temperature drops below freezing, this heat is used to retard cooling. During the summer melt season the radiation absorbed in the upper 10 cm of ice just below the melt pond is used to heat the pond.
If the net surface energy balance over a lead is positive, all the energy goes into the lead, whereas lateral melting. The fraction of solar radiation absorbed between the water surface and the base of an ice thickness category follows Ebert and Curry [1993]. This absorbed radiation leads to lateral melting of the ice, which is assumed to produce an area change in each ice thickness category according to

$$\frac{\partial g_i}{\partial t} = g_i a \frac{F_{\text{sw}}}{(1 - a)} q_i (H_i + q_i h_i)$$

(18)

where $a$ is the lead fraction and $q_i$ and $h_i$ are the volumetric heat of fusion of ice and snow, respectively. The solar flux under each ice category is now

$$SW_0 = \frac{a g_i}{(1 - a)} M F_{\text{sw}} + g_i \left[ (1 - p_i) F_{\text{un}} + p_i F_{\text{up}} \right]$$

(19)

where the first term is the solar flux entering the part of the lead adjacent to the $i$th category, penetrating below the ice, and is used to heat the water below the ice. The second term is the solar flux penetrating directly through the ice.

### 3. Results

Fifteen categories of ice thickness plus open water are used in this simulation. Each category covers a variable fractional area which is used to calculate an area-weighted average of the surface, lead, and pond conditions. The model is integrated for 25 years, producing an annually averaged ice thickness of 3.15 m. At the time of maximum melt the number of ice thickness categories decreases to 10, as ridging ceases and the thinnest ice categories have melted away.

Figure 2 shows the seasonal cycle of incoming shortwave radiation at the surface (following Curry and Ebert [1992]) and the area-averaged lead fraction, pond fraction, and pond depth, as determined by the model. Incident SW radiation begins to warm the surface in early March and increases to its maximum of 330 W/m$^2$ by mid-June. This is well before the lead or pond fractions have reached their maxima. The lead fraction increases gradually from May through early September and decreases sharply in early September. The ice thickness distribution model allows for a rapid autumnal freezing with a sharp drop in the lead fraction and the creation of thin ice, whereas the slab model [Ebert and Curry, 1993] produces a gradual decrease in the lead fraction in the fall months. The pond fraction increases abruptly in early July and decreases gradually over a period of about a month until freeze-up in mid-August. The presence of a small pond fraction in late-June arises from the melting of the thinnest ice. These ponds disappear when the thinnest ice melts away, the ponds returning when the thicker ice begins to melt. The pond depth reaches its maximum about 1 month after the peak in pond fraction.

The partitioning of shortwave radiation between reflection, surface absorption, interior absorption, and transmission de-
pends mainly on the lead and pond fractions, with the thicker ice categories having the smaller pond fractions. To illustrate the seasonal evolution of this partitioning, we examine the following three different regimes: (1) before the onset of the melt season, with snow cover still present on the thicker ice categories, and the incoming SW near its maximum; (2) early in the melt season, when melt ponds cover a large horizontal area but are quite shallow; and (3) late in the melt season, after considerable surface melting has taken place and the ponds are at their deepest, but their horizontal extent is reduced (as shown in Figures 2c and 2d).

