A numerical study of melt ponds

Eric D. Skyllingstad\textsuperscript{1} and Clayton A. Paulson\textsuperscript{1}

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[1] High-resolution turbulence simulations are used to examine the importance of melt pond geometry in setting pond growth rates and albedo. Modeling the circulation of water in melt ponds using large-eddy simulation shows that both convective and wind-forced conditions generate well-mixed ponds, suggesting that stratification is not a significant factor in pond circulation. Simulations with a variety of pond shapes and sizes indicate that the basic ratio of sidewall area to bottom area, $R$, can be used to characterize melting rates for ponds with simple shapes. Ponds with large values of $R$ will generally melt more rapidly in the horizontal direction at the expense of bottom melting. Consequently, small and elongated ponds will have a relatively larger lateral growth rate in comparison with large, symmetric ponds, assuming minimal lateral flux of meltwater. Simulations also show that pond shape can affect the sidewall and bottom turbulence transfer rates. Ponds with large $R$ tend to have reduced transfer rates because of weaker circulations. A bulk pond model is developed on the basis of a rectangular geometry and an assumption of uniform mixing as suggested by the turbulence model and pond scaling using $R$. Comparison of the bulk model with results from the large-eddy simulation cases shows good agreement.


1. Introduction

[2] One of the main difficulties in modeling climate change is the accurate simulation of physical processes that change the radiative balance of the coupled atmosphere-ocean system. For example, processes that control ice and snow coverage directly affect the radiative budget by setting the amount of solar radiation reflected back into space by the Earth's surface. If climate warming causes a decrease in surface ice and snow coverage and the corresponding surface albedo, then the greater absorption of solar radiation could accelerate the increase in global temperatures. This nonlinear interaction is commonly referred to as the ice-albedo feedback.

[3] One of the primary motivations for the yearlong Surface Heat Budget of the Arctic Ocean (SHEBA) Project. Results from the SHEBA experiment indicated five distinct phases in total surface albedo dependent on specific processes [Perovich et al., 2002a, 2003]. For most of the year, surface albedo ranged between 0.9 and 0.8 because of the uniform ice and snow coverage. Beginning in late May, albedo slowly decreased to about 0.8–0.7 as the surface snow melted. Formation of melt ponds in mid-June caused a large decrease in average albedo from 0.7 to 0.5, although some regions remained pond free. Total albedo values then decreased steadily as the melt ponds evolved in July before reaching a minimum of $\sim$0.4 in early August. Freezing conditions returned to the site in mid-August, resulting in a rapid increase in albedo to wintertime values of $\sim$0.8 by early September [Perovich et al., 2002a].

[4] Overall, decreases in albedo measured during SHEBA were a strong function of melt pond area coverage and the albedo of the individual melt ponds. Early in the melt season, melt ponds were relatively shallow (0.1 m) and had albedo values of $\sim$0.3–0.4. As ponds increased in depth, the pond albedo decreased to values as low as 0.1 near the end of July. Leads also affected the average albedo, but only occupied about 5% of the surface until August when the first fall storms caused ice movement and a rapid increase in regional lead coverage to about 20% [Perovich et al., 2002b, 2003]. These observations are consistent with previous field programs showing that pond coverage is a critical element in setting the ice surface albedo [Grenfell and Maykut, 1977; Grenfell and Perovich, 1984; Barber and Yackel, 1999; Fetterer and Untersteiner, 1998; Hanesiak et al., 2001].

[5] From a climate modeling standpoint, observations indicate that accurate representation of the ice-albedo feedback is highly dependent on predicting the melt pond coverage and applying the correct pond albedo for each phase of the pond evolution. One way to accomplish this goal is to develop an empirical parameterization of albedo based on the heat and radiation fluxes observed during field experiments. However, this approach assumes that field
conditions are representative of the average Arctic ice conditions, which may not be accurate for changing climate.

A second method for representing pond albedo considers the physics of melt ponds and the underlying ice to construct an approximate model of ice surface processes [e.g., Taylor and Feltham, 2004]. Although this approach provides a sound foundation for modeling melt ponds and ice processes, it presents major obstacles that need to be overcome. One significant issue concerns estimating the areal extent of melt ponds and the individual pond depths. Ponds form much like water features on land; melting snow and ice form melt ponds in preexisting depressions in the ice, which are created by ice deformation and variations in snow depth. Melt ponds grow by collecting meltwater from the ice surface and by melting the surrounding ice and snow, gradually deepening until, in some cases, reaching the ocean beneath the ice. Fresh water also percolates through the ice in a manner similar to groundwater moving through soil, and can form a communication link between melt ponds and the underlying ocean-ice interface. The flux of fresh water from melt ponds to the upper ocean may have been a significant factor during the SHEBA experiment, where a 0.5 m fresh layer was observed beneath the ice late in the summer melt period [Note et al., 2003].

Using data from Arctic field experiments, Eicken et al. [2002] developed a four-stage evolution process for areal melt pond coverage during the summer season. In the first phase of the evolution process, melt ponds form as snow-melts, and meltwater accumulates in existing depressions and cracks. Stage two begins when the snow is completely melted, but meltwater is largely confined to the ice surface without significant percolation through the ice. Pressure head from the meltwater is high at the beginning of this phase, forcing water into the ice through pores and ice defects. Eventually, as the ice melts, pathways form allowing the fresh water ponds to drain through the ice and into the surrounding ocean. The third phase begins when the meltwater has reached equilibrium with the sea surface so that pressure head is less significant. Ice permeability during this stage is very high, causing brine pockets to drain through the ice and be replaced by much fresher meltwater. On first year ice, melt ponds also increase in size during this phase because of lateral and bottom ice melting as ponds warmed by the sun melt the surrounding ice. Ponds on multiyear ice, on the other hand, may remain elevated above the sea surface and actually shrink during this phase as water continues to drain through the ice [Fetterer and Untersteiner, 1998]. Given enough solar heating, ponds in this stage can melt through the ice cover with the ice surface disintegrating. The final stage in the melt pond evolution is fall freezeup when ice reforms on melt ponds effectively ending the surface melt season.

