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TECHNICAL NOTE*

Sea Ice Prediction Physics

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OPC CONTRIBUTIONS

No. 1. Burroughs, L. D., 1986: Development of Forecast Guidance for Santa Ana Conditions. National Weather Digest, Vol. 12 No. 1, 8pp.

No. 2. Richardson, W. S., D. J. Schwab, Y. Y. Chao, and D. M. Wright, 1986: Lake Erie Wave Height Forecasts Generated by Empirical and Dynamical Methods -- Comparison and Verification. Technical Note, 23pp.

No. 3. Auer, S. J., 1986: Determination of Errors in LFM Forecasts Surface Lows Over the Northwest Atlantic Ocean. Technical Note/NMC Office Note No. 313, 17pp.

No. 4. Rao, D. B., S. D. Steenrod, and B. V. Sanchez, 1987: A Method of Calculating the Total Flow from A Given Sea Surface Topography. NASA Technical Memorandum 87799., 19pp.

No. 5. Feit, D. M., 1986: Compendium of Marine Meteorological and Oceanographic Products of the Ocean Products Center. NOAA Technical Memorandum NWS NMC 68, 93pp.

No. 6. Auer, S. J., 1986: A Comparison of the LFM, Spectral, and ECMWF Numerical Model Forecasts of Deepening Oceanic Cyclones During One Cool Season. Technical Note/NMC Office Note No. 312, 20pp.

No. 7. Burroughs, L. D., 1987: Development of Open Fog Forecasting Regions. Technical Note/NMC Office Note. No. 323., 36pp.

No. 8. Yu, T. W., 1987: A Technique of Deducing Wind Direction from Satellite Measurements of Wind Speed. Monthly Weather Review, 115, 1929-1939.

No. 9. Auer, S. J., 1987: Five-Year Climatological Survey of the Gulf Stream System and Its Associated Rings. Journal of Geophysical Research, 92, 11,709-11,726.

No. 10. Chao, Y. Y., 1987: Forecasting Wave Conditions Affected by Currents and Bottom Topography. Technical Note, 11pp.

No. 11. Esteva, D. C., 1987: The Editing and Averaging of Altimeter Wave and Wind Data. Technical Note, 4pp.

No. 12. Feit, D. M., 1987: Forecasting Superstructure Icing for Alaskan Waters. National Weather Digest, 12, 5-10.

No. 13. Sanchez, B. V., D. B. Rao, S. D. Steenrod, 1987: Tidal Estimation in the Atlantic and Indian Oceans. Marine Geodesy, 10, 309-350.

No. 14. Gemmill, W.H., T.W. Yu, and D.M. Feit 1988: Performance of Techniques Used to Derive Ocean Surface Winds. Technical Note/NMC Office Note No. 330, 34pp.

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No. 16. Yu, T.W., 1988: A Method for Determining Equivalent Depths of the Atmospheric Boundary Layer Over the Oceans. Journal of Geophysical Research. 93, 3655-3661.

No. 17. Yu, T.W., 1987: Analysis of the Atmospheric Mixed Layer Heights Over the Oceans. Conference Preprint, Workshop Proceedings AES/CMOS 2nd Workshop of Operational Meteorology, Halifax, Nova Scotia, 2, 425-432.

No. 18. Feit, D. M., 1987: An Operational Forecast System for Superstructure Icing. Proceedings Fourth Conference Meteorology and Oceanography of the Coastal Zone. 4pp.

Contents

Abstract

In this note we review the physics of sea ice as a system forced by the atmosphere and ocean as they are relevant to the problem of making forecasts of the future state of the sea ice pack. Statistical analyses [Walsh, 1981; Chapman and Walsh, 1991; Mysak et al, 1991] and a limited theoretical analysis [Grumbine, 1993] suggest that the period of predictability may be weeks to months, making sea ice one of the most predictable components of the climate system. A feature of sea ice physics is that less is known about it than related physical systems such as the atmosphere and ocean. Consequently this review should be taken in the sense of a progress report on the state of sea ice understanding. A bibliography including sea ice papers not directly quoted is included at the end to permit the reader to identify groups actively working on aspects of the problem relevant to the reader's specific interest.

1 Introduction

The sea ice prediction problem is less well posed than those for the atmosphere and oceans. This is because less is known about the ice, and because it has been studied for a shorter period than the more familiar geophysical fluids. In this note we will review the physics relevant to the sea ice prediction problem as they are currently understood.

A feature of the sea ice prediction problem is that the items of most interest are not directly predictable. For most users, the most important fact about the sea ice is the location of the ice edge. More demanding users might also want to know the concentration of all ice (though not discriminating between ice 10 cm and 300 cm thick) behind the ice edge. Both the ice edge and ice concentration are derivable from the forecast variables.

Before becoming involved in the processes, we should understand what the variables are. In the atmosphere or ocean, the minimal variables are the three dimensional velocity field, pressure, density (mass per unit volume), temperature, and concentration of a scalar (water vapor in the atmosphere, salinity in the ocean). For sea ice, the velocity is a two dimensional field. Density is a mass per unit area. Pressure is force per unit length. And temperature, in the thermodynamic sense of a measure of the random component of velocity of the particles which make up the continuum, has not yet been defined for sea ice. So far, no scalar field has been identified as important for sea ice forecasting.

An element which appears to be important for sea ice prediction is the ice thickness distribution. The ice thickness distribution describes the fraction of the continuum area occupied by ice of a given thickness. It could be considered as being analogous to gas concentration in an atmosphere where the gasses are not well mixed, and where each gas has different physical properties (such as the thermosphere). We will discuss the thickness dependence of the different processes through this note.

The governing equations for the atmosphere and ocean are: conservation of mass, conservation of momentum, conservation of thermodynamic energy, conservation of scalars, an equation of state, and a rheology. The rheology of a continuum is part of the conservation of momentum and describes the amount of stress (units of pressure) generated by the degree (elastic) or rate (fluid) of deformation. For sea ice, the conservation of mass is similar to that for the atmosphere, given that we have multiple thicknesses of ice in our continuum element and that the thickness distribution can change by means other than advection (freezing will thicken the ice floes, for example). Conservation of momentum is also similar to the atmospheric case. Terms are added because sea ice rests on the atmosphere-ocean boundary. Conservation of thermodynamic energy is undefined for sea ice, and any scalar conservation has yet to be shown to be relevant. The equation of state for sea ice is not well understood. We will discuss it at more length in the dynamics section. The rheology of sea ice is not well known at all, with several different rheologies giving similar results for ice drift in sea ice simulations [Flato and Hibler, 1992; Ip et al., 1991]. This will also be discussed in the dynamics section.

This note will discuss the conservation of momentum, including the equation of state and the rheology, the conservation of heat energy (in exchange with the atmosphere and ocean), and the conservation of mass (the evolution of the ice thickness distribution). Given the physical basis from those discussions, we will examine the sea ice prediction problem in terms of the accuracy which might be expected given forcing from atmospheric and oceanic models or climatologies. A later note will examine the coupled problems. A feature of sea ice as a continuum to bear firmly in mind is that the continuum particles, ice floes, are macroscopic objects which have a great range in thickness (less than 10 cm to over 10 m) and in area (10 m^2 to over 100 km^2).

