Changes in the modeled ice thickness distribution near the Surface Heat Budget of the Arctic Ocean (SHEBA) drifting ice camp

R. W. Lindsay
Polar Science Center, Applied Physics Laboratory, University of Washington, Seattle, Washington, USA

Received 17 January 2001; revised 21 November 2002; accepted 16 December 2002; published 19 June 2003.

[1] In the polar oceans the ice thickness distribution controls the exchange of heat between the ocean and the atmosphere and determines the strength of the ice. The Surface Heat Budget of the Arctic Ocean (SHEBA) experiment included a year-long field program centered on a drifting ice station in the Beaufort and Chukchi Seas in the Arctic Ocean from October 1997 through October 1998. Here we use camp observations and develop methods to assimilate ice thickness and open water observations into a model in order to estimate the evolution of the thickness distribution in the vicinity of the camp. A thermodynamic model is used to simulate the ice growth and melt, and an ice redistribution model is used to simulate the opening and ridging processes. Data assimilation procedures are developed and then used to assimilate observations of the thickness distribution. Assimilated observations include those of the thin end of the distribution determined by aircraft surveys of the surface temperature and helicopter photographic surveys and aircraft microwave estimates of the open water fraction. The deformation of the ice was determined primarily from buoy and RADARSAT Geophysical Processor System (RGPS) measurements of the ice velocity. Because of the substantial convergence and ridging observed in the spring and summer, the estimated mean ice thickness increases by 59%, from 1.53 to 2.44 m, over the year in spite of a net thermodynamic ice loss for most multiyear ice. INDEX TERMS: 1863 Hydrology: Snow and ice (1827); 3337 Meteorology and Atmospheric Dynamics: Numerical modeling and data assimilation; 3339 Meteorology and Atmospheric Dynamics: Ocean/atmosphere interactions (0312, 4504); 4207 Oceanography: General: Arctic and Antarctic oceanography; 4540 Oceanography: Physical: Ice mechanics and air/sea/ice exchange processes; KEYWORDS: Lagrangian, assimilation, characteristics, deformation, Kalman, albedo


1. Introduction

[2] The ice thickness distribution has long been recognized as a key state variable for the Arctic climate system. Rates of heat transfer between the ocean and the atmosphere as well as rates of ice production and melt are closely related to the ice thickness distribution [Maykut, 1982]. The surface heating and evaporation rates affect the lower atmospheric stratification, cloudiness, and circulation, while melt and freezing rates and the resulting salt or fresh water fluxes affect the ocean stratification and circulation. Coupled global atmospheric and oceanic general circulation models must accurately represent the evolution of sea ice and the ice thickness distribution to properly represent surface processes in the polar oceans. Many global climate models show an early and more extreme response in the polar regions to increased levels of CO₂, a response that is intimately connected to changes in the sea ice coverage [e.g., Pollard and Thompson, 1994; Rind et al., 1995]. However, these models have very rudimentary sea ice components. The treatment of the ice thickness in several regional coupled ice-ocean models is discussed by Kreyscher et al. [2000].

[3] The Surface Heat Budget of the Arctic Ocean (SHEBA) experiment, which included a year-long field program centered on a drifting ice station in the Beaufort Sea [Perovich et al., 1999b], offered an unprecedented opportunity to combine detailed observations of the weather, ice and snow properties, ice thickness, open water, ice motion, and ice deformation in order to estimate the changes in the ice thickness distribution over a full seasonal cycle. These observations can be used to both evaluate methods of modeling the ice as well as to gain insight on the principal physical processes that affect the ice thickness.

[4] Here we develop methods for assimilation of ice thickness and open water observations into a thickness distribution model in order to make as accurate an estimate as possible of the evolution of the distribution in the vicinity of the SHEBA camp. The model uses as many of the field observations that can be usefully incorporated, so that many of the free variables that must be calculated in a climate model are taken from observations. For example, the pond fraction, a key component of the surface albedo, is prescribed in this model from aerial observations. Thus the motivation
for formulating the model is quite distinct from a climate model, which must calculate many of the parameters for which we have observations. The resulting thickness distribution may be more accurate than that determined by a climate model and may be of use to investigators testing and developing climate models who do not have the luxury of assimilating observations of ice thickness or extent or of having ancillary observations to aid in constraining the model state.

[5] This paper is organized as follows: the essential properties of the ice thickness model are reviewed (section 2), followed by a discussion of the observations used to drive the model (section 3). A discussion of data assimilation issues and methods and the sources of ice thickness data that are assimilated follows (section 4), and finally the results are presented for the SHEBA year (section 5).

2. Model Description

[6] The theory of the evolution of the ice thickness distribution $g(h)$ was first outlined by Thorndike et al. [1975]. The distribution is defined as

$$ g(h) = \text{fraction of area covered by ice of thickness between } h \text{ and } h + dh $$

and the governing equation for $g(h)$ is

$$ \frac{\partial g}{\partial t} = - \nabla \cdot (vg) - \frac{\partial}{\partial h} (fg) + \Psi + \phi, $$

where $v$ is the ice velocity vector, $f$ is the growth rate, $\Psi$ is a redistribution function, and $\phi$ is the lateral melt term added by Hibler [1980]. Flato and Hibler [1995] investigated the importance of accounting for both ridged ice and undeformed ice as separate but related distributions. Discussions of the nature of the ice thickness model are given by Thorndike et al. [1975], Hibler [1980], Rothrock [1986], and Flato and Hibler [1995]. A recent review of ice models is given by Steele and Flato [2000].

[7] The model is based on a set of Lagrangian cells, each representing a limited region of ice. The initial spacing of the Lagrangian cells is 25 km. The motion and deformation of each cell are estimated from a variety of ice motion observations (see section 3). The motion and various output parameters are also represented on a grid box that follows the motion of the SHEBA camp but does not deform. The camp remains in the center of the grid. This moving grid measures 200 km on a side and has a spacing of 25 km. Thus we have a Lagrangian model within a rigid moving grid. This is similar in concept to the particle-in-cell (PIC) model of Flato [1993], which is used for ice edge forecasting. However, because our primary region of interest is the vicinity of the SHEBA camp, the grid of forcing parameters and the output grid must be mobile.

[8] The Lagrangian cells represent the thickness distribution on a “characteristic grid” suggested by Rothrock [1986] and used by Bjork [1992], Schramm et al. [1997], and Bitz et al. [2001]. A characteristic line $h(t)$ represents a solution of the thermodynamic growth equation

$$ \frac{dh(t)}{dt} = f(h, t), $$

where $f$ is the growth rate. The thickness distribution is represented by a set of discrete ice thickness values $h$ and a fractional area $g(h)$ associated with each. With characteristics the change in ice thickness due to thermodynamic processes is explicitly calculated for each characteristic so the model is Lagrangian in thickness as well as in space. A characteristic is created whenever there is an episode of open water creation (from divergence or shear) or when there is an episode of ridging. Each characteristic has associated with it a thickness $h_i$, an area fraction $g_i$, age $A_i$, and a ridging status $r_i$, which specifies if the characteristic represents undeformed ice ($r_i = 0$) or ridged ice ($r_i = 1$). In addition to the ice thickness distribution, the following parameters are determined for each cell: position, velocity, total area, and the uncertainties in each of these parameters. Note that while total area is calculated from the integral of the divergence, the shape of the cell boundary is not determined. Cells can be created in the model at any time to represent a parcel of ice that we wish to track, for example the location of a buoy or of a survey area. Cells are also created in regions where divergence has reduced the number density of cells below a critical value. The newly created cells take on the mean properties of the surrounding cells. In this simulation, with an initial configuration of 17 $\times$ 17 cells on a 25-km grid (a 400-km box), nearly half (46%) were created during the year to fill gaps that developed in the spatial coverage or, more commonly, to fill edge regions. Within the 200-km box centered on the camp, our principal area of interest, 23% of the cells were created during the year.