Figure 3 shows the variation of surface reflection (REFL), solar absorption in the snow layer (ABSS), absorption in the top 10 cm of the ice (ABST), absorption of solar radiation into the ice interior (ABSI), absorption in melt ponds (ABSP), and transmission of solar radiation through the ice into the ocean (TRAN), as a function of ice thickness for these three regimes. The relative units of radiation shown in Figure 3 are the weighted-area averages of the solar energy incident on ponded and pond-free areas. At the beginning of the melt season there are 15 ice categories. The thinnest FY categories are snowfree, which allows absorption of solar radiation in the surface layer and the ice interior (Figure 3a). The amount of ABSI decreases with ice thickness, due to the logarithmic dependance of albedo on ice thickness for FY ice that is less than 1 m (Table 1). For ice less than 1.6 m some shortwave is transmitted into the ocean mixed layer. The presence of a snow cover on the thicker categories causes most of the solar radiation to be reflected by the surface, with the remainder being absorbed in the snow layer (Figure 3a). In Figure 3b the snow has melted on all ice thicknesses, melt ponds have formed, and the number of ice categories has decreased to 13. The area-averaged pond fraction for this regime is 0.19, and the area-averaged pond depth is 0.10 m. The values of the pond fraction and pond depth corresponding to Figure 3b are shown in Figure 4. A significant amount of solar energy is absorbed in the deeper, more extensive, melt ponds, especially for the first three categories, which are still FY ice. The thicker (MY) ice has enhanced absorption in the 10-cm granular layer and less absorption in the shallower, less extensive melt ponds. Near the end of the melt season (Figure 3c) the thinnest ice categories have melted away, leaving 11 ice categories and a large area of open water and leading to a correspondingly large amount of transmitted radiation. The disappearance of all but the thickest FY ice has decreased the amount of SW absorbed in the ice interior, due to the increase in absorption in the 10-cm granular layer. Although transmission through leads is large, the incoming SW is decreasing rapidly, and the leads will refreeze in about 30 days. The pond fraction has decreased (0.11), and the average pond depth has increased (0.40 m) since the early melt season regime of Figure 3b, increasing the ice area that can absorb radiation at the surface and decreasing the amount of penetrating solar radiation.

The contrast in solar reflection, surface absorption, and interior absorption between unpended and ponded ice is highlighted in Figure 3. As expected, more SW is reflected from unpended ice than from ponded ice. About 35% of the incoming SW is absorbed in the top 10 cm for unpended ice (Figure 5a). The presence of melt ponds allows about 50% more solar energy to penetrate into the ice interior. The majority of the incoming SW is absorbed in the melt ponds. More SW is absorbed in ponds on FY ice (70%) than on ponded MY ice (55%), due to the larger fraction of deeper ponds (Figures 4 and 5b). Slightly more SW is transmitted to the ocean through ponded ice. Maykut and Grenfell [1975] measured spectral transmission beneath first-year sea ice with various surface conditions. For an ice thickness of 1.85 m, ice covered with

Figure 3. Variation of surface reflection (REFL), solar absorption in the snow layer (ABSS), absorption in the upper 10 cm of the ice (ABST), absorption in melt ponds (ABSP), penetration of solar radiation into the ice interior (ABSI), and transmission of solar radiation through the ice (TRAN), as a function of ice thickness for the following three regimes: (a) before the onset of the melt season (mid June), (b) at the start of the melt season (mid July), and (c) at the end of the melt season (mid August). The premelt snow depth is shown as a function of ice thickness at the top of Figure 3a.
melt ponds transmitted 3.3 times as much energy as the adjacent white ice, which in turn transmitted 10 times more energy than ice covered by 25 cm of snow. The model results, of course, depend upon the assumptions about melt pond area and depth, while we believe that we have chosen plausible melt pond characteristics in view of the available observations, additional observations are required to validate and improve the model.

The disposition of solar radiation in leads is shown in Figure 6. Here LATL represents the proportion of solar energy used for lateral melting at the edge of ice floes. Following Parkinson and Washington [1979], it is assumed that all incident shortwave goes into heating the water column below the lead if the net flux at the surface of the lead is negative. If the net flux is positive, it is assumed that the SW penetrating down to the depth of the ice goes into lateral melting, with the remainder going to heat the mixed layer. This assumption obviously influences our model results, and unfortunately, there is insufficient data at present to improve on this assumption. Using this assumption for 3 m ice, about 30% of the solar input is transmitted through the lead to the mixed layer beneath the ice, where it contributes to warming of the water and basal melting of the ice.

To interpret the results shown in Figures 4-6 in the context of the distribution of ice thickness, the time evolution of the ice thickness distribution must be examined. Figure 7 shows the cumulative ice thickness distributions \( G(H) \) for the middle of months June, July, and August (dates corresponding to Figure 3), where

\[
G(H) = \sum_{i=1}^{n} g_i
\]

and \( i \) is the number of the ice category, \( H_i \) is the ice thickness, and \( g_i \) is the fraction of the total area occupied by the ice class \( j \). The ice thickness distribution for mid-June (Figure 7a) consists of a ridged category, with an average ice thickness of 5.8 m, covering an area of about 10%. There are four ice classes with a thickness less than 1.7 m, covering a very small area (1.6%), and finally, the open water class, covering an area of 2%. By July (Figure 7b) the open water coverage has increased to 6% and two of the thin ice categories have melted away. The multiyear and ridging categories are thinner than in June but cover similar areas. In August (Figure 7c) the lead fraction has increased to 8%, almost to its maximum, and two more thin ice categories have melted away, leaving 11 categories.