2 Conservation of Momentum

2.1 Free Drift Ice Dynamics

The dynamics of the ice pack are most easily understood by considering what forces can act on an individual floe. If floe-floe interactions are neglected, which we'll start with, the approximated dynamics are the free drift. First, since the floe is on the earth, we will adopt the usual geophysical practice and take an Eulerian coordinate system fixed to the earth. Symbols used in this section are defined in appendix A.

Wind passing over the floe induces a stress on the floe, typically taken to be in the form of equation 1. Also, as water moves relative to the ice, a stress is induced, taken to be in the form of equation 2, as a bulk aerodynamic skin drag. Turning angles are used when the atmospheric winds or ocean currents are given by their geostrophic values. Given that the ocean currents and ice velocities may be comparable, a more detailed computation which computes the mutual stress balance between the ice, ocean, and atmosphere is probably desirable, as in Steele et al. [1989]. The form drag (drag due to the shape of the ice floe, as opposed to the skin drag which is caused by viscous dissipation against the surface area of the floe) is an additional term, determined to be important in the marginal ice zone by Steele et al. [1989] in the form of equation 3. The sea surface may also be elevated or depressed relative to the geoid (equipotential surface). If so, the ice floes, which rest on the sea surface, then attempt to slide to the lower potential level. The acceleration due to this is given in equation 4.

$$\vec{\tau_a} = \rho_A C_{da} |U_a| \overline{\overline{R}}(\theta) \vec{U_a}^T \tag{1}$$

$$\vec{\tau_o} = \rho_O C_{do} |\vec{U_o} - \vec{U_i}| \overline{\overline{R}}(\phi) (\vec{U_o} - \vec{U_i})^T \tag{2}$$

$$\vec{\tau_f} = \frac{1}{2} \rho_O [h'/L] \Gamma \, |\vec{U_i} - < \vec{U_o} >_D| (\vec{U_i} - < \vec{U_o} >_d) \tag{3}$$

$$mg \nabla H \tag{4}$$

The resulting dynamical equation is:

$$m \frac{\partial \vec{U}}{\partial t} + m \vec{U} \cdot \nabla \vec{U} + m f \vec{k} \times \vec{U} = \vec{\tau_a} + \vec{\tau_o} + \vec{\tau_f} + mg \nabla H \tag{5}$$

where:

$$\overline{\overline{R}}(\theta) = \begin{matrix} \cos(\theta) & \sin(\theta) \\ -\sin(\theta) & \cos(\theta) \end{matrix} \tag{6}$$

$$\Gamma = [1 - (h'/L_f)^{(1/2)}]^2 \tag{7}$$

$$L_f = L\sqrt{(1/A) - 1} \tag{8}$$

and h' is the draft of the ice floes, L is their typical diameter, m is the mass of ice per unit area, A is the ice concentration, H is the dynamic topography, and θ, ϕ are constant turning angles of 23°, 25° for the atmosphere and ocean, respectively when geostrophic winds or currents are used.

If we scale the terms in equation 5 with a typical atmospheric speed of 5 $\mathrm{m\,s^{-1}}$, ocean speed of 5 $\mathrm{cm\,s^{-1}}$, ice speed of 10 $\mathrm{cm\,s^{-1}}$, floe draft of 1 m, floe diameter of 100 m, mass per unit area of 1000 $\mathrm{kg\,m^{-3}}$ (total ice cover), dynamic height of 10 cm, length scale of 100 km, coriolis parameter of 10^{-4} $\mathrm{s^{-1}}$, C_{da}, C_{do} are 1.2 10^{-3} and 5.5 10^{-3} and a time scale of a week (a typical forecast period) we find that the magnitude of the terms, respectively, is:

$$8 * 10^{-5} : 2 * 10^{-5} : 10^{-2} = 3 * 10^{-2} : 10^{-2} : 10^{-2} : 10^{-2} \tag{9}$$

Ice accelerations, linear or nonlinear, are the only negligible terms. The ice velocity is then a function of the air velocity, ocean velocity, and sea surface topography, shown in equation 10.

$$f\vec{k} \times \vec{U} = g\nabla H + \frac{\vec{\tau_a} + \vec{\tau_o} + \vec{\tau_f}}{m} \tag{10}$$

This is the free drift governing equation. An empirical simplification is that the ice drifts at some fraction of the wind speed, and at some angle to its direction. This wind-drift rule has been used for sea ice since at least the turn of the century [Nansen, 1902]. The thing which has changed in applying this rule are the constants, and the amount of observation lying behind them. Thorndike and Colony [1982] found that they could explain 70% of the variance in floe velocity by such a rule in the central Arctic basin, applying to 7937 buoy-days of observation.

2.2 Ice Rheology

Ice floes can also collide with each other. These collisions exchange momentum, which produces (in principal) an isotropic stress (pressure) and a deviatoric stress. The relation between the displacement or motion fields and the stresses is described by the rheology. As for air and water, the total stress divergence produces accelerations, as shown in equation 11. The pressure gradient, familiar from atmospheric and oceanic dynamics, is derived from the leading term in equation 11. The second term is the deviatoric stress. In air or water, the deviatoric stress is responsible for the viscosity of the fluid. It is also linearly related to the rate of strain in the fluid. For sea ice, the deviatoric stress is not apparently linear, nor is the pressure (equation of state) well-constrained. The most commonly-used relations are due to Hibler [1979, 1980], the equation of state being given in equation 12, and the rheology (deviatoric stress to rate of strain relation) is given in equations 13-17.

$$\sigma_{ij} = -P\delta_{ij}/2 + \tau_{ij} \tag{11}$$

$$P = P^* h \exp\left(-C(1-A)\right) \tag{12}$$

$$\tau_{ij} = 2\eta \dot{\epsilon_{ij}} + [\zeta - \eta]\dot{\epsilon_{kk}}\delta_{ij} \tag{13}$$

$$\zeta = P\min(\frac{1}{2\Delta}, 2.5 * 10^8 s) \tag{14}$$

$$\eta = \zeta/e^2 \tag{15}$$

$$\Delta = ((\dot{\epsilon}_{11}^2 + \dot{\epsilon}_{22}^2(1+e^{-2}) + 4e^{-2}\dot{\epsilon}_{12}^2 + 2\dot{\epsilon}_{11}\dot{\epsilon}_{22}(1-e^{-2}))^{(1/2)} \tag{16}$$

$$\dot{\epsilon_{ij}} = \frac{1}{2}(\frac{\partial U_i}{\partial x_j} + \frac{\partial U_j}{\partial x_i}) \tag{17}$$

where P is the pressure (ice strength), C is an arbitrary constant taken to be 20, τ_{ij} is the deviatoric stress tensor, η is the shear viscosity, ζ is the bulk viscosity, e is the ratio of principal axes in the assumed elliptical plastic yield curve and is taken equal to 2, and $\dot{\epsilon_{ij}}$ is the strain rate tensor. The viscosities have units of kg s^{-1}, and stresses are in N m^{-1}. Stress divergence is then in units of N m^{-2}, as required for dimensional consistence. Note that the deviatoric stress and the isotropic stress are proportional to pressure (strength). As a consequence, isotropic and deviatoric stresses are comparable, and depend quite sensitively on the description of pressure for the ice field. P* is approximately 10^4 N m^{-2}, which gives a force of about 0.1 N m^{-2} for ice about 1 m thick on length scales of about 100 km. This is significant compared to wind stresses of 0.03 N m^{-2}. In practice the pressure will only change over this range near the ice edge or near shore.