[9] An important aspect of the model is that the thermodynamic growth and melt rates are calculated only at the camp location and only for a set of fixed ice thickness bins. The growth and melt rates for each characteristic are then interpolated from these fixed bins. This means that many computationally intensive operations can be performed for a few ice thickness bins and at a single location. The maintenance of ice temperature profiles only for fixed thickness bins and for one location means that the temperature profiles do not need to be merged when characteristics are merged, or be created when ice ridges. The model can be configured to account for spatially varying thermal forcing fields by computing the growth and melt rates at a grid of points instead of at only one point. The growth and melt rates would then be interpolated in space to the locations of the Lagrangian cells. However, in this study the thermal forcing parameters are available only at the camp, so the growth rates are calculated for the camp location and assumed to be applicable to the entire region.

[10] The Lagrangian representation avoids the necessity of calculating approximations for the advective terms for the ice thickness in geometric space, and the characteristic grid avoids the need to calculate approximations for the growth or melt of ice that moves between fixed thickness bins. The model consists of a thermodynamic component to determine bottom and top melt and growth rates, and an ice redistribution component to account for opening, closing, and ridging.

2.1. Thermodynamic Model

[11] The second term in equation (2) includes the growth or melt rate factor $f$ which is determined by the thermodynamic evolution of the ice. The terms of the energy balance for a slab of ice include the net radiation of the surface $F_r$, the
sensible heat flux at the surface $F_s$, the latent heat flux at the surface $F_l$, the conductive flux from the bottom of the ice $F_b$, and the penetrating solar flux $F_p$. The energy balance for an entire slab of ice is then expressed as a sum of the top and bottom fluxes minus the flux of energy stored in the ice $S$ and the flux of energy available for melting ice $M$.

$$F_t + F_s + F_q + F_b - F_p - S - M = 0.$$  (4)

All fluxes are considered positive if directed toward the ice. The meltwater, determined by $M$, is assumed to either stay in ponds or to run off the ice.

[12] The thermodynamic model follows that of Maykut and Untersteiner [1971] in that it is a time-dependent thermal model of the ice with multiple layers. The energy balance is similar to the treatments found in Maykut [1982], Bjork [1992], and Ebert and Curry [1993]. It is more fully described by Lindsay [1998]. Here we will give a brief outline of the salient features of the model.

[13] The thermodynamic fluxes are computed for a set number of fixed ice thickness classes, $n_s = 8$ (0.0, 0.25, 0.5, 1.0, 2.0, 3.0, 4.0, and 8.0 m), with the goal of determining the ice growth or melt rate for each of the classes. These classes are not related to the ice thickness distribution, and the thermal profiles and energy fluxes are computed for each class even if no ice in the vicinity is currently of a similar thickness. Furthermore, the thickness of each class does not change, for it is the growth or melt rates only that are needed for application to the characteristics. The number of layers in the vertical is fixed at $n_z = 7$ for all thicknesses: 2 in the snow (top surface and half way down) and 5 in the ice (at the snow-ice interface, 0.1$h$, 0.5$h$, and 0.9$h$ down from the interface, and at the bottom). The time step for the thermodynamic model is 1 hour and the heat equations are solved with a fully implicit method, a flux boundary condition at the top, and a fixed temperature (the melting point) at the bottom.

[14] The net radiation at the surface is

$$F_t = (1 - \alpha)F_{dsw}(1 - L_o) + F_{dsw} - \varepsilon\sigma T^4_{sfc},$$  (5)

where $F_{dsw}$ and $F_{dwb}$ are the downwelling shortwave and longwave fluxes at the surface, $\alpha$ is the albedo of the ice, $L_o$ is the fraction of the net shortwave flux that penetrates through the top layer, $\varepsilon$ is the surface emissivity, $\sigma$ is the Boltzmann constant, and $T_{sfc}$ is the surface temperature. The treatment of the albedo is discussed in Section 3.3. The net solar radiative flux is allowed to penetrate the surface and be absorbed within the ice [Grenfell, 1979]. The amount of penetration depends on the snow cover, the cloud fraction, and the depth in the ice. When snow is present, virtually all of the net solar flux is absorbed within the snow; when it is absent, about 50% passes through the top 0.25 m of ice. During the summer, the penetrating solar flux causes much of the ice melt to occur below the surface. The snow depth and the snow density are prescribed on the basis of observations from the SHEBA camp [Perovich et al., 1999a].

[15] The sensible heat flux is determined by the difference between the model-calculated surface temperature and the measured air temperature at a reference height of 10 m, using a stability-dependent exchange coefficient [Jordan et al., 1999] appropriate for sea ice,

$$F_s = \rho c_p C_s U_i(T_r - T_{sfc}).$$  (6)

where $\rho$ and $c_p$ are the air density and heat capacity, $U_i$ is the wind speed, and $T_r$ is the measured air temperature. The reference height is $r = 10$ m. $C_s$ is the heat transfer coefficient and is a function of the bulk Richardson number of the surface layer. When the wind speed is near zero the heat flux (in W m$^{-2}$) is not allowed to fall below $T_r - T_{sfc}$ (in K) [Jordan et al., 1999]. The latent heat is determined with the model-calculated surface mixing ratio $q_{sfc}$ and the mixing ratio at the reference height $q_r$.

$$F_q = \rho L C_q U_i(q_r - q_{sfc}).$$  (7)

[16] The heat conducted in the snow is proportional to the temperature gradient and the thermal conductivity of the snow $k_{snow}$, Maykut [1982] suggests a constant $k_{snow} = 0.31$ W m$^{-1}$ K$^{-1}$. There are at least three other possibilities used in the recent literature: two based on theoretical considerations, including vapor transport, by Ebert and Curry [1993] and by Jordan et al. [1999], and a third by Sturm et al. [1997] based on a wide range of field measurements. While the Ebert and Curry [1993] and Jordan et al. [1999] formulations are similar in magnitude, the Sturm et al. [1997] values are about half as large. The Sturm et al. [1997] parameterization is also in agreement with the conductivity measured by Sturm et al. [2002] in the vicinity of the SHEBA camp in the early spring. The average value they measured with 101 samples was 0.123 W m$^{-1}$ K$^{-1}$. In order to decide which conductivity formulation to use, we calculated the model-derived sensible heat flux for a constant value of $k_{snow} = 0.31$ W m$^{-1}$ K$^{-1}$ as well as for each of these three formulations for an ice thickness of 2 m and a snow depth as measured at the camp. The model daily averaged sensible heat flux was compared to the flux measured by the SHEBA Atmospheric Surface Fluxes Group (SAFG). The RMS difference between the measured and the modeled heat flux was very similar for the constant value (4.2 W m$^{-2}$) and for the first two density-based formulations (4.3 W m$^{-2}$), while it was worse for the Sturm et al. formulation (5.2 W m$^{-2}$). The improved comparison using a larger conductivity value than that measured by Sturm et al. near the camp may be related to wind pumping of heat within the snow. As a result of these comparisons, the conductivity is set to the Maykut [1982] value, $k_{snow} = 0.31$ W m$^{-1}$ K$^{-1}$.