Eicken et al. [2004] found that movement of fresh water through the ice was strongly controlled by the ice age and the initial roughness of the ice at the beginning of the melt season. Similar behavior was noted by Barber and Yackel [1999], who showed that pond albedo was strongly controlled by the age of the ice. One consequence of this behavior is that changes in the life cycle of ice could have important feedbacks on the climate system if, for example, multiyear ice concentrations are reduced [Eicken et al., 2004].

Simulation of melt pond evolution in climate system models have mostly considered bulk ice models where melt ponds affect the ice surface albedo as a component of the average ice properties [Ebert and Curry, 1993; Bitz and Lipscomb, 1999]. Solar radiative transfer and thermodynamic properties of the ice system are included in these approaches; however, the effects of pond size are not directly simulated. Similarities between meltwater evolution and surface water over land prompted Eicken et al. [2004] to propose a melt pond model based on a hydrological approach. In their model, measured statistics of ice topography and permeability are used in a basic groundwater formulation that relates the surface elevation of melt ponds to the meltwater production rates. Using this model, Eicken et al. [2004] examined pond area coverage for two sites (SHEBA and a location near Barrow AK) having different ice age, permeability and corresponding pond coverage and albedo. They found that the model reproduced observed pond features and variability for both locations when forced with a typical range of ice topography, melting rates, and permeabilities. Although the model performed well given prescribed melting rates, it did not explicitly calculate changes in melt pond coverage produced by lateral and bottom melting. Pond coverage statistics were computed on the basis of the available fresh water production and input ice topography. Because the hydrological model depends on knowing the ice topography, changes in the melt pond width and depth need to be predicted as the summer melt season progresses. These changes are a direct element in the ice-albedo feedback process; as melt ponds absorb solar heat, they increase in size and depth, thereby decreasing the average ice albedo.

Lüthje et al. [2006] present a model that combines aspects of the layer approach described by Taylor and Feltham [2004] with the Eicken et al. [2002] method, yielding a spatial simulation of melt ponds. Their approach includes enhanced melting beneath melt ponds and simulates the budget of surface water as the ice pack melts. Melting rates are prescribed in this model without direct consideration of the radiation budget, but do depend on the pond depth and thus indirectly include changes in the pond albedo.

Neither past observational studies nor pond modeling experiments have explicitly included heat transport within ponds, or considered the direct effects of lateral and bottom melting on pond size and albedo. Consequently, the dependence of pond solar input on changing pond size, shape, and depth has not been directly studied. Our goal in this paper is to better understand how the transport of solar heat in melt ponds controls side and bottom melting, addressing a major issue in predicting the melt pond albedo feedback. We investigate the thermodynamic behavior of melt ponds using two modeling approaches. Our first approach addresses the heat budget of a melt pond using a large-eddy simulation (LES) model that can directly simulate the motion of water within a melt pond. Use of the LES model allows us to estimate the importance of wind-forcing and heat transport within the pond. However, because of the high computational cost of LES, we can only examine the heat transport processes over periods of a few hours.

To examine longer-term pond behavior, a less complicated bulk model is developed, which concentrates on the radiative heating within the pond. The bulk model simulates pond growth by treating the pond as well mixed and tracks the width, length, and depth of the pond as a function of
time. This method differs from previous bulk models by considering the pond size and applying the pond budget equation with a depth-dependent solar flux. The model provides a useful tool that can bridge the gap between expensive, high-resolution LES and single layer ice models as described by Taylor and Feltham [2004]. In both the LES and bulk models, we do not account for fluxes of heat and water from lateral transport through porous ice. These factors require a more complete ice model and surface process model, which are not of primary interest in this study.

[13] The paper begins with a description of the modeling approaches and pond configurations in section 2. Section 3 describes results from the LES and bulk pond model. Conclusions are presented in section 4.

2. Model Approaches

2.1. Large Eddy Simulation Pond Model

[14] One of the main uncertainties in modeling melt ponds is quantifying how solar heat is transferred from the pond interior to the bottom and sidewalls where melting occurs. Accurate estimates of the edge and bottom melting rates are crucial for calculating the role that ponds have in the ice-albedo feedback. For example, if most of the heat is used to melt the pond bottom, then the absorption of solar radiation in the pond will rise because of the deeper path of solar radiation, but the overall area of pond coverage will not increase significantly. In contrast, larger edge melting has the effect of increasing pond size, but the individual pond albedo will not decrease as rapidly. Clearly, both of these factors come into play in defining the overall albedo of melting sea ice, even when ponds are also changing in size because of lateral meltwater transport [e.g., Eicken et al., 2004]. To estimate the relative importance of side versus edge melting, we apply an LES model that directly simulates the motion of water throughout the pond and the transport of heat from the pond interior to the bottom and sides.

[15] The LES model is based on the filtered nonhydrostatic, Navier-Stokes equations with a subgrid-scale closure provided by Ducros et al. [1996]. Pond boundaries are simulated using a volume of fluid approach following Steppler et al. [2002] and Adcroft et al. [1997]. The equations of motion are defined using an enstrophy conserving scheme following tripodli [1992],

\[
\frac{\partial u_i}{\partial t} = \varepsilon_{ij} u_j \eta_k - \frac{\partial}{\partial x_i} KE - \frac{\partial}{\partial x_j} (u_j u_i) - \frac{\partial P}{\partial x_i} - \delta_{1g} \frac{g \rho'}{\rho_0} + \gamma \partial^2 u_i,
\]

where

\[
\langle u_i u_j \rangle = -K_{ij} \left( \frac{\partial u_i}{\partial x_j} + \frac{\partial u_j}{\partial x_i} \right), \quad P = \frac{\rho_0}{\rho} + \frac{2}{3} g^2,
\]

\[
\eta_k = \zeta_k + f_k, \quad \zeta_k = \varepsilon_{ikj} \frac{\partial u_j}{\partial x_i},
\]

\[
KE = \frac{1}{2} \bar{u}^2,
\]