Including the internal stress field of the ice pack does not change the scaling arguments described before, but does add another two order one terms to the dynamic system – pressure gradient and deviatoric stress divergence. Since the pressure is nearly constant for a wide range of concentrations (by design, Hibler [1979]), we should reconsider the stress terms by examining the gradients (or divergence of the deviatoric stress) driving the floes. The isotropic stress gradient is proportional to $(\nabla h/C + C\nabla A)hP * \exp(-C(1-A))$. This term will be large when: A is near 1 and either or both the ice concentration or ice thickness are varying rapidly. Rapid ice concentration variations are most likely near the ice edge. Rapid thickness variations are most probable near the multi-year ice pack in spring and early winter — when the very thick multi-year pack is fringed by much thinner ice. The deviatoric stress in all the formulations is proportional to the gradient of the pressure dotted with a tensorial function of the strain, plus pressure times the divergence of the same tensor. The terms in the tensor (primarily proportional to the strain rate) are largest and most rapidly varying near coastlines, where boundary constraints are imposed physically over short distances [Hibler, 1979].

The drawback with the Hibler rheology is that it is fundamentally non-physical [Smith, 1983]. The failure rests in the fact that the relation between shear strength and dilation is specified, rather than derived by experiment from physical postulates [Smith, 1983]. The successes of Hibler's rheology are likely due to a good estimate of the physical processes and tuning.

Smith [1983] outlined physical principals which a proposed ice rheology must obey, and gave an example of a rheology which satisfied these constraints. This type of rheology was used by Häkkinen [1987], and a variant was used by Overland and Pease [1988] in a regional ice model. The notation in this section is different from that in the original papers. The fundamental postulates are [Smith, 1983]:

1) Ice has no equilibrium pressure (i.e. no tendency to expand of its own accord)

2) Ice has no ability to support tension – the ice is already cracked in several directions.

3) Resistance to deviatoric shearing is proportional to the isotropic compressive stress holding the ice floes together, and independent of the strain-rate and ice thickness – this is analogous to the Coulomb law for granular materials

4) The ice is nearly non-divergent until a critical value of isotropic stress is reached.

5) Ice is horizontally isotropic.

Smith [1983] then proposes the Reiner-Rivlin equation as the most general which satisfies constraint 5:

$$\sigma_{ij} = \gamma(S_i, \theta_1, \theta_2)\delta_{ij} + \hat{\mu}(S_i, \theta_1, \theta_2)\dot{\epsilon_{ij}} \tag{18}$$

where S_i are the state variables and

$$\theta_1 = \dot{\epsilon_{ii}} \tag{19}$$

$$\theta_2 = (\dot{\epsilon_{ij}}\dot{\epsilon_{ij}})^{1/2} \tag{20}$$

Hakkinen's rheology for $\theta_1 \leq 0$ is

$$\sigma_{ij} = -P\delta_{ij} + \phi_1\dot{\epsilon}_{ij} \tag{21}$$

where:

$$P = m\mu_0 e^{-C(1-A))} \tag{22}$$

$$\phi_1 = P(\mu_1/\mu_0)e^{-\gamma\theta_1\theta_2} \tag{23}$$

and μ_0 is 1 N m kg^{-1}, μ_1 is 10^4 m^2 s^{-1}, C is 15, γ is 3 10^8 s^{-2}. For $\theta_1 \geq 0$ the total stress is taken to be zero.

Overland and Pease [1988] modified postulates 3 and 4 in constructing their rheology, using:

3) Coulomb's law is valid for compressive stress states near the compressive strength limit.

4) Compressive strength limit is a function of ice thickness and compactness.

The motivation of the change was to obtain a more reasonable limit for the viscosity as the strain rate approached zero [Overland and Pease, 1988].

The Overland and Pease [1988] rheology is:

$$\sigma_{ij} = -P(A,h)/\sqrt{2}\delta_{ij} + \phi(\dot{\epsilon}_{ij} - \dot{\epsilon}_{ii}\delta_{ij}) \quad \theta_1 \leq 0 \tag{24}$$

$$\sigma_{ij} = 0 \quad \theta_1 \geq 0 \tag{25}$$

where

$$P = \rho_i \mu h^2 \exp(-C(1-A)) \quad (26)$$

$$\phi = \frac{DP^*}{\bar{\theta}_2 + \theta_2} \quad (27)$$

and θ_1, θ_2, and $\dot{\epsilon}$ are as already noted. P is the ice strength, ϕ controls the deviatoric stress and D is related to the Coulomb strength of the material. The tensor multiplying ϕ in equation 24 is zero when i does not equal j. The preferred constants used are C=15, D=0.6, $\bar{\theta}_2 = 5*10^{-3}\ s^{-1}$, and μ =1.6 N kg^{-1}, for a resolution of 1 km. The rheology is scale dependent, in that D, $\bar{\theta}_2$, and μ are functions of length scale [Overland and Pease, 1988]. Note that in this rheology, the ice pressure (strength) is proportional to the square of the thickness, rather than linear as for Hibler's [1979, 1980] equation of state. This was motivated by the observation that the Hibler equation of state gave excessive near shore ridging [Pease, personal communication]. The difficulty is that the equation of state (the dependence of pressure on other thermodynamic properties of the ice pack) for the ice pack has not been derivable from first principals, and apparently that differing assumptions can lead to comparably good results away from the coast.

3 Thermodynamics

Although formally it would be possible to write an ice model which included only the dynamical processes – but included all of them, including the ice interactions – this has not apparently been done outside of model tests by Hibler [1979]. An explanation is that the ice interaction processe are so complex and uncertain, adding thermodynamics as well is a realistic and not relatively expensive procedure. The thermodynamics, since they are better known, also provide a skilled element to the model, which inhibits solutions from becoming extremely unrealistic.