[17] The heat flux from the ocean is computed following McPhee et al. [1999] from

$$F_w = \rho c_p C_{ml} u_w \delta T,$$  (8)

where $\delta T = T_{ml} - T_{sfc}(S_{ml})$ is the difference between the temperature of the mixed layer $T_{ml}$ (obtained from observations) and the freezing temperature (a function of the observed mixed layer salinity $S_{ml}$). The friction velocity $u_w$ is determined by solving the Rosby similarity drag law for the difference between the complex ice velocity $U_i$ and the geostrophic ocean current velocity $U_g$,

$$U_i - U_g = \frac{u_w}{\kappa} \left[ \log \left( \frac{u_w}{F_0} \right) - A - iB \right]$$  (9)

with similarity parameters $A = 2.0$ and $B = 2.5$. $\kappa$ is von Karman’s constant and $f$ is the Coriolis parameter. The
roughness length \( z_0 = 1.5 \text{ cm} \) and the heat transfer coefficient \( C_{hi} = 0.006 \) are determined from SHEBA observations [McPhee, 2002; M. McPhee, personal communication, 2000]. When positive, the ocean flux is divided evenly between bottom ablation and lateral melt. The ratio of bottom flux to lateral melt is often taken as 0.5 [e.g., Hibler, 1980] but is known to depend strongly on the floe size distribution [Steele, 1992]. Experiments with the current model including assimilation of open water fractions show that if the assumed fraction of the ocean flux that goes to lateral melt is taken to be 0.25, the mean ice thickness at the end of the year is 7% less (and 7% more for an assumed fraction of 0.75). The modest sensitivity of the model mean ice thickness may be related to the fact that the ocean flux is derived from the observed water temperature and is not computed in an interactive fashion dependent on the estimated absorbed solar flux. This later method would amplify the sensitivity of the model to changing open water fractions.

[18] The bottom flux \( F_b \) in equation (2) is the conductive flux in the lowest layer. The temperature of the bottom of the ice is assumed to be at the freezing point of the mixed layer \( T(S_{ml}) \). Consequently, in the winter there is a substantial heat flux from the bottom surface. This flux largely comes from the latent heat of freezing at the bottom surface. In summer the ice temperature rises to the melting point which, because of the low salinity of the ice, is warmer than the mixed layer freezing temperature \( T(S_{ml}) \). Hence the bottom heat flux reverses sign.

[19] The growth rate at the bottom is determined for each thickness category from the difference between the conductive flux at the bottom and the oceanic heat flux:

\[
f_{\text{fem}} = L_{\text{ice}}(F_b - F_\text{a}),
\]

where \( L_{\text{ice}} \) is the latent heat of freezing of ice at a salinity of 3 ppt.

[20] The flux available for melt is found from the energy that accumulates within each layer that would raise the temperature above the melting temperature. The melt rate within the ice and at the surface is found from the total of the flux available for melt in each layer and the latent heat of fusion of the ice \( L_{\text{ice}} \)

\[
f_{\text{melt}} = L_{\text{ice}}q_{\text{ice}} \sum_{i=1}^{n_i} M_i.
\]

[21] The penetrating solar flux \( F_\text{p} \) is calculated but it is not used in the ocean mixed layer model because we are able to use camp observations of the mixed layer temperature. The flux storage term \( S \) is determined from the rate of change of the model ice temperature.

2.2. Ice Redistribution Model

[22] The thickness distribution is modified by ice deformation through the creation of open water during divergence and by ridging during convergence and shear. The model approach follows closely that outlined by Thorndike et al. [1975], Björk [1992], Schramm et al. [1997], and Bitz et al. [1996] use similar, but slightly different, techniques. Ridging is controlled by two key parameters, \( G^* \), the fraction of area that participates in ridging, and \( k \), the multiplier that determines the thickness of ridged ice. At each time step the characteristics are sorted by thickness (smooth and ridged together, starting with open water) and the number of characteristics that participate in ridging is determined from the distribution and \( G^* \). An equal number of new characteristics are created with thickness equal to the product \( kh_i \), where \( h_i \) is the original thickness of a characteristic. The rate of change of the cumulative distribution function \( dG/dt \) is found following Thorndike et al. [1975] with the ice growth, \( dh/dt = f(h) \), computed in the thermodynamic model described above. The time integration of \( G \) and \( h \) is performed simultaneously with a fourth-order Runge-Kutta procedure. The time step for the redistribution model is one day.

[23] The ice deformation process continually creates new characteristics, so in order to keep the number of characteristics to a reasonable value, a merging procedure is required. Björk [1992], Schramm et al. [1997], and Bitz et al. [2000] discuss merging methods. Our method is similar, but somewhat simpler, since there is no requirement to merge temperature profiles along with the characteristics. All characteristics with no ice (\( g(h) = 0 \)) are removed and all those with zero thickness (\( h = 0 \), open water) are combined. Characteristics with less than 0.1% of the area are merged with the nearest characteristic in thickness. If the number of characteristics in a cell exceeds the maximum (usually 50) the characteristics are examined in order of increasing thickness for smooth and ridged ice independently. If the difference in thickness of adjacent characteristics is less than 1%, the characteristics are merged with a weighted average. The procedure is repeated with increasing thresholds until the number of characteristics is reduced to the maximum. This simple fractional rule keeps the thickness resolution higher for thin ice than for thick ice.

[24] The model results are sensitive to the maximum number of thickness characteristics allowed, similar to what Schramm et al. [1997] found. Figure 1a shows how the mean ice thickness in the last six years of a 10-year run depends on the maximum number of characteristics allowed. The thermodynamic growth rates for these simulations are computed with climatological forcing values derived from the drifting ice stations of the former Soviet Union [Lindsay, 1998]. The mean ice thickness first increases and then decreases with the maximum number of characteristics allowed, but at 50 characteristics it is within 1% of the 100 characteristic value. The computation time is not a strong function of the number of characteristics allowed, for while the redistribution model may require more time as more characteristics are allowed, the procedures for pruning the number of characteristics take less time. In the rest of this study 50 characteristics are used.

[25] The parameters \( G^* \) and \( k \) are not well known and we use values suggested by Thorndike et al. [1975], \( k = 5 \) and \( G^* = 0.15 \), although Schramm used \( k = 15 \) and \( G^* = 0.1 \). Comparing submarine measurements of the thickness distribution, Babko et al. [2002] found that rafting, for which \( k = 2 \), plays a significant role in the ice redistribution. Figure 1b indicates that the mean ice thickness increases with both \( k \) and \( G^* \) and is indeed sensitive to these parameters. Model simulations using the SHEBA
camp conditions, including data assimilation, indicate that if $k$ is reduced from 5 to 2, the final mean ice thickness is reduced by 5%, and if $k$ is increased to 15, the mean thickness increases by 5%. If $G^*$ is decreased from 0.15 to 0.05, the mean ice thickness decreases by 24%, while if it is increased to 0.20, the mean ice thickness increases by 11%. These results are consistent with the representation in Figure 1. There are currently insufficient measurements of the evolution of the ice thickness distribution and deformation histories to adequately select the proper values and ultimately it is likely that they will be found to change with the nature of the distribution and the nature of the deformation.