\[u_i\] are the Cartesian velocity components, \(t\) is time, \(g\) is the acceleration of gravity, \(\delta_{1g}\) is the Kronecker delta, \(\varepsilon_{ijk}\) is the antisymmetric tensor, \(P\) is pressure, \(\rho'\) is the perturbation density, \(\rho_0\) is a constant average water density, \(K_m\) is the subgrid-scale eddy viscosity, \(q^2\) is the subgrid-scale turbulent kinetic energy, \(f_k\) are components of the Coriolis term, and \(\gamma \partial^2 u_i\) is a twelfth-order filter with filter coefficient \(\gamma = 0.07\) to remove a \(2\Delta x\) artifact of the differencing scheme [see Denbo and Skyllingstad, 1995]. Angle brackets, \(\langle \rangle\), denote subgrid-scale fluxes which are parameterized using Ducros et al. [1996]. A complete description of the LES model is given by Skyllingstad et al. [2005].

[16] Wind stress at the top of the pond is transferred directly as a prescribed momentum flux, \(\tau_{w,0}\), whereas friction along the pond edges and bottom is set using a similarity approach as presented by Skyllingstad et al. [2005]. For the ice edge, friction velocity, \(u_* = \sqrt{\tau_{w,0}}\), is defined as

\[
\langle u_i u_j \rangle = C_D \Delta u_i |u_i| = u_*^2,
\]

where

\[
C_D = \left[ \frac{K}{\ln(\frac{d}{z_o})} \right]^2, \quad \Delta u_i = \sqrt{\left( \bar{u}_i^2 + \bar{u}_j^2 \right)},
\]

\(z_o\) is the ice aerodynamic roughness set to 0.002 m, \(\delta x\) is set to 1/2\(\Delta x\) denoting the distance from the ice edge to the seawater grid cell center, and \(i, j\) represent the indices of the seawater grid cell adjacent to the ice edge. We use a roughness value that is less than the estimate \((z_o = 0.005 \text{ m})\) made by McPhee [2002] using measurements of turbulence beneath sea ice, primarily to avoid problems with grid spacing approaching \(z_o\). However, test simulations indicate that pond circulation and melting rates are not overly sensitive to the value of \(z_o\).

[17] Temperature, \(T\), and salinity, \(S\), are simulated using

\[
\frac{\partial T}{\partial t} + u_i \frac{\partial T}{\partial x_i} = \frac{\partial F_T}{\partial x_i} - \frac{\partial (u_i T)}{\partial x_i},
\]

\[
\frac{\partial S}{\partial t} + u_i \frac{\partial S}{\partial x_i} = -\frac{\partial (u_i S)}{\partial x_i},
\]

where

\[
\langle u_i T \rangle = K_{T} \frac{\partial T}{\partial x_i}, \quad \langle u_i S \rangle = K_{S} \frac{\partial S}{\partial x_i},
\]

[18] \(K_T\) is the eddy diffusivity from Ducros et al. [1996] and \(F_T\) is the solar radiative flux. Melting along the pond bottom and sidewalls is parameterized using transfer coefficients developed by McPhee et al. [1987] that account for the differing diffusivities of salt and heat. Ice melting rate is calculated using

\[
\langle u_i T_o \rangle = w_{ice} Q_t,
\]

\[
\langle u_i S_o \rangle = w_{ice} (S_w - S_i),
\]
where \( w_{\text{ice}} = \rho_i / \rho_s \), \( d \) represents a velocity of the ice surface associated with ice growth or melting, \( \langle u_i T_i \rangle \) and \( \langle u_i S_i \rangle \) represent the subgrid heat and salinity boundary fluxes into the pond near the ice edge, respectively, \( d \) is the ice melting or growth rate, \( \rho_i = 920 \text{ kg m}^{-3} \) is the ice density, and \( Q_L = L / c_p \), where \( L = 3.34 \times 10^3 \text{ J kg}^{-1} \) is the latent heat of fusion and \( c_p = 4000 \text{ J (Kg}^{-1}\text{ °C}^{-1}) \) is the specific heat of water. Scaling equations (4) and (5) with \( u^* \) as defined above, and integrating vertically results in two nondimensional functions

\[
\Phi_T = \frac{T(x_i) - T_w}{(w_{\text{ice}} Q_L) / u^*} \quad \text{(6)}
\]

\[
\Phi_S = \frac{S(x_i) - S_w}{w_{\text{ice}} (S_w - S_i) / u^*} \quad \text{(7)}
\]

where \( T(x_i) \) and \( S(x_i) \) are the temperature and salinity at the nearest grid point to the pond bottom or edge.

Equations (6) and (7) can be combined and simplified by replacing the wall temperature with the freezing temperature at \( S_w \), or \( T_w = -m S_w \), where \( m = -0.054 \) yielding

\[
(1 + m) S_w^2 + \left[ T(x_i) - m S_i + \frac{\Phi_T Q_L}{\Phi_S} \right] S_w - \left[ T(x_i) S_i + \frac{\Phi_T Q_L}{\Phi_S} S(x_i) \right] = 0.
\]

For most ponds, the salinity is low (< 4 psu [Eicken et al., 2002]), consequently the salinity of the pond water does not have a large influence on the edge and bottom melting rates.

A schematic of a typical melt pond heat budget is shown in Figure 1. Solar flux, \( F_s \), is parameterized using a radiative transfer equation developed using observations from fresh water leads taken between 17 June and 4 August during the SHEBA experiment [Pegau, 2002]. Radiative fluxes are calculated using,

\[
F_r(z) = P_m F_m (1 - e^{-K_m z}) \quad \text{(9)}
\]

where \( P_m \) is the proportion of shortwave energy in the band \( m \), \( F_m \) is the net shortwave radiation at the sea surface, \( K_m \) is the diffuse extinction coefficient, and \( z \) is the depth below the surface. Information on the band characteristics is provided in Table 1. The use of four exponents is driven by the need to resolve radiative heating in a very shallow surface layer and to capture the rapid change with wavelength in the absorption coefficient of water from less than 0.1 to greater than 300 m\(^{-1}\). Accurate estimation of the near infrared absorption requires expanding the number of exponents in regions near the surface where absorption coefficients change rapidly.