Ice thermodynamics are also complicated by the heterogeneous nature of the ice pack and the floes themselves. The ice pack is usually composed of floes of varying thicknesses. The heat flux through floes is a strong, nonlinear, function of thickness. The heat flux between floes and the ocean and atmosphere is also dependent on the amount of snow cover. The snow acts to insulate the ice. The snow can also cause ice formation at the surface of the floes by adding sufficient weight to depress the ice surface (only 10% of the ice thickness is above the water line in the absence of the snow) below the water line. In the melt season, particularly in regions of thick ice, ice floes can melt from the top, and collect the melt water into ponds. This again affects the floe thermodynamics, as well as the microwave signature of the pack. The latter is the reason that melt season identification of ice is difficult with passive microwave techniques [c.f. Parkinson, et al., 1987]

The principal terms in the thermal balance of an ice floe are: sensible heat flux from the ocean, sensible heat exchange with the atmosphere, latent heat exchange with the atmosphere, short wave radiation absorption from the atmosphere, long wave absorption from the atmosphere, long wave emission to the atmosphere, and thermal conduction. In the presence of a snow cover, shortwave absorption from the atmosphere, conduction, and a sensible heat flux from the ocean are retained. The snow layer keeps each of the terms mentioned for the ice-only case. The approximate magnitudes of the terms are given in table 1. Symbols used in this section are defined in appendix B. There is some overlap with the dynamics section, but it should be easy to distinguish the cosine of the solar zenith angle from viscosity (both denoted by μ in the appropriate section).

Table 1. Magnitude of thermodynamic terms over the ocean and over sea ice. In units of W m^{-2}.

Term	Size - Ice	Size - Water
SW $\downarrow$	0-350	0-350
SW $\uparrow$	200	35
LW $\uparrow\downarrow$	250	250
Internal SW	0-150	0-315
H	2-200	200
Conduction	2-300	2-30
FW $\uparrow$	2-30	2-30
LE $\uparrow$	2?	2?

The thermal energetics of ice floes (or rather, small homogeneous patches) may be divided into three classes of terms: those with no direct dependence on the ice/ocean/snow surface, those with a direct dependence, and those that are internal to the material. Variables with no direct dependence on the surface material include the downwelling longwave and heat flux from the lower ocean into the mixed layer. Internal variables are the conductive heat flux and the internal absorption of shortwave energy. Directly dependent variables are the latent and sensible heat fluxes, outgoing longwave radiation, outgoing shortwave radiation, and incoming shortwave radiation.

The downwelling shortwave energy depends on the surface when multiple reflection between surface and cloud is permitted. This term can account for 30-50% of the total shortwave flux [Shine, 1984]. The relation developed by Shine [1984] for cloudy skies is

$$F_N = \frac{(53.5 + 1274.5\mu)\mu^{0.5}(1 - 0.996\alpha)}{[1 + 0.139\tau(1 - 0.9345\alpha)]} \tag{28}$$

Where α is the albedo of the ice, τ is 3/2 LWP/r_e, with r_e being the equivalent drop size in the clouds, and μ is the cosine of the solar zenith angle. For clear sky, Shine [1984] reconsidered the relation developed by Zillman [1972] and adjusted the parameters to fit his more detailed radiative transfer model:

$$F_N = \frac{S_o\mu^2(1 - \alpha)}{1.2\mu + (1.0 + \mu)e_a 10^{-3} + 0.0455} \tag{29}$$

where S_o is the solar constant and e_a is the water vapor pressure in millibars.

The albedo of sea ice is obviously an important parameter, because of its direct role in controlling the surface energy budget for both the atmosphere and the sea ice. Unfortunately, there are limited observations of the albedo. This would not be such a concern if it weren't also for the fact that the albedo is highly variable, reaching 0.8 with a fresh snow layer, and as low as 0.5 when there are large melt ponds present [Shine and Henderson-Sellers, 1985]. The full albedo parameterization used by Shine and Henderson-Sellers [1985] is reproduced in table 2a. A different albedo scheme, which was developed by Ross and Walsh [1987] for use in comparing an ice model to observed albedoes, is given in table 2b. The albedo of bare puddled ice appears to be particularly important in controlling whether ice is seasonal or multi-year [Shine and Henderson-Sellers, 1985].

Table 2a. Albedo representation for different sea ice and snow states (Shine and Henderson-Sellers, 1985).

Albedo Class	Symbol	Value	
Dry Snow	α_d	0.80	
Thick Melting Snow	α_m	0.65	
Thin Melting Snow	α_{mb}	$\alpha_b + (h_s/0.10)(\alpha_m - \alpha_b)$	
Bare Puddled Ice	α_b	0.53	
Bare Frozen Ice	α_{bf}	0.72	
Thin Forming Ice	α_{btf}	α_{btm}	$0.0 \leq h_i \leq 1.0$
		$0.472 + 2.0(\alpha_{bf} - 0.472)(h_i - 1.0)$	$1.0 \leq h_i \leq 1.5$
Thin Melting Ice	α_{btm}	$0.472 + 2.0(\alpha_b - 0.472)(h_i - 1.0)$	$1.0 \leq h_i \leq 1.5$
		$0.2467 + 0.7049h_i - 0.8608h_i^2 + 0.3812h_i^3$	$0.05 \leq h_i \leq 1.0$
		$0.1 + 3.6h_i$	$h_i \leq 0.05$
Thin Snow On Frozen Ice	α_{df}	α_d	$h_s \geq 0.05$
		$\alpha_{btf} + (h_s/0.05)(0.8 - \alpha_{btf})$	$h_s \leq 0.05$ $h_i \leq 1.5$
		$\alpha_{bf} + (h_s/0.05)(0.08 - \alpha_{bf}$	$h_s \leq 0.05$ $h_i \geq 1.5$

Table 2b. Albedo representation from Ross and Walsh [1987].

α_{snow}	0.80	$T_s \leq$ -5 °C
	0.65 + 0.03(-T_s)	-5 $\leq T_s \leq$ 0 °C
	0.65	$T_s = 0$
α_{ice}	0.65	$T_s \leq$ 0 °C
	0.45 + 0.04 T_a	$0 \leq T_a \leq 5$
	0.45	$T_a \geq$ 5 °C

Total outgoing longwave radiation is composed of the reflection of downwelling (proportional to 1-ϵ_i, where ϵ_i is the longwave emissivity of the surface), and thermal emission. The combined effect is:

$$(1 - \epsilon_i)LW \downarrow + \sigma\epsilon_i T_s^4 \tag{30}$$

The latent and sensible heat fluxes to the atmosphere require, in principle, a coupled boundary layer analysis. The coupling is between the oceanic boundary layer, the ice boundary, and the atmospheric boundary layer. Such models are being developed [c.f. Stössel, 1991] for ice modelling use, but lie outside our present scope. A commonly used scheme is the bulk formulation:

$$H = \rho_a C_p C_H |U_{ag}|(T_a - T_s) \tag{31}$$

$$L_E = \rho_a L_v C_E |U_{ag}|(q_{10m} - q_s) \tag{32}$$

Where ρ_a is the air density, C_p is the specific heat of air at constant pressure, C_H is the sensible heat transfer coefficient, L_v is the latent heat of vaporization, C_E is the latent heat transfer coefficient, q_{10m} is the specific humidity at 10 m, q_s is the saturation specific humidity of the atmosphere at 10 m. Note that the typical sea ice modelling practice has been to use the geostrophic 10 m winds, rather than actual winds.

