ON MODELING THE WEDDELL SEA PACK ICE

by

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ABSTRACT

Some results from a dynamic-thermodynamic simulation of the seasonal cycle of the Weddell Sea pack ice are described. The model used for the study is similar to that developed by Hibler (1979) for a numerical investigation of the Arctic ice cover. It employs a plastic ice rheology coupled to a two-level ice-thickness distribution. The thickness characteristics evolve in response to ice dynamics, and to ice growth and decay rates dictated by surface heat-budget calculations and by heat storage in a fixed depth oceanic boundary layer. Observed time-varying wind, temperature, and humidity fields based on the 1979 Australian analysis are used together with empirical radiation fields and fixed ocean currents to drive the model. Employing these fields, the model is integrated over two seasonal cycles. The second-year results are compared to observed ice-drift data obtained by Ackley (1981[b]) and to ice-edge characteristics determined from satellite imagery. In analyzing the results, particular attention is paid to ascertaining the relative roles of ice dynamics and thermodynamics in determining the ice-edge characteristics. A sensitivity analysis of the effect of a different parameterization of the atmospheric temperature and humidity fields is also carried out.

Overall, the results suggest that (1) ice dynamics are essential in describing the seasonal cycle, and (2) a feedback between the atmospheric temperature and the presence of ice may be a major cause of the rapid decay of the Antarctic ice cover during the spring-summer period.

1. INTRODUCTION

Pack-ice growth and decay are controlled by both the dynamic and thermodynamic characteristics of the ice. Owing to the unconstrained nature of the Antarctic sea-ice cover, it is likely that dynamic effects play a dominant role in its behavior. The Weddell Sea pack ice comprises an important portion of the Antarctic ice cover and has both representative and unique characteristics. The eastern portion tends to remain during the spring and summer. High productivity is often associated with such a residual ice edge, making sea-ice fluctuations particularly relevant to biological activity (Ackley and others 1979, Alexander 1980).

In addition to these features, sea-ice drift and air-temperature data from drifting buoys (Ackley 1981[b]), make the Weddell Sea a particularly useful region for a first test of an Antarctic ice model. In this paper we examine some initial results from a seasonal dynamic-thermodynamic simulation of the Weddell Sea pack ice. In these simulations we make use of an extension of the dynamic-thermodynamic sea-ice model developed by Hibler (1979). This study differs from previous seasonal Antarctic simulations (Pease 1975, Washington and others 1976, Parkinson and Washington 1979) in two important respects: (1) real-time varying forcing is used to drive the model instead of climatological data, and (2) a realistic rheology and thickness coupling is used to describe the ice dynamics instead of either no ice motion or ad hoc ice-velocity corrections without reference to an ice rheology.

2. DESCRIPTION OF THE MODEL

Basically the model consists of a momentum balance coupled to equations describing the evolution of ice thickness characteristics. Except for the surface heat budget (Hibler 1980[b]) a complete description of these equations is given in Hibler (1979). We summarize briefly the main characteristics here. In Cartesian coordinates the governing equations are:

\[
\begin{align*}
\frac{D\mathbf{u}}{Dt} &= m\mathbf{f} + \mathbf{J}_{\alpha} + \mathbf{j}_W - mg\mathbf{V} + \mathbf{F}, \\
\frac{Dh}{Dt} &= -\nabla \cdot (uh) + \mathbf{S}_h, \\
\frac{D\mathbf{A}}{Dt} &= -\nabla \cdot (u\mathbf{A}) + \mathbf{S}_A,
\end{align*}
\]

where \(D/Dt = \partial/\partial t + \mathbf{u} \cdot \nabla\) is the total time derivative, \(\mathbf{u}\) a unit vector normal to the surface, \(\mathbf{f}\) the Coriolis parameter, \(m\) the ice mass per unit area, \(\mathbf{J}_\alpha\) and \(\mathbf{j}_W\) forces due to air and water stress, \(H\) the sea-surface dynamic height, \(g\) the acceleration due to gravity, \(A\) the fraction of area covered by ice, \(h\) the average areal ice thickness (equal to the ice-mass per unit area divided by the ice density), and \(F\) the force due to variations in internal ice stress. \(S_A\) and \(S_h\) are thermodynamic growth and decay terms and are described below. For modeling the ice interaction the ice is considered to have a nonlinear viscous-plastic constitutive law \(\sigma_{ij} = \alpha_{ij}(\mathbf{\varepsilon}_{ij}, P^*)\) where \(\sigma_{ij}\) is the two-dimensional stress tensor, \(\mathbf{\varepsilon}_{ij}\) the strain-rate tensor, and \(P^*\)