Table 1. Band Characteristics Used to Determine the Shortwave Radiation Absorbed in a Freshwater Layer

<table>
<thead>
<tr>
<th>Wavelength</th>
<th>Range</th>
<th>m = 1</th>
<th>m = 2</th>
<th>m = 3</th>
<th>m = 4</th>
</tr>
</thead>
<tbody>
<tr>
<td>( P_m )</td>
<td></td>
<td>0.481</td>
<td>0.194</td>
<td>0.123</td>
<td>0.202</td>
</tr>
<tr>
<td>( K_m )</td>
<td></td>
<td>0.18</td>
<td>3.25</td>
<td>27.5</td>
<td>300</td>
</tr>
</tbody>
</table>

*\( P \) is a function of cloud conditions, and \( K \) is a function of material in the water.
Using equations (9) and (10), pond albedo can be calculated as a function of depth by integrating the downwelling and upwelling radiation,

\[ Q_r = \int_{h} F_r(z) \, dz + \int_{h} F_r(z) \, dz, \quad (11) \]

and accounting for the radiation transmitted through the pond bottom, \((1 - a_b) F_r(z_b)\). A plot of the predicted albedo is presented in Figure 2 for three different pond bottom albedo values. In even the most shallow pond, which in this case is 0.004 m, the absorption of the near-infrared portion of the solar spectrum causes a ~0.1 reduction in the albedo as predicted by equations (9) and (10). For ponds with depths of ~0.4 m, albedo ranges from 0.2–0.3, which is in agreement for first year pond albedo measured by Hanesiak et al. [2001] and for “light” pond albedo measured by Barber and Yackel [1999]. Measured albedo values for shallow ponds (0.1 m) reported by Fetterer and Untersteiner [1998] ranged from 0.37 to 0.47 depending on the pond bottom albedo, also supporting the modeled values shown in Figure 2.

By using the one-dimensional empirical formula presented in equations (9) and (10), we ignore the effects of sunlight that enters the pond at an angle and interacts with the pond sidewall. For small ponds, this could be significant, but will in general be much smaller than the integrated sunlight defined by the pond surface area. We also assume that the pond radiative properties, cloud conditions and overall weather are similar to those of the fresh water lead that was used to formulate equations (9) and (10).

Surface fluxes of longwave radiation along with sensible and latent heating, \(F_r\), are prescribed by assigning a subgrid-scale flux boundary term at surface grid points, via \(\langle u_i T \rangle_{sf}\), in equation (2). For the idealized simulations presented here, \(F_r\) is held constant with a value of ~30 W m\(^{-2}\), representing a typical summer average value, for example, as measured during SHEBA using bulk formula for sensible and latent heat and measured longwave radiative fluxes [Persson et al., 2002].

### 2.2. Bulk Pond Model

Our goal in developing a bulk pond model was to examine the longer-term behavior of ponds given estimates of the initial pond size, depth, and solar radiative flux. As a base model, we selected a simple rectangular pond configuration with ice melting rates set using the McPhee et al. [1987] heat and salt flux approximations described above. The heat balance for this configuration yields

\[ \frac{dT}{dt} = \frac{Q_r}{h} + \frac{F_r}{h} - \frac{2u_{ice} Q_L}{L} - \frac{2u_{ice} Q_L}{W} - \frac{4w_{ice} Q_L}{h}, \quad (12) \]

where \(T\) is the pond temperature, \(Q_r\) is the bulk radiative flux divergence computed using equations (9) and (10) over the pond depth, \(h\), \(L\) is the length of the pond, \(W\) is the width of the pond, \(u_{ice}\) is the lateral pond melting rate, and \(w_{ice}\) is the bottom melting rate. Equation (12) assumes that mixing in the pond is uniform and that the pond temperature changes simultaneously at all locations. Calculations of the ice melting rates, \(u_{ice}\) and \(w_{ice}\), are as described above in the LES model. We note that equation (12) does not account for heating of the surrounding ice, which may be significant early in the melt season when the ice has a temperature below freezing from winter cooling. Because we examine only the midsummer melt period, we assume that the ice is near the freezing temperature, and ignore this term. In the present application, calculation of \(u_{ice}\) and \(w_{ice}\) requires values for the friction velocity at the ice edge and bottom (\(u_*\) and \(w_*\), respectively). Here we use estimates of \(u_*\) and \(w_*\) from the LES simulations.

![Figure 2. Albedo predicted using radiative transfer equation presented in text for pond bottom albedo of 0.7, 0.6, and 0.5.](image-url)
2.3. LES Pond Configuration

Three basic pond configurations were considered in the LES experiments, each designed to test the importance of the pond horizontal area versus perimeter. We characterized the pond geometry using \( R = P h / A \), and circularity, \( C = P^2 / A \), where \( P \) is the perimeter length, \( h \) is the pond depth, and \( A \) is the pond surface area. Circularity provides a statistic that does not depend on pond depth and can therefore be estimated from aerial photography [Perovich et al., 2002b]. The ratio \( R \) provides a better measure of the pond melting characteristics, but requires a measure of the pond depth.

[26] Our first configuration is based on a simple rectangular pond following the bulk model described by equation (12). Circularity of a rectangle increases as the aspect ratio, defined as length/width, increases; for a square \( C = 16 \), but for large aspect ratio, \( C \sim 4 \times \text{length} \). The second configuration we considered was a circle, which yields a circularity of 20.4. This value differs from the analytical value of \( C \) for a circle (4\( \pi \)) because of the discrete shape of the pond edge on the model grid. Note that the square configuration value for \( C \) is less than the circle in this discrete case, in contrast to the analytical results. The third configuration we considered was a star shape built by varying the radius of a circle with sinusoidal functions,

\[
r_{\text{new}} = D / 2 - 0.1 r [1.5 - 0.5\{a_1 \cos(\alpha k_1) + a_2 \sin(\alpha k_2) + a_3 \sin(\alpha k_3)\}],
\]

where \( D \) is the width of the model domain in grid points, \( r = D / 2 - 4 \), \( a_1 - 3 \) are the amplitudes of the sinusoidal functions for waves with wave number \( k_1 - 3 \), and \( \alpha \) is the angle of rotation between \(-\pi \) and \( \pi \) radians. Values for \( a \) and \( k \) are presented in Table 2.