Downwelling longwave also depends on the nature and presence of clouds. Greater cloudiness leads to increases in the thermal blanketing effect. An approximation to this flux is [Maykut and Church, 1973]

$$FL \downarrow = (0.7855 + 0.2232C^{2.75})\sigma T_a^4 \tag{33}$$

where C is the cloud cover fraction.

The flux of heat from below the mixed layer into the mixed layer (F_W) is difficult term to quantify. Ice models have been run with a fixed flux of 2 W m^{-2} in the arctic [Hibler, 1979; Parkinson and Washington, 1979]; that figure being derived from a modelling study of Maykut and Untersteiner [1971] which gave (assuming the many other parameters and parameterizations to be correct) the best equilibrium central arctic ice pack thickness for the heat flux. Later studies with an ice model coupled to an ocean model [Hibler and Bryan, 1987, 1984; Semtner, 1987], though not including a mixed layer, did include a heat flux which depended on oceanic conditions and transports. These models found much improved agreement with observations even for the simple heat transfer included, as compared to results with a fixed heat flux. Finally, an increasing number and type of coupled ice-mixed layer models have been developed [c.f. Ikeda, 1985, Häkkinen, 1986, Mellor and Kantha, 1989, Lemke, et al., 1990, Stössel, 1991] which confirm the importance of spatially and temporally varying heat fluxes to the mixed layer for modelling sea ice.

The internal variables, conductive heat flux and shortwave transmission control the temperature profile within the ice and snow layers. These processes are complicated near the melting point by the formation of brine pockets [Maykut and Untersteiner, 1971]. In the heterogeneous process of ice floe formation, local portions of the floe can have elevated salinity, so depressed melting point. Consequently, these sections melt first. Once melted, they act as thermal reservoirs within the ice. They will also affect the radiative transmissivity of the ice. The thermal conductance and heat capacity through the ice, allowing for brine, may be modelled as [Maykut and Untersteiner, 1971]

$$(\rho c)_i = (\rho c)_{ip} + \frac{\gamma S(z)}{(T-273)^2} \tag{34}$$

$$k_i = k_{ip} + \frac{\beta S(z)}{(T-273)} \tag{35}$$

where c is the specific heat of ice, γ gives the dependence of the specific heat of ice on salinity, k is the thermal conductivity of the ice, β is the dependence of the thermal conductivity on salinity, and S is the salinity of the ice. Subscript i refers to saline ice, and subscript ip refers to pure ice. It is common to ignore the brine pocket effect, in which case the conductance and heat capacities are taken as their values for a reference salinity, usually 5 parts per thousand.

Including radiative transfer through a geophysical solid is an unusual and counterintuitive idea. It turns out to be an important process, however, particularly for relatively thin ice (less than 1 meter) [Grenfell, 1991] as is found in the Bering Sea, the Eurasian continental shelf, and most of Antarctica. The significance lies in the fact that radiation which penetrates into the snow and ice is unavailable for causing melting of the surface, thus delaying the melt process. Further, for thin ice, some of the radiation penetrates through the ice and into the ocean, warming the ocean. This warmed ocean can then sensibly contribute to basal and lateral melting of the floes. The model developed by Grenfell [1991] is a full radiative transfer computation within the snow and ice. A simpler treatment would be to use an extinction coefficient (Beer's law) derivable from work such as Grenfell's, which would provide a vertically-varying internal energy source. This has not yet been done. Maykut and Untersteiner [1971] did take the first step in this direction by considering the ice to have two layers: a thin layer where most of the radiation was absorbed, and the rest of the ice where the rest of the radiation was absorbed. The current procedure is to invoke a heat storage within the ice (without a corresponding change in temperature) up to some limit point representing the point at which is is presumed that the internally-melted water would escape the floe [Semtner, 1976].

The temperature profile through the ice is needed in order to accurately model the onset of melting and the heat exchange with the atmosphere. Maykut and Untersteiner [1971] used a highly detailed numerical scheme in their study. That scheme had the drawback of requiring a large amount of computation, and converging to its equilibrium quite slowly, prompting the development of a simpler model [Semtner, 1976]. The three layer version of Semtner's [1976] thermodynamic model is quite widely used in ice modelling due to its simplicity. The scheme is to solve the thermal evolution equations for snow (if present) and ice in a reduced number of layers.

$$(\rho c)_s \frac{\partial T}{\partial t} = k_s \frac{\partial^2 T}{\partial z^2} \tag{36}$$

$$(\rho c)_i \frac{\partial T}{\partial t} = k_i \frac{\partial^2 T}{\partial z^2} \tag{37}$$

where the subscripts i, s refer to the ice and snow properties, respectively. This is solves subject to the boundary conditions:

flux balance at the air-ice interface

$$k_i \frac{\partial T}{\partial z} = 0 \quad T(z_{atm}) < T_{melt} \tag{38}$$

$$k_i \frac{\partial T}{\partial z} = \frac{1}{\rho_i L_f} \frac{\partial H_i}{\partial t} \quad T(z_{atm}) \geq T_{melt} \tag{39}$$

flux continuity at the snow-ice interface

$$at\ z = z_{si}, \quad k_s \frac{\partial T_s}{\partial z}|_{z_{si}} = k_i \frac{\partial T_i}{\partial z}|_{z_{si}} \tag{40}$$

freezing and melting at the ice-ocean boundary

$$at\ z = z_w, \quad k_i \frac{\partial T}{\partial z} - F_W = \frac{1}{\rho_i L_f} \frac{\partial H_i}{\partial t} \tag{41}$$

where z_{atm}, z_{si}, z_w mark the boundaries between the ice floe and the atmosphere, snow and ice, and ice and water respectively. F_W is the heat flux from the ocean to the ice floe. The penetrative radiation is accumulated in a reservoir F_{io}. Whenever the upper layer ice temperature would otherwise be below freezing, heat is released from this reservoir to maintain the temperature at the freezing point. The particular savings of the method is that there is only a single point in the snow layer, and only two in the ice layer.

4 Conservation of Mass

We now step back from considering individual floes to examine properties of the ice pack. In considering an individual floe, we saw that the thickness was quite important both in heat fluxes and dynamic processes. When we broaden our focus to include many floes, we cannot speak of the ice thickness, but of a thickness distribution instead. Another feature we notice from a broader vantage is that the sea ice cover is not always continuous; the ice is only a fractional (though often a large fraction) cover on the sea. Symbols from this section are defined in appendix C. Note that contrary to common mathematical usage, the delta function carries units here, m^{-1}. Our presentation follows Thorndike et al. [1975].

Both the ice thickness distribution and areal coverage may be represented by g(h;x,y,t), where g is the fraction of an area centered at x,y at time t, which is covered by ice between thickness h and h+dh [Thorndike, et al., 1975]. The evolution of g is governed by:

$$\frac{\partial g}{\partial t} + \nabla \cdot (\vec{U} g) + \frac{\partial f g}{\partial h} = \psi + F_L \tag{42}$$

where U is the velocity field, f is the growth rate of ice (f=f(h)), ψ is a function which describes the mechanical redistribution of ice from one thickness class into another class, and F_L is the lateral growth of ice of thickness h (added by Hibler [1980]). U is determined by ice dynamics while f and F_L are controlled by ice floe thermodynamics.