3. Camp Observations

[26] A wide variety of observations from the camp and in its vicinity were used for forcing parameters and for adjusting parameterizations within the model. Thermal forcing data were obtained from measurements made at the camp by the SHEBA Project Office (SPO), the SAFG, and the Atmospheric Radiation Monitoring program (ARM). Data from these groups were obtained from the National Center for Atmospheric Research (NCAR) Joint Office for Scientific Support (JOSS) data catalog. At the beginning of the experiment, before 8 October 1997, climatological values were used. This was necessary because the thickness distribution used for initializing the model run was from the SCICEX survey taken on September 28th through October 1st and it was necessary to project this initial measurement of the thickness distribution to the time of the first camp observations in a realistic manner.

3.1. Wind, Temperature, and Radiation

[27] The 10-m wind speed and 10-m air temperature were obtained from the SPO meteorological towers. Hourly composite values were formed from the 10 minute averages reported in the raw data. The composites consisted of the average of the observations from both towers after values were discarded following various quality checks. Lack of accurate humidity measurements forced us to set the relative humidity of the air at 10 m to be 100% with respect to ice. This is in agreement with spot measurements with a dew point hygrometer at the camp (E. Andreas, personal communication, 1998). The computed humidity fluxes over thick ice are hence quite small but over thin ice they are often large. Camp observations extended from day 282.0 (9 October 1997) to day 648.0 (10 October 1998).

[28] The downwelling longwave and shortwave radiative fluxes are mostly from the SAFG measurements but these are supplemented with SPO and ARM measurements when needed. A composite data set was formed in which 82% of the observations were from the SAFG radiometers, 6% from the ARM radiometers, 7% from the SPO radiometers, and 5% from climatology.

3.2. Snow Depth

[29] The snow depth on thick ice is taken from the measurements of the SHEBA Snow and Ice Studies Group (SSISG). The measurements from all of the snow depth measurement lines on thick ice were combined in a 5-day running mean. The snow depth measurements came from the Snow and Ice Studies CD-ROM [Perovich et al., 1999].

[30] We use a simple procedure to estimate the snow depth on thin ice based on the accumulation rates seen in the measured snow depths. The snowfall rates measured at the camp are not used because of the uncertainties in converting these values into snow accumulation. The snow depth is estimated as a function of ice age for each of 15 ice age categories (from 1 to 400 days). Categories that are older than the time since the start of the accumulation season (15 August) are given the mean depth measured at the camp. Younger categories accumulate snow at the rate of the change of the measured snow depth except that ice younger than 60 days does not have the snow depth reduced during the winter when the measured snow depth falls (presumably because of settling and scouring). In the spring, all ice age categories have the snow depth reduced at the same rate as that measured on the thick ice. Thus the thin ice snow depth is reduced to near zero before that of the thick ice because the snow was thinner to start with. The minimum snow depth is 1 cm. This thin snow layer represents a surface granular layer commonly observed on bare ice in the summer [Perovich et al., 2002]. The annual
cycle of snow density is taken from the climatology of the snow density measured at the North Pole drifting ice stations of the former Soviet Union [National Snow and Ice Data Center, 1996].

3.3. Albedo

[31] The albedo algorithm is based on the Ebert and Curry [1993] albedo parameterization with an improved pond parameterization as reported by Schramm et al. [1997]. The improvements are based on the pond albedo observations reported by Morassutti and LeDrew [1996]. The model includes five surface types (dry snow, melting snow, bare ice, melt ponds, and open water) and four spectral bands (0.25–0.69, 0.69–1.19, 1.19–2.38, and 2.38–4.00 μm). In order to determine the broadband albedo, the four bands must be averaged with appropriate spectral weighting. The spectral weighting was determined with the radiative transfer program STREAMER [Key and Schweiger, 1998]. For clear skies the weights are 0.500, 0.346, 0.144, and 0.010 and for cloudy skies they are 0.582, 0.343, 0.074, and 0.0002 (J. Key, personal communication, 2000).

[32] Both direct and diffuse albedos are included for dry snow and water but only one albedo is included for the other surface types. The albedo for open water follows the observations at the SHEBA camp of Peggau and Paulson [2001] and includes a wind speed dependence for clear skies. Their formulation is a modification of the Briegleb et al. [1986] formula that is based only on the solar zenith angle. For cloudy skies the Peggau and Paulson [2001] observation of 0.066 is used. The area-averaged albedo is then a function of the surface types present, cloud fraction, solar zenith angle, ice thickness, snow depth, melt pond fraction, and melt pond depth. Curry et al. [2002] compared the Schramm et al. [1997] parameterization as well as others to the observations made along the albedo line at the SHEBA camp and found good agreement. It was within 1 to 2% for the dry snow and ponded periods but 7% too low for melting snow and 8% too high during fall freeze up. In the simulations reported here the match with observations during the summer was best if the ice was assumed to be first-year ice for the purposes of the albedo calculations. The snow depth is set to 1 cm on 24 June and the surface is assumed to be melting throughout the melt season [Curry et al., 2002].

[33] The melt pond fraction is obtained from the aerial surveys conducted during the summer and reported by Perovich et al. [2002]. These observations are 0.05 to 0.10 lower than those of Tschudi et al. [2002]. The Perovich et al. values are used here because of the larger area surveyed (140 km² per flight vs. 43 km²) and the greater temporal coverage (June through October vs. July only). The depth of the ponds was taken to increase linearly starting on 6 June and rising to 0.5 m on 1 September (following the observations reported on the SHEBA Snow and Ice Studies CD-ROM [Perovich et al., 1999a]). Figure 2 shows the modeled albedo for 2-m ice, the observed average albedo from the albedo line, as well as the observed snow depth and pond fraction. While the estimated snow depth depends on the age of the ice, it is the same for smooth and ridged ice and the pond fractions are assumed to be the same as well. These assumptions may add additional error to the ridged ice albedo, but there are few data available for comparison.

3.4. Deformation

[34] The deformation used to force the ice redistribution model was obtained from a data fusion procedure that merged observations from buoys, the RADARSAT Geophysical Processor System (RGPS), the Advanced Very High Resolution Radiometer (AVHRR), and the Special Sensor Microwave/Imager (SSMI) to obtain daily estimates of the ice motion and deformation on the same 25-km grid, centered on the ship, that is used in the present model. The methods and results of the data fusion procedures are presented by Lindsay [2002] and are archived at the NCAR JOSS SHEBA data archive. The ice motion is derived from all four of the data sources but the ice deformation at the camp location relies principally on the RGPS displacement measurements. Note that here we used observed motion and deformation values and did not derive these quantities from a dynamic model that relies on solving a force balance equation to determine the ice motion.

[35] The motion and deformation estimates include spatial variability within the 200-km grid so that the thickness distributions calculated from them will reflect this variability. The deformations are determined for a spatial scale of 100 km. While shear plays an important role in the redistribution model, it is divergence that most directly produces

Figure 2. Snow depth, pond fraction, and albedo at the camp from observations (diamonds) and as used in the model for 2-m thick ice (lines). The snow depths are from observations along all the multiyear ice snow lines near the camp [Perovich et al., 1999a], the pond fractions are from aerial surveys [Perovich et al., 2002], and the albedo observations are from the camp 100-m albedo line [Perovich et al., 1999a].