[29] Five LES domain sizes were used depending on the size of the melt pond. A description of these domains is presented in Table 3. Actual pond dimensions depended on the pond shape; however in all cases the sidewall and bottom grid points were assigned as ice over at least 4 grid points from the domain edge.

Table 2. Large-Eddy Simulation Melt Pond Experiments With Pond Configuration Constants

<table>
<thead>
<tr>
<th>Case</th>
<th>( a_1 )</th>
<th>( a_2 )</th>
<th>( a_3 )</th>
<th>( k_1 )</th>
<th>( k_2 )</th>
<th>( k_3 )</th>
</tr>
</thead>
<tbody>
<tr>
<td>P1</td>
<td>0</td>
<td>6</td>
<td>0</td>
<td>0</td>
<td>6</td>
<td>0</td>
</tr>
<tr>
<td>P2</td>
<td>1</td>
<td>4</td>
<td>1</td>
<td>25</td>
<td>6</td>
<td>70</td>
</tr>
</tbody>
</table>

*Pond shape is described in the text.

2.4. Model Initialization and Surface Forcing

Two pond initializations were used in the simulations representing well-mixed conditions and a salt stratified case. For well-mixed cases, ponds were initialized with a water temperature of 0.3°C and salinity of 0.2 psu, representing fresh water conditions expected early in the melt season just after snowmelt has completed [Eicken et al., 2004]. Ice salinity is set to 4 psu with a temperature at the freezing point. Experiments with stratification were initialized with salinity increasing from 0.2 psu at the pond surface to 4.0 psu at a depth of 0.2 m. All LES cases were forced with surface fluxes based on approximate averages from the SHEBA experiment during July [Perovich et al., 2003]. Representative values of 160 W m\(^{-2}\) for the solar flux (\( F_{\text{sun}} \)) and \(-30\) W m\(^{-2}\) for the combined sensible, latent, and infrared heat flux at the pond surface (\( F_{\text{ir}} \)) were used. Solar heating over the first half of the melting period averaged \( \sim 180 \) W m\(^{-2}\) and \( \sim 140 \) W m\(^{-2}\) over the second half (assuming a pond surface albedo of 0.07, as estimated using observations in the work of Pegau and Paulson [2001]), bracketing the 160 W m\(^{-2}\) used in our LES simulations. Melt pond simulations with the LES model were typically conducted for a 4-hour period, except in the case of weak winds and stratification, which was simulated over a 20 hour period. Simulations were generally long enough for the pond heat budget (side and bottom melting) and water temperature to reach a near equilibrium state.

[32] Initial conditions for the bulk model were set identical to the LES case assuming a well mixed pond. Because simulations were conducted for longer periods with the bulk model, we employed both constant and variable solar radiative flux and pond bottom albedo. For the constant cases, we used the same conditions as the LES cases. In the
time varying experiment, a more realistic time varying solar flux was applied, decreasing linearly from 250 W m\(^{-2}\) to 120 W m\(^{-2}\) over the 30-day duration of the simulation. These values roughly follow the observed incident solar radiation reported by Perovich et al. [2003] for the SHEBA experiment during July and early August. We also varied the pond bottom albedo linearly over the same simulation, reducing the albedo from 0.7 to 0.2, representing the darkening of the ice beneath the ponds as the ice melts.

3. Results

3.1. LES Cases

[33] As noted in section 1, our ability to parameterize melt ponds is limited by a poor understanding of pond circulation and the role of pond geometry in controlling pond growth rates. It is not clear, for example, if ponds can be treated as well-mixed or if stratification can affect the transport of heat through ponds. Consequently, our first goal was to determine how winds affect pond circulation and the heat budget, and examine the role of stratification. For these cases, we used a square pond configuration using domain 4 m as defined in Table 3.

3.1.1. Wind-Forcing

[34] We begin by presenting plots of the horizontal \(u\) velocity component and temperature for a square pond without wind stress and with wind stress of 0.01 N m\(^{-2}\) (Figure 4). Note that the wind stress value used in the simulations is very small and would be equivalent to a 10-m wind speed of about 2 m s\(^{-1}\), assuming a typical bulk parameterization for wind stress. As shown by Figure 4, a convective circulation pattern develops in the no wind case, with sinking water in the center of the pond and rising motion along the pond edges. The highest density for water occurs at about 2°C, with colder temperatures having lower density. As a result, colder water near the ice edge rises, while water warmed by solar heating in the middle of the pond sinks. The highest water temperatures are about 1°C, which corresponds to observations of ponds during nearly calm wind conditions (T. Grenfell, personal communication, 2006). Application of wind-forcing causes noticeable differences in the pond circulation as shown by the asymmetric velocity field and much colder temperatures. Horizontal velocity variations in the wind-forced case are maximized on the downwind side of the pond and have a banded structure similar to roll vortices observed during convection with shear.

[35] Plots of the sidewall heat flux for these two cases presented in Figure 5 also show significant differences. Heat fluxes were computed using equation (4) and represent the melting of ice. For example, a heat flux of 39 W m\(^{-2}\) translates to a melting rate of \(\sim 1\) cm/d with ice density of 920 kg m\(^{-3}\). In the case without wind, fluxes are similar on both of the sidewalls presented in the plot with magnitudes ranging up to \(\sim 60\) W m\(^{-2}\). Addition of wind-forcing generates a highly skewed melting rate as wind generated currents transport heat to the downwind edge of the pond. Plots of the cross wind pond sides in the wind-forced case (not shown) also show high heat fluxes on the downwind side of the pond with rapid reduction in heat flux or melting rate toward the upwind pond side.