The redistribution function, ψ, is determined by the mechanics of floe interaction, subject to global constraints. The first constraint is the floe interaction must not change the total volume of ice per unit area (i.e. the mean thickness is conserved):

$$\int_0^\infty \psi h dh = 0 \tag{43}$$

ψ must also compensate for ice convergence/divergence by creation or destruction of leads, and possibly ridging thinner ice into thicker:

$$\int_0^\infty \psi dh = \nabla \cdot \vec{U} \tag{44}$$

A conservation of energy can also be stated [Hibler, 1980]

$$C \int_0^\infty h^2 \psi dh = \sigma_{ij} \dot{\epsilon}_{ij} \tag{45}$$

where C relates the potential energy change (the integral) to the amount of work done (the right hand side, with terms as defined in the dynamics discussion).
For pure divergence, lead formation:

$$\psi = \delta(h) \nabla \cdot \vec{U} \tag{46}$$

Under convergence, ice ridging:

$$\psi = W_r(h,g)\nabla \cdot \vec{U} \tag{47}$$

W_r is constrained to conserve volume per unit area and to match the divergence.

$$\int_0^\infty W_r(h,g)dh = -1 \tag{48}$$

$$\int_0^\infty W_r(h,g)hdh = 0 \tag{49}$$

The ridging kernel, W_r, may be specified in several ways. Thorndike et al. [1975] used the mathematically simple, but physically questionable [Hibler, 1980] assumption that ice ridged into stacks k times thicker than the initial thickness, with k selected as 5. Hibler [1980] outlined a more general ridging kernel:

$$W_r(h,g) = \frac{-P(h)g(h) + \int_0^\infty \gamma(h,h')P(h')g(h')dh'}{\int_0^\infty (P(h)g(h) - (\int_0^\infty \gamma(h,h')P(h')g(h')dh')dh)} \tag{50}$$

Where P is the probability that ice of thickness h ridges [Thorndike et al., 1975].

$$P(h) = \max(1 - \int_0^h g(h)\frac{dh}{c_1}, 0) \tag{51}$$

where c_1 is taken to be 0.15

$$\gamma(h_1,h_2) = \delta(h_2 - kh_1)/k \tag{52}$$

and k=5 for Thorndike et al., [1975]

$$\gamma(h_1,h_2) = \begin{matrix} \frac{0.5}{H^* - h_1} & 2h_1 < h_2 < 2\sqrt{H^* h_1} \\ 0 & otherwise \end{matrix} \tag{53}$$

and H^*=100 m for Hibler [1980].

The ice thickness distribution will change most rapidly when: the flow is strongly divergent or convergent (as under a strong storm system), the thickness distribution is rapidly varying in space (as near the marginal ice zone or in spring and fall near the multiyear pack) or when the growth rate (f) is strongly thickness dependent (winter). As for the times and places of most difficult dynamic forecast, the times of most difficult thickness distribution prediction are those where the interest is greatest — in the fall and spring, and always in the marginal ice zone.

The partition of thermal energy between the ice and ocean is another feature of the ice pack at larger scales. Determining the fraction of energy received by the ocean which is used for warming the ocean, causing sidewall ice melting, or causing basal ice melting is the difficulty [Maykut and Perovich, 1987]. Ocean temperatures as high as 10 °C have been observed in ice-surrounded water regions (polynyas) [Maykut and Perovich, 1987].

In the short term there is no difference in melt rate between sidewall and basal melting. The longer term effects can be quite different because sidewall melting increases the fraction of the surface which is covered by low albedo water. Modelling the difference between sidewall and basal melting requires a coupled ocean-ice model. The techniques used in an uncoupled mode completely ignore this element. The difference between sidewall and basal melting shows up geophysically as an enhanced or decreased, respectively, sensitivity to the ice albedo feedback [Maykut and Perovich, 1987]. An important parameter identified by [Maykut and Perovich, 1987] is the floe size distribution; small floes expose relatively more sidewall to the ocean, so should be more prone to lateral melting than large floes.

The momentum partitioning between floes is even less well understood. It is known [Rothrock, 1975] that at length scales much below 100 km, the ice pack ceases to behave as a continuum. One element of this failure is that local averaged floe velocities differ significantly from areal average velocities. This results in sub-regions of the ice pack colliding with each other (and then ridging or fragmenting floes) or separating (producing leads and polynyas) at rates not directly derivable from the large-scale ice flow field. It appears that the ice pack, even when examined at scales small compared to the continuum scale, often flows as a solid unit [Lepparanta and Hibler, 1985]. This has been the justification for applying large-scale ice rheologies to problems with much smaller length scales [Lepparanta and Hibler, 1985]. It is also the probable reason that such uses have had some success [c.f. Hibler, 1979, 1980; Walsh et al., 1985; Lepparanta and Hibler, 1984; Hibler and Ackley, 1983; Walsh and Zwally, 1987; Semtner, 1987]. At other times and regions, the differential motions may be quite important [Preller et al., 1989]. The current operational practice is to ignore the non-continuum effects.

5 Desirable Accuracies for Prediction

5.1 Dynamics

The dynamic features of greatest operational interest are the ice edge location and ice motion field. Ice concentration and thickness are also desirable. The ice edge is forecast weekly by the Navy/NOAA Joint Ice Center for 7 days ahead for the Arctic and Antarctic [Feit, 1989]. The ice motion field is currently not forecast, but is of interest to users such as offshore drilling companies. The desirable precision for the ice edge forecast is the resolution limit (or better of course) of the analyses.

In cloudy areas, the analysis accuracy is approximately 25 km, improving to about 1 km in cloud-free areas [Feit, 1989]. Over one week, these precisions correspond to speeds of 4.1 and 0.17 $\mathrm{cm\,s^{-1}}$ respectively. The cloudy region precision can be reached by a mesoscale ice model on an Eulerian grid [c.f. Preller et al., 1989]. But the cloud-free analysis precision of 1 km would require approximately 10^4 times the present computing load, suggesting that other schemes will merit consideration as the models become more skilled.

The precision required in the speeds imposes some constraints on the precision of the forcing. The ice velocity from the Thorndike and Colony [1982] drift rule is approximately 0.008 times the wind speed. So a 0.17 $\mathrm{cm\,s^{-1}}$ ice velocity precision requirement corresponds to approximately 0.2 $\mathrm{m\,s^{-1}}$ in the wind speed. The corresponding limit for a 25 km ice edge location precision is 5 m/s.

Sea surface topography (balanced by the Coriolis force on ice) at 100 km oceanic resolution is needed to 0.22 cm for the 1 km ice edge precision, or 5.3 cm for 25 km precision. Variations in the Coriolis parameter, the beta effect, can probably be neglected for 25 km precision, but will need to be retained for 1 km precision. The constraint is imposed by requiring the error in f(y)*u to be less than the allowable error in u, given a reference ice velocity of 10 cm/s. The 25 km precision corresponds to a 40% error, while 1 km precision requires f accurate to about 2%. The constraints noted here are listed in table 3.