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the ice strength. The constitutive law employs an elliptical yield curve passing through the origin with a no-stress condition applying for pure divergence (see Hibler (1977) for exact equations). Rigid plastic behavior is approximated in the rheology by allowing the ice to flow plastically for normal strain-rates and to creep in a linear viscous manner for very small strain-rates. The viscosity used for small strain-rates is, however, very large. As a consequence, in the linear regime, simulated deformation rates are small enough to approximate rigid behavior well. The coupling of the ice strength to the ice-thickness characteristics is given by

\[ p^* = P_0 he \]

where \( P_0 \) and \( C \) are strength constants.

This formulation makes the ice strength \( p^* \) strongly dependent on the amount of thin ice while also allowing the ice to strengthen as it becomes thicker. For the calculations here we take

\[ P_0 = 2.75 \times 10^4 \text{ N m}^{-2} \quad \text{and} \quad C = 20. \]

This \( P_0 \) value is about five times larger than the \( P_0 \) value used by Hibler (1979). This increase was necessitated by the fact that daily wind fields were used here as opposed to the eight-day average wind fields used by Hibler (1979). On the average, these daily wind fields yielded higher wind stresses than the eight-day averaged data. As a consequence a higher ice strength was needed to obtain realistic drift rates. The value of \( p^* \) used here is the same as the best fit value obtained by Hibler and Walsh (to be published) for the Arctic.

The thermodynamic terms \( S_h \) and \( S_A \) are given by

\[ S_h = f(h) f(A) + (1 - A) f(0), \]

\[ S_A = \begin{cases} 
(f(0)/h_0)(1 - A), & \text{if } f(0) > 0, \\
0, & \text{if } f(0) < 0, \\
0, & \text{if } S_A > 0, \\
(A/2h)S_h & \text{if } S_A < 0,
\end{cases} \]

with \( f(h) \) the growth rate of ice of thickness \( h \), and \( h_0 \) a fixed demarcation thickness between thin and thick ice. The \( S_h \) term gives the net growth or melt of ice while \( S_A \) characterizes the way in which growth and decay change the relative areal extents of open water and ice. The \( h_0 \) parameter dictates how rapidly open water is removed by growth. A more complete discussion of these equations is given by Hibler (1979).

For these Antarctic simulations \( h_0 \) is set equal to 1.0, as opposed to 0.5 used in Arctic simulations (Hibler 1979, Hibler and Walsh to be published). This larger value is based on field observations reported in Ackley and others (1980) and Gow and others (1982) which suggest that frazil ice formation may well progress into ice production in open-water regions. Other than \( h_0 \) and \( P_0 \), the numerical parameters used in the simulations are identical to those used by Hibler (1979).

The vertical growth rates \( f(h) \) are estimated by employing Sermter's (1976) sea-ice thermodynamic model in conjunction with a surface heat-budget calculation similar to that of Parkinson and Washington (1979) and Manabe and others (1979). The basic surface energy-budget calculation (for details see Hibler 1980[b]: appendix B) includes incoming shortwave and long-wave radiation and sensible and latent heat fluxes, and provides for the determination of the surface temperature of the ice by numerical iteration.