The deformations are determined for a spatial scale of 100 km, while shear plays an important role in the redistribution model, it is divergence that most directly produces
open water, thin ice, and ridged ice. Figure 3 shows the time series of the relative area of a cell centered at the camp. There was significant divergence in the early part of the winter, followed by steady convergence during the spring and summer. A large divergence/convergence event occurred in early August. The net area change at the end of the year was a net increase of 12%.

4. Data Assimilation

[36] In order to improve the estimates of the ice thickness, it can be advantageous to assimilate observations that relate to the ice thickness distribution. These observations might include only the open water fraction or may include all or part of the full distribution. Some interesting issues arise with assimilation of ice thickness or open water observations. Because the thickness is represented by a probability distribution function (PDF), a change in the fractional area of any one thickness or bin must change the areas of one or more of the other bins in the distribution. It is possible to formulate a method that relies strictly on the error covariance properties of the model and the observations, as is done below. However, the method can occasionally lead to strange results because the error covariance structure of the model is poorly known.

[37] The model error covariance depends on the sources of the errors. Errors in the model thickness distribution can arise from errors either in the forcing parameters or in the model physics. Without extensive validation data, disentangling the two is difficult at best. First consider only observations of the open water fraction and errors in the forcings. If the prescribed divergence rates are incorrect, any discrepancy in the model and observed open water fraction might be best understood as arising from an incorrect estimation of the changes in the area of a cell, and any adjustments in the distribution should be shared equally by all elements of the distribution. For example, if the observed open water fraction is less than what is estimated by the model, the revised model estimate of the open water should be reduced, and all other classes should be increased in proportion to their fractional areas and the size of the discrepancy. Because the area of thick ice is increased, the estimated mean thickness increases, an increase that can be substantial if there is a large discrepancy. On the other hand, if it is assumed that the discrepancy comes from incorrect thermal forcing parameters or growth rate calculations, it may make more sense to modify the thinnest part of the distribution, creating ice just at the thin end of the distribution and thereby minimize the changes in the mean ice thickness. Another alternative is to create ice in the model of a single thickness, a thickness that would be thick enough to persist during the melt season and provide a good match to subsequent observations of open water. This would both minimize the changes in the model ice volume and the jumps in the model open water fraction. This last alternative is used here and is explained further below.

[38] For SHEBA, the observations available include submarine, aircraft, and satellite estimates of the open water fraction, of just the thin ice part of the thickness distribution, or of the full distribution. The procedure used here depends heavily on an assessment of the errors of both the model and the observations. These errors are often poorly known and require some broad assumptions but the method may prove increasingly useful as we gain knowledge of these errors. Each time ice observations are assimilated into the model, ice must be created or removed to account for the new information. The resultant change in the ice volume is a measure of the agreement between the observations and the model and serves as a validation (or repudiation) of the model calculations.

4.1. Assimilation Procedure Using Error Covariances

[39] The assimilation procedure begins with estimates of the distribution for some set of fixed thickness categories. Let the model distribution be \( g_{\text{mod}}(h) \) and the observed distribution be \( g_{\text{obs}}(h) \). They are typically defined on different sets of characteristics with thickness \( h \) and each has an associated uncertainty (error standard deviation) of \( \epsilon_{\text{mod}}(h) \) and \( \epsilon_{\text{obs}}(h) \). The observations need not include the entire distribution and, in the case of ice concentration observations, only the open water characteristic \( g_{\text{obs}}(0) \) is present.

[40] First both the model and the observed distributions are determined for a set of \( n_h = 50 \) fixed thickness bins. Then the model and observed fractions are merged with a procedure related to the Kalman filter (KF) [Dee, 1991; Thomas and Rothrock, 1993]. Let \( F \) be a column vector of length \( n_h \) consisting of the fractional area coverage of each fixed bin from the model. Let \( P \) be the \((n_h \times n_h)\) error covariance matrix for \( F \) and let \( Z \) be the vector of observed fractions with measurement error covariance matrix \( R \). In order to account for the hard constraint that the fractional areas remain a probability distribution function, an additional pseudo-measurement is added to the observation vector that represents the sum of all the fractions. This sum must be exactly 1 and the error is zero. The observation vector can be estimated from the model values of \( g \) with a linear operator. This estimate is

\[
Z = HF. \tag{12}
\]

For example, if \( n_h = 3 \),

\[
H = \begin{bmatrix}
1 & 0 & 0 \\
0 & 1 & 0 \\
0 & 0 & 1 \\
1 & 1 & 1
\end{bmatrix}. \tag{13}
\]
where the final row of ones accounts for the pseudo-measurement. The measurement error covariance matrix is
\[
R_z = \begin{bmatrix}
\varepsilon_{ew}^2 & 0 & 0 & 0 \\
0 & \varepsilon_1^2 & 0 & 0 \\
0 & 0 & \varepsilon_3^2 & 0 \\
0 & 0 & 0 & 0
\end{bmatrix}.
\] (14)

If only the open water fraction is observed, the errors for the other bins are set to large numbers. For simplicity the errors are assumed to be independent but in fact there may be significant correlations due to the PDF constraint. The KF update step yields a refined estimate of the area fractions, the vector \( \mathbf{G} \), that is based on the difference between the observations and the model-based estimate of the observations,
\[
\mathbf{G} = \mathbf{F} + \mathbf{K} \left( \mathbf{Z} - \mathbf{\hat{Z}} \right).
\] (15)
The Kalman gain matrix is
\[
\mathbf{K} = \mathbf{P}_F \mathbf{H}^T \left[ \mathbf{H} \mathbf{P}_F \mathbf{H}^T + R_z \right]^{-1}
\] (16)
and the estimate of the error covariance matrix of the new model values is updated by
\[
\mathbf{P}_\mathbf{G} = (\mathbf{I} - \mathbf{K} \mathbf{H}) \mathbf{P}_F.
\] (17)
where \( \mathbf{I} \) is the identity matrix. In equation (15) the updated ice thickness distribution vector \( \mathbf{G} \) is a weighted combination of the first guess vector \( \mathbf{F} \) and the measurement vector \( \mathbf{Z} \). Finally the ice in each fixed thickness bin is proportionately distributed among the characteristics that originally formed the bin.

4.2. Model Errors

[41] An important and perhaps unique aspect of this model is that estimates of the uncertainty of both the ice thickness of each characteristic \( h_i \) and the area fraction \( g(h) \) are computed every time step. The methods of estimating these errors are approximate and the efficacy is still not well known. For this study, the uncertainty in the thickness of a generic characteristic is based on the simple assumption that the growth in the uncertainty of a parameter is linearly related to the rate of change of the parameter, and that the growth in the uncertainty does not fall below a minimum value, either absolute, in the case of \( h \), or fractional, in the case of \( g \).