Table 3. Precision required in forcing terms for desired accuracy in forecast ice motion.

Precision (1 wk)	H_{topo} (cm)	f	U_A (5 m/s)	U_O (10 cm/s)
1 km	0.22	2%	±0.02	± 0.08
25 km	5.3	40%	±0.40	± 2.3

5.2 Ice Floe Thermodynamics

The fields computed by considering ice floe thermodynamics are the ice and snow thickness and the vertical temperature profile within the ice. The ice temperature is not operationally useful, but is required in predicting thickness. Ice thickness is an operational interest, since ships which may pass safely through thinner (10 cm) ice cannot attempt the passage if the ice is thick. Ice thickness is not currently analyzed or forecast as such. Instead, ice type (young, thin, first year, multiyear) is used as a proxy.

The precision requirement for ice thickness prediction is set by the thickness at which ice first becomes reliably detectable by satellite passive microwave observing systems, and the thickness which may hamper ship operations. For both cases, the thinnest ice is 10 cm [Zwally et al., 1983, Callahan, 1991, respectively]. We will consider time scales of one week, one month, and a year. Forecasts are issued for a week and a month, while the year time scale corresponds to climatic simulations. The required precision in W m^{-2} for predicting the growth or melt of 10 cm and 1.0 m ice at each of these time scales is given in table 4. The table also casts this precision in terms of the relative precision needed in each of the thermal forcing terms. The table includes the present observation and modelling precisions.

Table 4. Required accuracy in thermodynamic fluxes to predict the growth of thin (10 cm) and thick ice (1 m) in W m^{-2} and the relative magnitude of this flux compared with the size of individual elements in the total flux. α is the albedo, LW is the longwave flux, FW is the ocean-supplied heat flux, K_{thin} is the thermal conduction through thin ice, K_{thick} is the thermal conduction through thick ice, and $S_{thin,thick}$ is the thermal conduction through thin or thick ice with a 10 cm layer of snow.

Time	Thin	Thick	α_{thin}	LW_{Thin}	FW_{Thin}	K_{Thin}	K_{Thick}	S_{Thin}	S_{Thick}
Week	50	500	0.14	0.25	—	0.125	—	1.0	—
Month	11	110	0.032	0.05	1.0	0.025	—	0.2	—
Year	1	10	0.003	0.005	0.1	0.0025	0.25	0.02	0.4

The precision required to predict the growth of 10 cm of ice in a week, about 50 W m^{-2}, is quite modest relative to the magnitude of the forcing terms, about 10% of the largest.

For a one month forecast, the relative precision is still only about 3%. Integration over an annual cycle to predict ice to an accuracy of 10 cm thickness requires 1 W m^{-2} accuracy, or about 0.3% relative precision. Note, though, that the requirement for predicting 1 meter thick ice for a climatic (year) simulation is only several percent. This is likely the reason that models have been more successful at predicting mean thicknesses than ice edges. Note too that the thermal conduction and snow blanketing are thickness dependent. The thicker snow blanket reduces the required thermodynamic precision in the ice forecast.

The ability of the thermodynamic model to predict ice thickness is more important than may seem. There are no data available on a regular basis for large areas on ice thickness. Ice type is available, but only at fairly low accuracy [Cavalieri et al., 1984] for the present, and is not the same feature as thickness. There is hope that synthetic aperture radar will improve the spatial resolution of the ice type analysis. The operational FNOC model consequently uses an ad hoc means of initializing the ice model thicknesses when the areal concentration of ice is different from the predicted. If the analysis shows no ice where the model had ice, then the ice thickness is set to zero [Preller and Posey, 1989]. If the analysis has ice where the model does not, for ice concentrations of 0.15 to 0.5, ice thickness is set to 0.5 m, and for ice concentrations greater than 0.5, the ice thickness is set to 1.0 m [Preller and Posey, 1989]. It is unclear what is done when the concentrations differ, but are not zero in the model or the observations. If the model is able to simulate the ice growth and decay well, the effect of the ad-hoc adjustments for initial conditions will be relatively minor.

6 Recommendations

We have discussed the physics of ice and its modelling in isolation from the ocean and atmosphere to the greatest extent possible. This has permitted us to examine the behaviour of the ice in relatively simple context. This simplification nonetheless retains many of the difficulties which remain (or are aggravated) on coupling into the fuller climate system. Consequently, there are certain directions of research or operational implementation which are evident, and which remain important in the fuller system.

Dynamics is paradoxically the easiest and hardest element of the sea ice forecast problem. It is the easiest in that quite simple models – including a drift rule – can account for much of the variance that quite complex physical models can explain. It is the most difficult in that the proper rheology, equation of state, and conservation of random motion (continuum energy) have yet to be derived rigorously. Consequently, any model of these elements should be viewed as an approximation to some unknown rheology.

That various models have similar skill in predicting ice velocity suggests that ice velocity is not a good measure of skill. A feature of the ice velocity field which does discriminate more between models is the divergence [Hibler, 1990]. Convergence also induces ridging and creates the thickest ice. Thick and ridged ice are particular hazards, so are important quantities to forecast well. So, divergence, rather than velocity should be used whenever possible as the parameter for verifying dynamic models.

Since there is no rigorous derivation of the proper rheology or equation of state for sea ice, we should prefer one which is most easily tuned, and which never makes physically unrealistic forecasts. The Hakkinen [1987] and Overland and Pease [1988] rheology is more apparently tuneable than the Hibler [1979]. This derives from the fact that the Hibler rheology includes branch points (maximum and minimum pressures and viscosities), while the other rheologies are are more nearly continuous functions. The Hibler rheology is also at root nonphysical [Smith, 1983], while the other rheologies represents an exact physical rheology, though one not yet rigorously proven for sea ice. Finally, the Hibler equation of state permits unrealistic ridging near coasts [Pease, personal communication]. In spite of these differences between rheologies, it is not clear that the other rheologies actually leads to better hemispheric sea ice forecasts over week to month time scales. Our preference is based on the belief that that rheology can be improved more readily than Hibler's. Between the two alternate rheologies to Hibler's, Hakkinen's [1987] appears the most readily modified.

It is clear from coupled ice-ocean studies that the ocean-ice stress needs to be modelled in the framework of a coupled system, rather than as ice simply being advected by the ocean currents. The ice-ocean stresses represent a nontrivial momentum source/sink for the ocean and the surface mixed layer. Consequently, we should migrate away from using simply the geostrophic winds and currents with turning angles as is currently common.

For ice thermodynamics it is even more obvious that uncoupled models will not produce satisfactory results. It is also clear from experiments [c.f. Hibler 1980] that different thermodynamic representations can lead to substantially different results, even under the same apparent forcing. Three boundary layers, air-ice, ice-ocean, and radiative, occur with respect to the sea ice, all of which need some degree of coupling eventually. The ocean mixed layer has received the most work. The high spatial and temporal variability of heat flux from the mixed layer to the ice in the marginal ice zone has been shown to be important to accurate prediction of the ice pack edge.