Following Manabe and others (1979), the effects of snow cover are approximated by allowing the ice surface albedo to be that of snow when the calculated surface temperature is below freezing, and that of snow-free ice when the surface temperature of the ice is at the melting point. The model also includes a mixed layer of fixed depth (30 m) and assumes a constant oceanic heat flux of 2 W m\(^{-2}\) (the same value used in Arctic simulations (Hibler 1980[b]) into this mixed layer from below. Following Equation 5, when open water is absorbing heat, all of the absorbed heat is used to melt the ice further. If there is no ice the absorbed heat raises the temperature of the mixed layer. Under growth conditions, no ice is allowed to form until the mixed layer reaches the freezing temperature of sea-water.

These coupled equations are solved numerically using a time-marching procedure with one-day time steps. In this procedure the momentum equations are solved implicitly using relaxation methods, and the thickness equations are solved explicitly. More details on the numerical scheme are presented by Hibler (1979). In addition documentation for all portions of the code except the heat budget is available in Hibler (1980[a]).

3. SIMULATION RESULTS

To investigate the applicability of this model to the Weddell Sea pack ice, a series of two-year seasonal simulations employing different input values of the atmospheric temperature and humidity fields were carried out. Daily atmospheric data from 1 January 1979 to December 1979 were used to drive the model with the same atmospheric forcing being repeated the second year. In all cases the initial conditions were chosen to be 2 m of ice everywhere. After two years the model was found to be close enough to equilibrium to yield useful sensitivity analyses. More specifically, after this time the model results were found to be relatively independent of initial conditions and to vary little between sequential seasonal cycles. The particular time interval used in these experiments coincided with the deployment of several drifting buoys in the Weddell Sea (Ackley 1981[b]) which supplied valuable drift, temperature, and atmospheric-pressure data. A grid with 2º latitudinal (222 km) resolution based on an azimuthal equidistant projection (Fig.1) was used for the simulations. Zero ice velocities were assumed at all boundaries. However, at the shaded grid cells the ice strength was taken to be zero. In addition, ice is only allowed to be transferred into or out of these cells by advection and once there is considered to have flowed out of the basin. In practice these cells provide a natural outflow and/or inflow boundary condition, and are discussed in more detail by Hibler (1979, 1980[a]).
Fig. 2. Comparison of observed surface temperatures with those derived from Australian analysis temperatures at location of buoy 1433.

(a). Input fields
Input fields to drive the model consisted of average daily surface atmospheric air temperatures, mixing ratios, and surface pressure taken from the 1979 Australian analysis. Preliminary simulations with these analyzed atmospheric temperature fields yielded unrealistically small ice concentrations in summer. Comparison of analyzed surface temperatures with buoy observations (Fig. 2) indicated that the main problem was the excessively warm temperatures in the late summer and fall (February to April) in the Australian surface analysis, which did not include these buoy observations. Consequently, a more realistic surface analysis was constructed by not allowing (except for January) air temperatures and mixing ratios to rise above 271.2 K and 0.25 g kg⁻¹ in the region designated by the dashed lines in Figure 1.

Experiments with slightly different regions had little effect on the results.

For the calculation of geostrophic ocean-current fields, mean dynamic topography values reported by Gordon and others (1978) were used. In the radiation calculations, parameterizations similar to those employed by Parkinson and Washington (1979) were used. Specifically, daily global solar radiation under cloudless skies was obtained by integrating an empirical equation by Zillman (1972) over solar zenith angles for any particular day. Zenith angles at half-hour intervals for this purpose were obtained from a numerical solution of Kepler's equation (Holloway personal communication). Incoming long-wave radiation was obtained using Idso and Jackson's (1969) formula for clear skies. For cloud cover, a constant value of 0.85 was assumed at all points. This value is commensurate with maximum estimates by Van Loon (1972). Experiments with other values were found to modify the simulation results minimally.

(b). Sensitivity of ice-edge fluctuations to ice dynamics
Using the input fields discussed above, two-year simulations were carried out using either the full coupled model or, for comparison, only the thermodynamic portion of the model. Comparison of these results gives some insight into the role of ice dynamics on the ice-edge fluctuations.