Observations of the summertime ice thickness distribution are reported by Bourke and Garrett [1987] and McLaren [1989]. Theoretical calculations are described by Thorndike et al.
during August for ice with thickness 0-50 cm. While these observations are not inconsistent with the ice thickness distribution shown in Figure 7, the observational data as presented in these papers are of limited utility in determining $h_c$. The second feature is that the category with the maximum ice thickness, the ridged ice category, has a thickness in August of about 5.6 m. McLaren [1989] and Bourke and Garrett [1987] each report approximately 20% areal coverage of ice thicker than 5 m during summer, while our model averages about 10% areal coverage of ice thicker than 5 m over summer. The amount of ridged ice will certainly depend on the deformation history and thus is expected to show substantial variability. For the disposition of shortwave radiation in sea ice and the upper ocean, details of the thickness distribution beyond 5 m are relatively unimportant, since no shortwave radiation will penetrate through ice this thick into the upper ocean.

Figure 8 shows the time evolution of the area-averaged disposition of solar radiation during the sunlit portion of the year for the modeled evolution of the ice thickness distribution. Before the onset of the melt season, reflection and absorption by snow are the dominant processes. There is also a small amount of reflection and absorption in the ice interior from the thin, relatively snow-free ($h_s < 10$ cm) ice categories. In early June the thin ice categories begin to melt away, producing areas of bare ice for a short time and increasing the area of open water through which radiation is transmitted and lateral melting can occur. In early July, snow begins to disappear from the thicker ice categories, allowing absorption and penetration of solar radiation in the ponds and ice. By the end of July all ice categories are in a state of melting, with variable melt pond coverage contributing to a reduction in surface reflection and an increase in shortwave absorption. After the melt season concludes in late August, surface melting ceases on MY ice and snow begins to accumulate. Penetration of SW into the ice interior continues as the thin snow cover transmits a large fraction of the incident radiation. The leads freeze over in late September, and the reflection and absorption of SW radiation by snow cover is again dominant. The timing and duration of the melt season are within the range of observations (summarized by Curry et al. [1993]).

The modeled annual cycle of shortwave radiation transmitted through the ice into the upper ocean is shown in Figure 9 for both the ice thickness distribution model ($n=15$) and the slab version of the model ($n=1$) described by Ebert and Curry [1993]. The annually averaged, area-weighted ice thickness from the distribution model was similar to the annually averaged ice thickness of the slab model as follows: 3.15 m versus 2.8 m, respectively, and the summertime growth of open water area was nearly identical for the two models. Including an ice thickness distribution increases the summertime peak value and the annual solar flux entering the ocean by a factor of 30% when compared with the slab model ($n=1$), in spite of the lower average ice thickness for the slab model. The sharp decrease in early September is an artifact of the assumption that when the net flux at the surface of the lead becomes negative, all of the SW entering the lead cools the water column and is not combined with that penetrating the ice. The value of the transmitted SW increases sharply when the leads freeze over, creating a 10% area of very thin, snow-free ice. It is extremely difficult to directly measure the solar radiation reaching the ocean beneath the sea ice. Using observed values of incident solar radiation and ice thickness, Perovich and Maykut [1990] calculated a transmitted SW flux of ~20 W/m² through ice thinning from

![Figure 7. Cumulative ice thickness distribution $G(H)$ versus ice thickness for the three regimes in Figure 3. Here $H$ is the spatial mean thickness of the ice.](image-url)
Figure 8. Time evolution of the area-averaged shortwave (SW) components during the sunlit portion of the year, normalized by the downward solar flux, including reflection from snow (REFLS) and ice (REFLI), ponds (REFLP), and leads (REFLL).

1.5 to 0.6 m during July-August 1982 in Mould Bay. Fichefet and Gaspar [1988] calculated a SW flux exceeding 5 W/m² for ice thinner than about 30 cm and only 0.1 to 0.2 W/m² for 3-m thick ice.