[42] For example, the uncertainty in the thickness at time step \( t+1 \)
\[
\varepsilon_{h,t+1} = \sqrt{\varepsilon_{h,t}^2 + (\varepsilon_{f,\min} + \varepsilon_f g')^2 (\Delta t)^2}
\] (18)
where \( \varepsilon_h \) is the uncertainty in the thickness of a characteristic, \( \varepsilon_{f,\min} \) is the minimum (absolute) rate of uncertainty increase, \( g' \) is the growth rate, and \( \varepsilon_f \) is the growth uncertainty. The constant \( \varepsilon_f \) is set at 0.20, reflecting an assumed uncertainty of 20% in the growth rate, and the minimum is \( \varepsilon_{f,\min} = 1 \text{ mm day}^{-1} \).

\[\text{[41]}\] The uncertainty in the area fraction is related to the uncertainty in the forcing deformation (divergence and shear) and in the ridging parameters \( G^* \) and \( k \). These sources are lumped into a minimum (fractional) growth rate \( \varepsilon_{g,\min} \) and a redistribution uncertainty
\[
\varepsilon_{g,t+1} = \sqrt{\varepsilon_{g,t}^2 + (\varepsilon_g |g_{t+1} - g_t|)^2 (\Delta t)^2}
\] (19)
where \( g_t = g(h) \) at time \( t \). The uncertainty growth rate factors are set at \( \varepsilon_{g,\min} = 0.001 \text{ day}^{-1} \) and \( \varepsilon_g = 0.20 \text{ day}^{-1} \). These factors, for both \( h \) and \( g \), are highly uncertain and future research with sufficient validation data will be needed to determine them more accurately.

4.3. Assimilation of Open Water Measurements

[44] Many of the observations are of just the open water fraction. The simplest method to deal with these observations is to determine the new open water fraction as a weighted sum of the model and the observation values
\[
g(0) = \frac{g_{\text{mod}}(0) + g_{\text{obs}}(0)}{\varepsilon_{g,\text{mod}}(0)^2 + \varepsilon_{g,\text{obs}}(0)^2},
\] (20)
for which a value of \( \varepsilon_{g,\text{mod}}(0) = 0.01 \) was used. If the model fraction of open water was less than the observed, the thinnest ice characteristics were removed to accommodate the increased open water fraction. If the model open water fraction was larger than the observed, a single characteristic was created in the model with a thickness of 0.5 m. This thickness was found to provide good continuity in the model fraction of open water from one observation to the next.

4.4. Observations of Ice Thickness and Open Water

[45] We used four different types of ice thickness observations: a submarine survey of the total distribution in the fall, Twin Otter aircraft surveys of the thin ice distribution in the winter, C130 aircraft surveys of lead fraction in spring, and helicopter and satellite surveys of the open water fraction in the summer. SSMI-based estimates of the open water fraction were found to be too uncertain to be useful in this application. In another context, however, where the ice edge is included in the domain, these satellite measurements could be helpful. Table 1 show the dates and the total length or area of each of the surveys.

4.4.1. SCICEX

[46] Two surveys were made during the SHEBA experiment, one at the end of September 1997 and a second in August 1998. In the first survey the ice draft near the camp was measured by the nuclear submarine USS Archerfish on 28 September through 1 October 1997 as part of the Scientific Ice Expeditions program (SCICEX). The survey consisted of a 150-km wide star-shaped pattern that was over 1300 km in total length. The distribution was estimated in 20-cm bins. The average thickness was 1.54 m with a standard deviation of 1.35 m and a maximum ridge thickness of 20 m. Ten percent of the area was covered by open water or ice less than 20 cm thick. The uncertainty in the mean ice thickness due to sampling errors is 0.10 m or less (Y. Yu, personal communication, 2000). The results of the September survey were used as the initial thickness distri-
bution in the model and an initial uncertainty in the fractional area of each bin was taken as 0.01. This uniform uncertainty value means the thick bins of ridged ice with very little fractional area are poorly measured while the bins near the mode are well measured. The results of the second survey are not yet available but when they are released they will be a valuable validation and assimilation data set.

[47] The initial thickness distribution was assumed to be all smooth, multiyear ice. This was done because it is not possible to determine just what part of the distribution is newly frozen ice or what part is ridged ice. The two categories, ridged and smooth, are treated the same within the model. The identification of the initial ice as multiyear is a convenient way to mark the initial distribution, since most of it is multiyear ice, and is not meant to be a definitive identification of the ice type.

4.4.2. Twin Otter Surface Properties Surveys

[48] Ice thickness distributions were derived from aircraft surveys performed by the Twin Otter logistics aircraft [Lindsay and Stern, 1999]. The surveys used thermal infrared measurements of the ice surface along the track of the aircraft to determine the surface temperature distribution, and an energy balance model was used to estimate the thickness distribution for ice up to 0.50 m thick. The instrument was a narrow beam KT-19 infrared radiometer. Here we used revised results based on new calibrations of the radiometer and a revised estimate of the snow depth and conductivity. Each survey was centered on the ship and ranged out about 30 km from the ship. Five campaigns were used: 14 October 1997, 25–27 November 1997, 8–9 January 1998, 10–12 March 1998, and 29 March 1998.

4.4.3. NCAR C-130

[49] Airborne surface surveys were also conducted with the NCAR C-130, based in Fairbanks, AK. Seven surveys conducted in May 1998 are analyzed by Tschudi et al. [2001]. The open water fraction was determined with the passive microwave Airborne Imaging Microwave Radiometer (AIMR) and ranged from 0.0040 to 0.027. An uncertainty of 10% of the measured fraction has been assumed. In a second campaign in July, a video camera was used to measure the open water fraction on five days [Tschudi et al., 2002]. The open water fraction was 0.05 to 0.09 at this time. The uncertainty, determined from the standard deviation of the 250 images analyzed for each flight, averaged 7% of the mean.

4.4.4. Helicopter Surveys

[50] Aerial surveys of the ice were also conducted during the summer from a ship-based helicopter within a 50-km region centered on the ship. The results are reported by Perovich et al. [2002]. Images were captured with a 35-mm camera and were subsequently analyzed digitally. The area fractions of ice, ponds, leads, and new young ice were determined with an estimated uncertainty of 2% of the mean fraction. This uncertainty is based on the sampling variability from individual photographs.

[51] Perovich et al. [2002] also report the analysis of a single U.S. National Reconnaissance (USNR) satellite image from 18 June 1998. The image is black and white, covers an area of 57 km², and has a resolution of 1 m. The open water fraction from this image is 0.031. Since the area sampled is about 1/3 of that for the helicopter surveys, the assumed error is $\sqrt{3}$ larger, or $\sqrt{3} \times 0.02 \times 0.031 = 0.001$.

4.5. With and Without Data Assimilation

[52] Figure 4 shows the evolution of the mean ice thickness, total ice volume, and the open water fraction with and without data assimilation. Most of the open water measurements had relatively low uncertainty, so the model estimates were forced to nearly match the measurements. In general, the observations indicated less open water or thin ice than was represented in the model so that the assimilation procedures increased the amount of ice. All of the surveys from the Twin Otter showed less thin ice than estimated by the model so that in each case the ice volume increased.

![Figure 4](image_url)

**Figure 4.** Time series comparing runs with and without data assimilation of ice thickness and open water observations: (a) the total ice volume in the cell centered on the camp with an initial area of 625 km², (b) the mean ice thickness, and (c) the open water fraction. The diamonds indicate observations of the open water fraction.
Observations from most of the different platforms also showed less open water than found in the model-only simulation. This is most evident in June and July but an observation in August showed more open water than the model estimates and consequently the ice volume decreased. The over abundance of open water in the model in June and July would indicate that either the melt rates for thin ice were too high or that more convergence occurred during this period than was estimated from the ice displacement measurements (Figure 3).