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**Figure 3.** Melt pond outlines used in simulations for cases (a) circular, (b) P1, and (c) P2.
Comparison of the wind and no wind (convective) cases suggest that ice melting rates (or heat fluxes) along the pond sides and bottom might be dependent on the type of turbulent forcing. To test this idea, we plot the average sidewall and bottom heat fluxes for these two cases as shown in Figure 6. In the wind-forced case, the pond thermal structure reaches an equilibrium after \( \frac{C}{24} \) hours of simulation time. In contrast, the no wind convective case requires a much longer time period to reach equilibrium; after 20 hours, fluxes are still slowly changing in the no wind case. Figure 6 shows that the presence of wind generated transport has a strong influence on the pond melting characteristics. In the no wind case, melting rates are greater along the pond bottom where convective plumes directly transport water warmed by the solar flux. This pattern is disrupted by wind-forcing, which moves heat laterally to the downwind edge generating strong melting along the side of the pond. By the time the water has reached the pond bottom, it has cooled and has less heat for melting. The result is that wind-forced ponds are more likely to have increased lateral growth in comparison with situations with calm winds, where bottom melting will dominate.

One observation that can be made from these experiments is that for most ponds, even slight winds prevent convection. Summer winds during SHEBA were frequently very light. However, winds of \( 2-3 \) m s\(^{-1}\) were still common and would probably be adequate to generate wind stress of \( \frac{0.01 \text{ N m}^{-2}}{24} \) or greater as noted above. For example, Persson et al. [2002] report that the average wind speed for SHEBA during July 1998 was \( \frac{3 \text{ ms}^{-1}}{24} \), with an average wind stress (over ice) of \( \frac{0.03 \text{ N m}^{-2}}{24} \). Although at times winds were reduced to near calm conditions, most of the record shows at least \( 2 \) m s\(^{-1}\) winds. Consequently, we view the no wind case as being infrequent for typical summer arctic conditions and focus the remainder of this paper on conditions with wind-forcing.

In our next set of experiments, we wanted to determine if stronger wind-forcing causes a significant change in the melting behavior of the simulated pond. For these experiments, we used the 4 m circular pond configuration, which is a more realistic pond shape, with wind stress of \( 0.01 \) and \( 0.05 \) N m\(^{-2}\). We also examined the effects of stratification by performing a simulation initialized with increased bottom salinity as described in section 2. Results from these simulations are presented in Figure 7 showing...
the average bottom heat flux over a 4 hour simulation. Sidewall fluxes are not shown since they converge to an equilibrium value similar to the square pond result in Figure 6a. As Figure 7 shows, increasing the wind speed does not cause a substantial change in the bottom melting rates after the model reaches steady state. Also, we find that wind-forcing quickly destroys the stratified layer in the pond so that a well-mixed state is achieved in about 1 hour. Similar results for stratification were obtained when the wind stress was reduced to 0.01 N m\(^{-2}\) (not shown); however, equilibrium was not reached until about 3 hours into the simulation.

3.1.2. Pond Shape and Size

Key parameters governing the effective albedo of melt ponds on sea ice are the horizontal size of the ponds and the depth of the pond water. These factors may also be important for determining how quickly a pond will grow since pond size is a function of both drainage of meltwater and melting along the pond bottom and sides. Conservation of heat within ponds requires that decreased bottom melting must be compensated either by transport of heat from outside of the pond by lateral heat flux, or by increased edge melting. In this paper, we cannot easily address the loss of heat through lateral fluxes without modeling the detailed physics of the ice, which was not our main goal. However, we know from observations that ponds often have complex shapes that would tend to increase the sidewall area relative to the total pond bottom area. Our next set of experiments examine how changing the pond shape and size affects melting rates.

Results are first presented showing the horizontal \(u\) component velocity and temperature for ponds with increasing complexity for domain size 4 m (Figure 8). As in the square pond case, each simulation shows velocities with increasing variability in the downwind direction. However, in these cases a more noticeable large-scale circulation pattern is evident as shown by the lower velocities in the center of the circular pond (C). The large-scale pattern is less noticeable in ponds P1 and P2. Temperature patterns in the three cases display systematic structures with colder water appearing in 2–3 bands aligned with the wind.
direction. Temperature variability appears to decrease as the pond complexity increases as shown by the reduction in extremes between case C and P2.

Plots of the pond average temperature and heat fluxes for the three pond shapes, along with the square pond (Figure 9), demonstrate how the complexity of the pond perimeter affects the pond heat budget. In each case, as the perimeter size of the pond increases, the average bottom flux decreases. Average sidewall fluxes also decrease, but the total sidewall area increases more rapidly than the bottom area as demonstrated by the increasing circularity of each case (see Figure 3). The net effect is that ponds with greater sidewall complexity will not deepen as rapidly as ponds that have uniform shape in all directions. The effect can be significant; bottom fluxes with case P2 are \(~25\) W m\(^{-2}\) in comparison with \(~35\) W m\(^{-2}\) for the

![Figure 6.](image1)

![Figure 7.](image2)

**Figure 6.** Time series plot of the average heat flux along the bottom and sidewalls of a square pond using domain 4 m with (a) no wind stress and (b) wind stress of 0.01 N m\(^{-2}\). Note the much longer simulation time in the case without wind forcing.

**Figure 7.** Average bottom heat flux used to melt ice in a circular pond with a domain size of 4 m with wind stress of 0.1 and 0.5 N m\(^{-2}\) and stratification.
circular pond, which translates into a melting rate decrease of ~30 percent for the complex pond.

We note that the pond temperature follows a slightly different trend in the four cases. The equilibrium pond temperature is determined by the pond circulation and the strength of turbulence near the sidewalls and pond bottom. In the more complex ponds (e.g., P2), the pond edge may act in a manner similar to radiator fins, increasing the cooling rates and lowering the equilibrium temperature. Cooler average temperatures for the square pond in comparison with the circular pond may be in response to the stronger current system and greater heat transport within the square pond.