Some ice-extent results from these simulations are illustrated in Figure 3, which shows plots on two specific days of the compactness for the full model and thin-ice contours for the thermodynamics-only case. The observations are based on charts of northern ice limits (prepared by the US Navy-NOAA Joint Ice Center, Suitland, Maryland, USA) which report ice having a concentration of at least 50%. In most
cases this 50% limit is effectively at the ice edge estimated in the charts. In this figure two different days were chosen to represent salient features of both the simulated and observed ice extents. Day 18 (January 18) represents a period close to the end of the rapid decay and shows a characteristic geometric tongue of ice often seen at this time of year. Day 298 (October 25), on the other hand, represents a typical maximum extent at a time just prior to the beginning of the rapid decay phase.

As can be seen from Figure 3 on day 18 the full model reproduces the tongue effect and ice extent much better than the thermodynamics-only case. Analysis of the ice-velocity field (see section (d)) shows that this tongue is largely due to eastward and northward advection of ice by the ice-velocity field. (The tongue is, however, slightly far south, partly due to insufficient transport of ice northward as discussed in the next section.) In the thermodynamics-only case the structure of the ice edge is totally incorrect and there is too much ice.

On day 298 the results are, however, quite similar, with both the full model and the thermodynamic case showing extents slightly less than observed. This in turn suggests that the expansion to maximum is largely thermodynamic in nature and controlled (in both simulations) by the temperature field.

The effect of dynamics on the seasonal variation is illustrated more graphically in Figure 4, which shows the observed time series of the total grid area covered by ice with concentrations greater than 15%. Also shown in this figure are the time series of the thermodynamics-only case where the area given is for ice of non-zero thickness. As can be seen, the coupled model with the dynamics included gives a bigger seasonal swing of extent and a more rapid decay of the ice cover. Both these effects are primarily due to the large amount of deformation and advection caused by the dynamics, which in turn create open water. As the periods of short-wave radiation increase during spring and summer this open water absorbs large amounts of heat and causes rapid decay of the ice cover. However, one difficulty with the full model (and the thermodynamics-only case) is a lack of persistence of the ice cover in the spring (day 250 on). Specifically, the simulated ice cover starts decaying sooner and in a smoother manner than observed. A feedback parameterization is proposed in the next section which modifies this feature considerably.

An additional feature of some interest is the effective area covered by ice (area x compactness) versus total area. Time series of these two quantities are compared in Figure 5 for the full model.

The difference emphasizes the large amounts of open water in the Antarctic pack ice in all seasons. This feature is in contrast to the ice charts of northern ice limits prepared by the US Navy-NOAA Joint Ice Center, but is in agreement with concentration estimates made using electically scanning microwave radiometer imagery (Zwally and others (1979) as referenced in Ackley (1981[a])

(c) An ice-atmosphere feedback mechanism

To improve the persistence of the simulated ice cover in spring, a simple feedback mechanism between ice extent and atmospheric temperature was constructed. The dynamic-thermodynamic coupled simulations including this mechanism yielded quite realistic results, and suggest that such a feedback may play much more of a role in sea-ice fluctuations than had previously been considered. However, the feedback with the thermodynamics only model produced highly unrealistic results, a fact which further indicates that realistic dynamics must be included in sea-ice models in order to obtain useful results.

The feedback mechanism consisted of specifying that the air temperatures and humidity are not allowed to rise above 271.2 K and 0.25 g kg−1 respectively if the ice-cover compactness is greater than 0.50. This correction was applied at all grid cells. The input fields used in this case were then taken to be the standard Australian analysis data without the simple re-analysis described in the previous section. It is notable that such a feedback mechanism is consistent with field observation of an abrupt change of air temperature as the ice edge is crossed. Also,
measurements of atmospheric boundary layer modification (Andreas and others 1979) support the assertion of a feedback between atmospheric temperature and ice extent.

Two-year simulations employing this feedback with both the full dynamic-thermodynamic model and a thermodynamics-only case were carried out. In the thermodynamics-only case the feedback was turned off when the ice thickness was less than 10 m. The seasonal variation of ice extent from these simulations is shown in Figure 6. As can be seen, the full model reproduces a realistic seasonal variation and geographical extent. In addition it produces the persistence in the ice extent in the spring (day 250 onward) that is in better agreement with observation than the standard case discussed above without significantly modifying the rapid decay seen in the summer period (December to January). Also, although not shown, the feedback produces a realistic ice tongue similar to the standard case discussed previously. The thermodynamics-only case, on the other hand, effectively produces a constant ice edge with very little seasonal cycle.