5. Results
5.1. Comparisons of Thermal Model Output to Camp Observations

The surface and bottom energy fluxes calculated for 2-m ice are shown as time series in Figure 5. The annual cycle is similar to what has been found by a number of investigators [e.g., Maykut, 1982] and follows the pattern of the energy balance found for 45 years of Soviet drifting ice stations [Lindsay, 1998]. A number of comparisons can be made between values calculated by the model and those measured at the SHEBA camp. The sensible heat flux and the net radiation were measured at the 20-m tower by the SHEBA Atmospheric Surface Flux Group. Six-hourly average values are compared in Figure 6. The model shows very little bias in the sensible heat flux, just 1.3 W m$^{-2}$, and the correlation is good, $R = 0.86$. However, the model overestimates the magnitude of some of the large positive or negative fluxes. The net radiation is also in good agreement with observations. However, the largest discrepancies occur during the summer, when there are small differences between the local albedo measured at the tower and the model albedo. The bias over the year is small, the
model averages $3.2$ W m$^{-2}$ greater, and the correlation is $R = 0.95$.

Figure 7 shows the cumulative ice production rates for five different fixed thickness ice classes. These plots illustrate the net production obtained by integrating the growth rate $f$ computed for each ice class over the year. Ice less than 1 m shows a net gain in ice production while ice 1 m or thicker shows a net loss under the conditions seen at SHEBA. These plots are in general agreement with the measurements made at the camp, which indicate that ice initially between 0.89 and 2.90 m grew by 0.50 to 0.75 m in the winter and showed a net loss of 0.10 to 0.75 m by the end of the summer [Perovich et al., 1999b]. The shapes of the curves are also similar, showing a gradual rise in thickness during the winter followed by a sharp drop in the summer season. The locations at the camp that had small growth rates were where depressions left by old melt ponds allowed the accumulation of more snow than average.

5.2. Ice Thickness Distribution

[55] The ice thickness distribution is derived from the characteristics. The evolution of the thickness of the individual smooth or ridged characteristics $h(t)$ is shown in Figure 8. Initially all of the ice is labeled as smooth, multiyear ice (although some is undoubtedly new or ridged) but as the ice deforms new smooth ice and ridged ice characteristics are created. New smooth ice characteristics are created whenever there is divergence and new ridged ice characteristics are created during convergence and shear. Characteristics disappear if all of the ice is melted or ridged, and characteristics merge when too many are created. The new smooth ice characteristics are seen to be frequently created in the winter but are then lost to ridging and, in the summer, to melt. New smooth ice characteristics, each with an initial thickness of 0.5 m, are created in the summer to account for the assimilation of open water observations. None are left by the end of August. The thickness resolution is seen to diminish for the thicker ice because the merging scheme favors higher resolution for thin ice. Rridged characteristics are created throughout the year but the thickness of the newly ridged ice increases as the fraction of thin ice is reduced by growth and by previous ridging events.

[56] Figure 9 shows the distribution of the model ice thickness on four dates: at the beginning of the experiment, in the early summer when the ice is the thickest, in the late summer when it is near the thinnest, and at the end of the

Figure 6. Comparison of the modeled and observed sensible heat flux and net radiation for 6-hour averages. The observations are from the SHEBA Atmospheric Surface Flux Group tower. For the (top) sensible heat flux the model averages just 1.3 W m$^{-2}$ higher than the observation, but the model overestimates the magnitude of the large positive or negative fluxes. The correlation is $R = 0.86$. For the (bottom) net radiation the model averages 3.1 W m$^{-2}$ higher, and the correlation is $R = 0.95$. The solid lines indicate the best fit lines.
experiment. Note that the mode increases from the initial value near 1 m to more than 2 m in June, and then declines to less than 1.5 m in August. There is very little thin ice present in the summer. The distribution is significantly different at the end of the experiment compared to the beginning. There is more very thick ridged ice at the end of the year and most of the thin ice is less than 0.5 m thick while at the beginning the distribution between 0 and 1 m is more uniform. At the end there are two principal modes, one at about 0.25 m representing new ice and one at 1.0 to 1.5 m representing multiyear and ridged ice.

[57] The fraction of the area in 7 fixed ice thickness bins and for open water is shown in Figure 10. Here again the dominant mode in the multiyear ice gradually moves, first from the 1–2-m bin to the 2–3-m bin during the winter and then returns to the thinner bin in the summer. Sharp jumps in the fraction in a bin occur when an important characteristic crosses from one bin to another. The jumps in the fractions of the thicker ice during the summer are artifacts of the data assimilation procedures. At the end of the year most of the ice greater than 3.0 m is ridged while the dominant mode between 1.0 m and 2.0 m is mostly multiyear ice.

[58] The total fractions of multiyear ice, new smooth ice, new ridged ice, and open water are shown in Figure 11 as

Figure 8. Time series of the thickness of individual characteristics for (top) multiyear and smooth and (bottom) ridged ice. Multiyear ice is orange, new smooth ice is blue, and ridged ice is red.

Figure 9. The distribution of ice thickness at four times during the year: the initial distribution (2 October 1997), 17 June (when the mean thickness was near the maximum), 8 August (when it was near the summer minimum), and the final distribution (6 October 1998). Green is open water ($h < 0.1$ m), orange is multiyear ice, blue is new smooth ice, and red is new ridged ice. The bar at 8 m represents ice 8-m thick or more.

Figure 10. Fractional area of the ice in each of seven fixed ice thickness bins and of open water. The numbers in the upper right of each plot show the thickness range for the bin, and the number at the left shows the maximum fractional coverage of the bin. Multiyear ice is orange, new smooth ice is blue, and ridged ice is red.
well as the mean thickness of each category. The multiyear ice is assumed to be all the ice present at the beginning of the year, as measured by the SCICEX survey. The survey indicated about 10% of the area had ice less than 0.20 m thick. This ice is divided evenly between open water and ice 0.1 m thick. The 5% of open water immediately turns to new smooth ice. The multiyear fraction decreases because of divergence in the fall and to ridging in the remainder of the year. The large decrease in the multiyear fraction in the fall may be a result of ridging of new ice that formed in the fall before the SCICEX survey. The model fraction of ridged ice grows steadily throughout the year, finally covering 45% of the area. Ridged ice accumulates even in the fall when there was net divergence because the redistribution function makes ridged ice from strong shear even in the presence of modest divergence. The new smooth ice covers 25% of the area in January but by the end of the summer there is none left.

The changes in the fractions of the different ice types or thicknesses are, for the most part, due to the changing deformation rates. For example the sharp rise in the fraction of new smooth ice in mid-January is due to an episode of divergence. This is followed by a period of convergence as the new smooth ice fraction decreases again and the ridged fraction increases sharply. The sharp drop in the mean thickness in August is caused by divergence, not melt, and the mean thickness increases again when the subsequent convergence episode occurs.

5.3. Aggregate Energy Fluxes

The aggregate fluxes are determined by an areal weighting of the fluxes for individual ice thickness classes. In order to determine the relative importance of each thickness class in establishing the areal average, the monthly average for each fixed thickness class is plotted in Figure 12. The width of each bar is proportional to the fraction of the area covered by each class. Two dominant terms of the surface energy balance are plotted: the net radiation and the sensible heat flux. Positive values indicate heat gain by the ice. Also plotted is the average ice growth rate for each month. The aggregate net radiation is positive only in May through August. The very thin ice shows more radiative heat loss in the winter and heat gain in the summer than the thick ice classes. The thin ice classes also show substantial sensible negative flux in the winter while for thicker ice...
the flux is positive. The aggregate net sensible heat flux is near zero except for February and March, when there is very little thin ice, and July, when the air temperature was commonly above freezing. In winter the thin ice, while occupying a small amount of the area, plays a significant role in establishing the average growth rate, similar to what Maykut [1982] found. In summer the melt rate is similar for all of the ice categories and the aggregate melt is largest in July.