Pond size is also important in setting the pond melting rates as shown by plots of bottom fluxes for different pond sizes in Figure 10. In general, as the pond size increases, the average bottom flux increases. This is a consequence of the ratio $R$ being inversely proportional to the pond width (for square ponds) or diameter (for circular ponds). Of course pond depth is also important for determining the relative strength of side versus bottom melting rates. Here we use pond depths that are appropriate for the pond size, using observations from the SHEBA data set as a standard.

Results shown in Figure 10 suggest that early in the melt season when ponds are relatively small, the total ice
albedo will change more through the expansion of ponds than by pond deepening (assuming that pond size is not dominated by drainage). As the season progresses and ponds increase in size, changes in the albedo will be more tied to the increased depth of the larger ponds. Late in the melt season, our results show that much of the solar energy entering the pond will be devoted to bottom melting.

[45] At first glance, the degree of pond edge complexity indicated by case P2 appears to be quite high. However, ground level photographs of ponds during SHEBA [see Perovich et al., 2003] show that ponds are often connected by long channels and have numerous inlets and embayments. In addition, the pond depth during SHEBA was not uniform and exhibited many shallow regions where the melting process could not be directly simulated using the simplified configurations used here.

[46] Another way to increase $R$ for a pond is to consider a channel configuration. Ponds are often observed as elongated features, which do not fit the circular or oval shapes considered above. The channel configuration also has the advantage of matching the geometry of the bulk pond model given by equation (12). Consequently, the LES results for

![Figure 9](image-url)  
Figure 9. (a) Horizontally average temperature, (b) heat flux from melting along the pond bottom, and (c) heat flux along the pond sidewalls for domain size 4 m.
this configuration can be used to set parameters for the bulk pond model and examine the longer-term pond evolution.

[47] We conducted a single experiment with a channel pond using the domain size 10 m as described in Table 3 as a test to determine if ponds with similar $R$ values behave in the same way regardless of shape. As shown in Table 4, the channel pond (labeled Rectangle) and pond P1 have reasonably close values of $R$. Close correspondence between bottom heating rates for these two cases as shown in Figure 11 suggest that $R$ could be a robust statistic for characterizing pond melting behavior. For actual ponds it may not be possible to easily characterize the pond geometry because of sloping sidewalls and variable pond depth as shown by pond cross sections taken during the SHEBA experiment (Figure 12). As these sections show, in some cases ponds have a fairly sharp edge and nearly flat bottom. Nevertheless, characterizing ponds in terms of sidewall and bottom area provides a first step for building a useful pond energy budget model, for example, as given by equation (12).

[48] Pond shape can also determine the strength of currents near the pond edges (e.g., see Figures 4 and 8).
Figure 11. Average heat flux along the pond bottom for a pond with pond shape P1 for a 4 m domain size and a rectangular pond with dimensions of 10.18 m × 0.56 m.

Figure 12. Observed pond depths taken from the Surface Heat Budget of the Arctic Ocean (SHEBA) albedo line on (a) 29 June 1998 and (b) 8 August 1998.
In general, if currents are reduced, then melting will be suppressed because of lower heat transport rates and local friction velocity. Reduced heat transport causes an imbalance in the heat budget and an increase in the pond temperature until the side and bottom fluxes increase to match the input solar heat. The effects of the pond shape can be quantified by plotting the average sidewall and bottom friction velocity, $u^*$ and $w^*$, for ponds with shapes of square, circle, and P2 using the 4 m domain (Figure 13). Friction velocities are greatest along the sidewalls in the square pond, which also shows the largest difference between the sidewall and bottom values. Friction velocities decreased overall as the pond complexity increased, and showed almost no difference between the bottom and sidewalls except in the square pond case.

### 3.2. Bulk Pond Model

Simulation of melt ponds with the LES/radiative transfer model can only be conducted for short time periods because of the high computational costs. To model longer time periods, we employ the simplified bulk approach as described in section 2. Results from the LES model show that mixing in ponds is in most cases very rapid unless winds are nearly calm. Even under calm conditions, natural

![Figure 13. Averaged friction velocity for the sidewalls ($u^*$) and bottom ($w^*$) of ponds with shapes (a) square, (b) circular, and (c) P2, using the 4 m domain size.](image-url)
convection in the pond causes mixing that eventually forces an equilibrium temperature and a balance between input solar radiation and melting of the surrounding ice. With or without winds, we do not see strong temperature stratification that would tend to enhance melting near the top of the pond edge relative to the bottom. Because ponds are generally well mixed, we selected a bulk approach that assumes vertically oriented sidewalls and partitions energy between the bottom and sides of the pond.

A number of physical parameters need to be estimated in the bulk model to replicate the basic behavior of ponds simulated with the LES. Probably the most critical factor in the model is the pond width and length, which account for the difference between the side and bottom area melting as discussed above. On the basis of our LES experiments, we use the ratio of the sidewall area to the pond bottom to characterize pond complexity, which for a rectangular pond is defined as

\[ R = \frac{Ph}{2(L + W)h} = \frac{Ph}{LW} \]  

A second key parameter for ice melting is the strength of the near-wall turbulence as indicated by the sidewall and bottom friction velocities, \( u_r \) and \( w_r \), respectively. As shown above in the LES experiments, turbulence strength near the wall controls the flux of heat and salinity as ice melts. Except for the square pond case, friction velocities predicted by the LES model were roughly the same for the pond bottom and sides with values between 0.008 and 0.005. The more complex the pond shape (i.e. larger \( R \)), the lower the friction velocity. For the tests presented here, we used friction velocities taken from LES cases with similar values of \( R \).