4. DISCUSSION

The initial simulations reported here have yielded several insights into modeling Antarctic sea-ice fluctuations. The main effect is the role of dynamics in creating a much larger seasonal swing of sea-ice extent in agreement with observation. A second effect of the dynamics is to decrease the sensitivity of the model to feedback between ice extent and the atmosphere. This result has important implications for climate models.

For the forcing employed here it is not necessary to increase the oceanic heat flux above typical Arctic values to obtain a realistic seasonal cycle when realistic dynamics are included. This is in contrast to calculations based primarily on thermodynamic considerations by Parkinson and Washington (1979) and Gordon (1981), which have suggested the need for a larger oceanic heat flux to explain the ice decay.

Further sensitivity studies and analyses are needed to examine the behavior of this dynamic-thermodynamic model of the Weddell Sea pack ice in more detail. Such work is in progress.

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REFERENCES


Ackley S F 1981[b] Sea-ice atmosphere interactions in the Weddell Sea using drifting buoys. International Association of Hydrological Sciences Publication 131 (Symposium at Canberra 1979-Sea level, ice and climatic change): 177-191


Andreas E L, Paulson C A, Williams R M, Lindsay R W, Businger J A 1979 The turbulent heat flux from Arctic leads. Boundary Layer Meteorology 17: 57-91


Physical and structural characteristics of
Antarctic sea ice. *Annals of Glaciology* 3:
113-117

Hibler W D III 1977 A viscous sea ice law as a
stochastic average of plasticity. *Journal of
Geophysical Research* 82(27): 3932-3938

Hibler W D III 1979 A dynamic thermodynamic sea
ice model. *Journal of Physical Oceanography*
9(4): 815-846

Hibler W D III 1980[a] Documentation for a two-
level dynamic thermodynamic sea ice model.
*CRREL Special Report* 80-8

Hibler W D III 1980[b] Modeling a variable
thickness sea ice cover. *Monthly Weather
Review* 108(12): 1943-1973

Idso S B, Jackson R D 1969 Thermal radiation from
the atmosphere. *Journal of Geophysical Research*
74(23): 5397-5403

Manabe S, Bryan K, Spelman M J 1979 A global
ocean atmosphere climate model with seasonal
variation for future studies of climate sensiti-
vity. *Dynamics of Atmospheres and Oceans* 3:
393-426

Parkinson C L, Washington W M 1979 A large-scale
numerical model of sea ice. *Journal of Geophysi-
cal Research* 84(C1): 311-337

Pease C H 1975 A model for the seasonal ablation
and accretion of Antarctic sea ice. *AJDNEG
Bulletin* 29: 151-172

Schwerdtfeger W 1979 Meteorological aspects of
the drift of ice from the Weddell Sea towards the
mid-latitude westerlies. *Journal of Geophysical
Research* 84(C10): 6321-6328

Semtner A J Jr 1976 A model for the thermodynamic
growth of sea ice in numerical investigations of
379-389

Van Loon H 1972 Cloudiness and precipitation in
the southern hemisphere. In Newton C W (ed)
Meteorology of the southern hemisphere. Boston,
MA, American Meteorological Society: 101-111
(Meteorological Monographs 13 (35))

Washington W M, Semtner A J Jr, Parkinson C,
Morrison L 1976 On the development of a seasonal
change sea-ice model. *Journal of Physical
Oceanography* 6(5): 679-685

Zillman J W 1972 Solar radiation and sea-air
interaction south of Australia. In Hayes D E (ed)
Antarctic oceanology II: The Australian-New
Zealand sector: 11-40 (Antarctic Research Series
19)

Zwally H J, Parkinson F, Carsey F, Gloersen P,
Campbell W J, Ramseier R D 1979 Antarctic sea
ice variations 1973-75. *NASA Weather Climate
Review* 56: 335-340