5.4. Spatial Variability

[62] There was considerable spatial variability in ice deformation experienced in the region surrounding the ship. This variation results in substantial variability in the model mean ice thickness at the end of the experiment. The thermal forcing and thickness-dependent growth rates are assumed to be the same as the values calculated at the camp, while the divergence and shear includes spatial variability. Figure 13 shows the mean ice thickness at the end of the experiment. Within this 200-km region the simulated ice thickness ranges between 1.7 and 2.5 m with the thickest ice to the north of the camp. The ice to the west and south of the camp is substantially thinner than the ice at the camp itself. Note that the uncertainty in the deformation estimates increases with the distance from the camp because of the location of buoy trajectories (used when RGPS data are unavailable) and hence the uncertainty in the estimated ice thickness also increases.

6. Comments and Conclusions

[63] All efforts to model the ice thickness distribution face formidable obstacles in obtaining accurate estimates of the distribution and the present study is no exception. The thermal forcing parameters are reasonably well measured: wind speed, air temperature, and downwelling radiative fluxes. The humidity proved hard to measure in the field, so the latent heat flux over the thick ice is not well known, however it is likely small compared to the flux over thin ice or to the net radiative flux. The snow depth on multiyear ice was well measured at the camp but few measurements have been reported for the depth on new ice; the snow on thin ice can greatly modify the ice growth rates. The model albedo matches the observed albedo reasonably well, however the albedo is very sensitive to the snow depth on thin ice and to the pond fraction in the summer.

[64] There are a number of parameters in the model that are not well known. Perhaps the ridging factor \( k \) and the fraction of area that participates in ridging \( G^* \) are most important for the thickness distribution. The relative importance of rafting and ridging under different conditions of ice thickness and deformation help determine \( k \). Our model simulations suggest that uncertainties in these parameters may account for a 10–20% uncertainty in the final mean ice thickness.

[65] The fraction of the energy flux from the ocean that goes to lateral melt is also poorly known. Steele [1992] shows how it is related to the floe size distribution but the floe size distribution is not computed in this model. As mentioned above, the selection of this fraction has only a modest influence on the mean ice thickness at the end of the year because changes in the open water fraction are constrained by observations and because the positive feedback loop of lateral melt increasing open water and lowering the aggregate albedo is not present in this model. It also does not greatly change the size of the corrections made for the observed open water fraction during data assimilation. Decreasing the fraction increases the melt rate of the ice but does not have a large effect on the model open water fraction.

[66] The simulated ice thickness increases 0.90 m over the course of the year, from 1.54 m to 2.44 m. What explains the increase in mean ice thickness when Perovich et al. [1999a] report a net loss of ice at all of the mass balance measurement sites? The reason is found in the relative area plot in Figure 3. In the fall there was significant opening, more than 40%, allowing for high rates of ice production in the newly opened leads. Figure 4 shows that during the fall the total ice volume was increasing while the mean ice thickness remained nearly constant. During the spring and summer there was significant closing, creating a great deal of thick ridged ice (Figure 11). The net area change over the year near the camp was small, on the order of 10% opening. The opening in the fall combined with the steady closing in the spring accounts for the increase in the mean thickness. The uncertainty in the final mean model ice thickness cannot be determined without validation data but it is likely about 0.5 m, well below the change in the simulated ice thickness. This uncertainty arises from uncertainty in
the forcing (most importantly the deformation rates) and from the model characterization of physical processes (most importantly the ridging process).

[67] It is also important to remember how unusual the ice pack appeared to the scientists who first deployed the ice camp in the fall of 1997. Most of the ice was very thin and there was considerable concern that a suitably thick ice floe could be found for the camp (R. Anderson, personal communication, 2002). Upper ocean salinity measurements also indicated the water was unusually fresh at the time of the camp deployment, a result attributed to an anomalous degree of melt the previous summer [McPhee et al., 1998]. Passive microwave observations indicated a great deal of open water along the Alaskan coast in September and that the camp was deployed within 200 km of the edge of the perennial ice. The initial mean ice thickness, just 1.54 m, is consistent with these observations. The final position of the camp was over 500 km farther north and in a region of consolidated pack ice, although the summer ice edge did get within 300 km of the camp in another summer of extraordinary melt. The unusual initial conditions, high rates of ice production in the fall, and the steady convergence through the rest of the year provided the conditions for a net increase in the mean ice thickness and a marked change in the nature of the thickness distribution with much more very thick ridged ice and less thin ice.

[68] There is considerable spatial variability in the deformation and the net area changes so that the net change in the ice thickness in the 200-km region analyzed here ranged from −41% to +7% compared to that of the camp (Figure 13). The ice deformation data shows that the changes in the relative areas computed from the divergence ranged from between 10% to 130%. This very large spatial variability highlights the difficulty that will be encountered when local measurements of the ice thickness are compared to model simulations. The ice thickness distribution is very sensitive to the nature of the deformation (net opening or net closing) and the season in which it occurred.

[69] A new method for assimilating ice thickness information into a model has been presented that depends heavily, as do all assimilation procedures, on estimates of the uncertainties in the model-estimated distribution. These uncertainties determine how ice is created or removed in the model to accommodate observations, observations that may indicate lesser or greater amounts of thin ice or open water are present. Here it is assumed that most of the change in the distribution is in the thinnest ice classes, so that changes in the total volume of ice are minimized. When ice in the model must be created, a thickness of 0.5 m is used, a value chosen to minimize both the size of the jumps in the simulated open water fraction and the changes in the mean ice thickness caused by the assimilation.

[70] Finally, it should be noted that the ice draft data from the second SCICEX cruise in August 1998 will eventually become available. These data will be useful in determining if the model is correct in its assumptions for this time and location, however it will be a challenge to attribute discrepancies between the model results and the SCICEX survey results to a single source of error in the forcing parameters or to a single model parameter because of the large number of parameters in the model and the long time period between the two thickness surveys.

[71] The numerical results of these model simulations are posted in the SHEBA data archive maintained by the NCAR JOSS (web site: www.joss.ucar.edu/codiac/codiac-www.html). Daily estimates of the forcing parameters, the terms of the ice energy balance, and the thickness distribution are included for the camp location.

[72] Acknowledgments. A large number of people have helped to collect and process data used in this study. I thank our colleagues in the SHEBA Atmospheric Surface Flux Group, Ed Andrews, Chris Fairall, Peter Guest, and Ola Persson, for collecting and processing the flux tower data. I also thank Don Perovich and the SHEBA Snow and Ice Studies Group for albedo, ice temperature, and pond fraction measurements; Dick Moritz and all of the SHEBA Project Office Observers for the SPO data; Bernie Zack and the Atmospheric Radiation Measurement Program personnel for their radiative flux observations; Miles McPhee, Tom Stanton, Roger Anderson, and Jamie Morison for ocean temperature data; Mark Tschudi and the NCAR C-130 team for pond fraction data; Ron Kwok and the Polar Remote Sensing Group at JPL for the RGPS data; and Yanling Yu and Drew Rothrock of the Polar Science Center for the SCICEX ice draft data. This work has been funded by the National Science Foundation Arctic System Science Program.

References


Lindsay, R. W., Temporal variability of the energy balance of thick Arctic pack ice, J. Clim., 11, 313–331, 1998.