Examination of the solid curve in Figure 9 shows that the peak value of shortwave penetration into the upper ocean occurs near the end of July, approximately 6 weeks past the peak value of solar insolation (Figure 2a). The displacement of the peak value of transmission from the peak value of insolation depends on the relative timing of the maximum and minimum ice thickness, the maximum downward SW flux, and maxima in the pond and lead fractions. The sequence of these events is shown in Figure 10. The melt season begins with the ice thickness at its maximum, with the incoming SW radiation increasing until mid-June. Ponds form on the surface, reaching their areal maximum in mid-July. As time progresses, the ponds deepen and decrease in extent, the ice thins due to surface and basal melting and recedes due to lateral melting, and SW transmission to the ocean continues to increase. Fifteen days after the peak in pond extent, the maximum amount of SW is transmitted to the ocean, while the lead fraction is only half its maximum value. At the end of the melt season, when the lead fraction is at a maximum, the ice is also at its thinnest, but the incoming SW radiation is one sixth of its maximum value. It is the displacement of these maxima (shown in Figure 10) that helps maintain the stability of the ice pack, since the surface...
features that increase the transmission of radiation through the ice into the upper ocean occur away from the peak insolation.

The annual totals of the terms related to the disposition of shortwave radiation and percent of total solar input are given in Table 3. The reflected component is by far the largest, accounting for 69% of the total solar energy. Snow cover absorbs 15% of the incoming solar flux, 12% is absorbed by the ice, and about 4% is transmitted to the ocean mixed layer through leads and thin ice. In the absence of an oceanic heat source it is this transmitted solar radiation that provides the energy for basal melting. Approximately two thirds of the transmitted solar radiation is transmitted directly through open leads, with the other one third being transmitted through thin ice. A decrease in ice thickness, due perhaps to an increased net flux at the surface, would lead to a greater amount of solar energy being transmitted through the ice and leads (provided that the lead fraction did not also change), leading to a positive feedback [Ebert and Curry, 1993].

4. Conclusions

An ice thickness distribution model has been used to examine the disposition of the incident surface shortwave radiation in sea ice, snow, melt ponds, and leads. Depending on the characteristics of the snow and ice, this solar energy is either reflected, absorbed in the snow or ice, penetrates the ice or melt ponds, contributes to the lateral melting in leads, or is transmitted through the leads or thinner ice into the upper ocean. The presence of a thin snow cover or snow-free thinner ice categories allows transmission of radiation through the ice. Melt ponds increase the penetration of SW into the ice interior by 50%, depending on the fractional coverage and depth of ponds, and increase the transmission of incident SW into the upper ocean relative to unpended ice. Leads reflect only 8% of the incoming radiation, while the remainder goes into transmission and lateral melting, with the relative fraction of melting increasing with ice thickness. Of the 4% of the annual incident solar input which reaches the ocean mixed layer, approximately 66% is transmitted through leads and 34% through thin ice.

For the same average ice thickness the net transmission of solar radiation into the ocean is roughly 30% greater for a distribution of ice thicknesses than for a single ice slab. The peak value of shortwave penetration into the upper ocean occurs near the beginning of August, approximately 6 weeks past the maximum value of solar insolation, 2 weeks past the maximum pond fraction, and 6 weeks before the maximum lead fraction. The displacement of these maxima contribute to the stability of the Arctic ice pack and reduce the potential ice-albedo feedback associated with the sea ice [e.g., Curry et al., 1995].

Accurate modeling of the coupled atmosphere-ocean-sea ice system must account for the disposition of shortwave radiation in sea ice which includes leads and ponded ice. This study shows that the use of an ice thickness distribution leads to improved simulations of the surface albedo and SW flux transmitted to the ocean, both important quantities for climate models.

This model has presented an internally consistent and plausible shortwave radiation budget of the Arctic sea ice and upper ocean. However, major uncertainties in the present calculations include the relative amount of shortwave radiation entering leads that goes to lateral melting, the modeled melt pond characteristics, and the validity of the ice thickness distribution. The planned Surface Heat Budget of the Arctic Ocean (SHEBA) field experiment [SHEBA Science Working Group, 1994] should provide data that will form the foundation of improved understanding of the disposition of shortwave radiation in sea ice and the upper ocean.

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