We first test the bulk pond model by simulating a square pond duplicating the LES 4 m domain case using values for \( u_r \) and \( w_r \) of 0.0011 and 0.0007 (taken from Figure 13) with a constant solar heat flux of 160 W m\(^{-2}\) and net surface flux of \(-30\) W m\(^{-2}\). A 30-day simulation is shown in Figure 14. Overall, predicted pond characteristics for this simple case vary almost linearly in time. Initially, the pond bottom flux of 36 W m\(^{-2}\) is nearly the same as the LES square pond bottom flux shown in Figure 6a for the wind-forced case. Over time, the albedo decreases from \(-0.35\) to \(-0.29\) as the depth increases from an initial value of 0.24 m to \(-0.55\) m. The final albedo for this case is considerably larger than observations of pond albedo (\(-0.1\) to \(-0.2\)) for ponds with similar depth [e.g., see Perovich et al., 2003], probably because we hold the pond bottom albedo to a constant value of 0.7.

Our next test with the bulk model examined a case matching the rectangular pond simulated with the LES pond model (see Figure 11). Results from this case are presented in Figure 15, and again show initial values that are very similar to the LES model output for bottom heat flux and temperature (not shown for the LES), and albedo values that are higher than comparable observations by the end of the 30-day period.

Overall, the bulk pond model described by equation (12) yields consistent results when compared with the LES model, suggesting that it might be used to parameterize variable shaped ponds by matching the value of \( R \) and including time varying solar heating and pond bottom albedo. For example, we present a simulation using the bulk model with solar radiation varying linearly from 250 W m\(^{-2}\) to 100 W m\(^{-2}\) [Perovich et al., 2003] and the pond bottom albedo varying linearly from 0.7 to 0.2 over a 30-day period. Pond dimensions for this case are as in the rectangular case discussed above, but with an initial depth of 0.05 m, which is closer to the observed initial pond depths from SHEBA.

With this forcing, the bulk model predicts a pond depth increasing to just under 0.4 m and albedo decreasing from \(-0.4\) to \(-0.1\) (Figure 16). These values are similar to the observed pond characteristics presented by Perovich et al. [2003, 2002a] for the month of July 1998, and with albedo observations presented by Barber and Yackel [1999]. Caution should be used when comparing the pond depth from the bulk model with observations. Here we do not account for pond depth variability from surface meltwater runoff or changes in the ice free board. Nevertheless, this demonstration suggests that the bulk model could form the basis of a thermodynamic pond model if combined with ice components as in the work of Taylor and Feltham [2004] and Lüthje et al. [2006].

4. Conclusions

Observations of the Arctic ice pack have clearly shown that melt ponds are a key component in setting the ice surface albedo and defining the local energy budget. In this study, we examine the heat budget of an idealized melt pond using a combination of an empirical radiative transfer model with a large-eddy simulation turbulence model and a simple bulk pond model. The empirical radiative transfer model is based on observed values of solar flux from a fresh water capped polar lead during the SHEBA experiment, and yields albedo values that are in good agreement with pond observations. Experiments using the radiative model assume a one-dimensional configuration, ignoring radiation that is reflected from the pond sidewalls. We also ignore the effects of heat transported into the pond from lateral meltwater flux.

Results from the LES pond model indicate that ponds are very likely to develop well-mixed conditions even when ponds begin with significant stratification. Simulations show that very light winds (2–3 m s\(^{-1}\)) are able to move pond water vertically, causing transport of heat throughout the pond system. When stratified with a 4 psu salinity gradient, the model predicts a delay in pond mixing of several hours, but still develops a well-mixed equilibrium state.

Well-mixed conditions in ponds suggest that heat will transfer equally to the side and bottom of the pond. For the most part, our experiments support this hypothesis; however, pond shape does limit the movement of water even when well mixed. Simulations with a variety of pond shapes and sizes show that the basic ratio of sidewall area to bottom area, \( R \), can be used to characterize most ponds. For example, as pond size increases, the bottom area increases in relation to the side area, decreasing \( R \). Consequently, relatively small ponds have a larger lateral growth rate from melting in comparison with large ponds.
Figure 14. Time series plot of pond parameters from the bulk pond model for a square pond with initial length of 3.72 m, depth of 0.24 m, and constant forcing as presented in text. Solar flux represents the total solar radiation absorbed by the pond water.
Pond shape can also change the value of $R$ as shown by the simple case of an elongated or channel-shaped pond, or in ponds with a complex shapes as typically observed by aerial surveys [e.g., Barber and Yackel, 1999]. Simulations with the LES model of rectangular ponds and ponds with increasingly convoluted sidewalls show that the more elongated or complicated the pond sidewalls, the more ponds lose heat through sidewall melting at the expense of bottom melting.

Results indicating that ponds are frequently well mixed prompted the development of a bulk pond model based on a rectangular geometry. Adjustable parameters in this model are limited to the length and width of the pond, which provides a means of changing $R$, and values for the friction velocity along the pond sidewalls and bottom. When forced with solar heating and net surface flux as applied in the LES model, the bulk pond predicts bottom heat fluxes and pond temperatures that are in good agreement with the LES model for ponds with similar $R$ values. A single test for a 30-day period initialized with a small, shallow rectangular pond shows that the bulk model develops a realistic pond depth and albedo in comparison with observed pond characteristics in late summer, even though

![Figure 15. Same as Figure 14 but for a rectangular pond with dimensions of 15.24 m × 0.87 m, matching the large-eddy simulation (LES) case.](image)
the bulk model does not consider water flux from surface melting or changes in the overall ice thickness.

Future research is planned to integrate the bulk pond model into a full melt pond sea ice model using an approach similar to that of Lüthje et al. [2006] and Eicken et al. [2004]. Experiments performed here are only able to give us estimates of the pond melting rates and do not address pond evolution forced by water inflow and outflow from the surrounding ice. Clearly, these aspects of pond behavior are also a key element in controlling the ice-albedo feedback.

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References

Figure 16. Bulk model simulation of a 15.27 × 0.87 m pond with an initial depth of 0.05 m. Variable solar forcing and pond bottom albedo are as described in the text.


Grenfell, T. C., and G. A. Maykut (1977), The optical properties of ice and snow in the Arctic basin, J. Glaciol., 18(80), 445–463.


C. A. Paulson and E. D. Skyllingstad, College of Oceanic and Atmospheric Sciences, Oregon State University, Corvallis, OR 97331, USA. (skylling@coas.oregonstate.edu)