The atmospheric thermal boundary layer over the ice can also exert strong influence over the ice edge location. So far, it has appeared sufficient to force the ice with an atmosphere which is aware of ice parameters (roughness, thickness, fractional cover) rather than fully coupling the models.

The existance of the third boundary layer, a radiative boundary layer, is now being recognized as important. Again, in this layer it appears most important to ensure that the atmosphere is aware of ice parameters (albedo, fractional coverage) rather than to make a fully coupled model.

In addition to the atmospheric and oceanic thermodynamic effects, the thermodynamics of the ice itself can be better represented. The classic work on this subject was by Maykut and Untersteiner [1971], and included non-equilibrium temperature profiles, ice salinity, and penetrative radiation. Most of the extant sea ice models use Semtner [1976] thermodynamics instead. This scheme was developed by Semtner for use in general circulation models, and was optimized to be the simplest scheme which preserved the sense of Maykut and Untersteiner's [1971] results. Present computational power makes this simplification unnecessary, and the improvement in thermodynamic properties possible by returning to the original scheme appears to be significant for ice models.

The conservation of mass (ice thickness distribution) includes a problematic term, ridging. The other terms are reasonably well understood. Two significantly different ridging models have been proposed, Thorndike et al. [1975] and Hibler [1980]. Hibler's is more physically based, at the cost of greater complexity. Consequently, we shall use the Hibler [1980] ridging. An area to research is an alternative formulation which would be simpler but still physically based.

The heart of the NMC interest in sea ice is to make predictions of the ice cover. Key variables are the ice edge, concentration, and thickness. Unfortunately, only two of these, ice edge and concentration, are observable directly on global scales. Consequently, a data assimilation scheme which uses observations of ice concentration to infer the ice thicknesses should be developed. This will also permit better tests of the ice forecasts, as ice concentration is forgiving of errors in model formulation [Hibler, 1990].

Tools for forecast verification need to be developed. It is common currently to verify only large scale, long term variables, such as mean annual ice thickness or total ice extent. The only smaller scale verification variable is the local ice velocity where buoys are available. As already mentioned, ice velocity is not very discriminatory between forecast schemes. What is needed are forecast variables which differ significantly between models, and which are local rather than global.

7 Appendix A

Symbols - Dynamics

Symbol	Value	Parameter
A		Ice cover fraction
C	20	Parameter in Hibler [1979] equation of state
	15	Parameter in Overland and Pease [1988] equation of state
C_{da}	$1.2\ 10^{-3}$	Air-ice drag coefficient
C_{do}	$5.5\ 10^{-3}$	Ocean-ice bulk drag coefficient
D	0.6	Coulomb strength in Overland and Pease [1988] rheology
e	2	Ratio of principal axes in stress rule for Hibler [1979] rheology
f	2Ω sin(latitude)	Coriolis parameter
g	$9.81\ m\ s^{-2}$	Acceleration due to gravity
h	m	Mean ice thickness
h'	m	Ice draft
H	m	Ocean dynamic topography
L	m	Mean radius of ice floes in region
m	$kg\ m^{-2}$	Ice surface density
	2700	Arctic typical value
	600	Antarctic typical value
P	$N\ m^{-1}$	Ice pressure
P*	$10^4\ N\ m^{-2}$	Failure strength of ice
R		Rotation matrix
$\vec{U}_a$	$m\ s^{-1}$	Wind velocity
$\vec{U}_i$	$m\ s^{-1}$	Ice velocity
$\vec{U}_o$	$m\ s^{-1}$	Ocean velocity
α		Angle of repose for Coloumb material limit in Overland and Pease [1988] rheology
δ_{ij}		Kronecker delta tensor
η	$1.7\ 10^{12}\ kg\ s^{-1}$	Maximum ice shear viscosity in Hibler [1979] rheology
$\dot{\epsilon}_{ij}$	s^{-1}	Rate of strain tensor
γ	$3\ 10^8\ s^{-2}$	In Hakkinen [1987] rheology
μ	$1.6\ N\ kg^{-1}$	Related to strength in Overland and Pease [1988] equation of state
μ_0	$1.0\ N\ m\ kg^{-1}$	Related to strength in Hakkinen [1987] equation of state
μ_1	$10^4\ m^2\ s^{-1}$	Hakkinen [1987] viscosity

Symbol	Value	Parameter
ϕ		In Overland and Pease [1988] rheology
Ω	$7.292\ 10^{-5}\ s^{-1}$	Rotation rate of earth
ϕ	$25°$	Ocean current turning angle
ρ_a	$1.29\ kg\ m^{-3}$	Density of air
ρ_o	$1027.8\ kg\ m^{-3}$	Density of ocean
σ_{ij}	$N\ m^{-1}$	Total ice stress tensor
τ_a	$N\ m^{-2}$	Air-ice stress
τ_o	$N\ m^{-2}$	Ocean-ice surface stress
τ_f	$N\ m^{-2}$	Ocean-ice form drag
τ_{ij}	$N\ m^{-1}$	Deviatoric ice stress tensor
θ	$23°$	Atmospheric winds turning angle
θ_1	$\dot{\epsilon}_{kk}$	First stress invariant
θ_2	$\dot{\epsilon}_{ij}\dot{\epsilon}_{ij}^{0.5}$	Second stress invariant
$\overline{\theta_2}$	$5\ 10^{-3}\ s^{-1}$	Reference strain rate in Overland and Pease [1988] rheology
ζ	$6.9\ 10^{12}\ kg\ s^{-1}$	Ice bulk viscosity in Hibler [1979] rheology

8 Appendix B

Symbols - Thermodynamics

Symbol	Value	Parameter
C		Cloud cover fraction
c_i		Specific heat of ice
c_{ip}	1880 J kg^{-1} $^{\circ}K^{-1}$	Specific heat of pure ice
c_s	690 J kg^{-1} $^{\circ}K^{-1}$	Specific heat of snow
C_E	1.75 10^{-3}	Bulk transfer coefficient for latent heat
C_H	1.75 10^{-3}	Bulk transfer coefficient for sensible heat
C_p	1004 J kg^{-1} K^{-1}	Specific heat of air at constant pressure
e_a	hPa	Vapor pressure of water vapor at 10 m
FL↓	W m^2	Downwelling long wave radiation
F_N	W m^{-2}	Downwelling short wave radiation
H	W m^{-2}	Sensible heat transfer between ice and atmosphere
k_i	2.2 W m^{-1} K^{-1}	Thermal conductivity of ice
k_{ip}	W m^{-1} K^{-1}	Thermal conductivity of pure ice
k_s	0.31 W m^{-1} K^{-1}	Thermal conductivity of snow
L_f	3.34 10^5 J kg^{-1}	Latent heat of fusion
L_e	2.5 10^6 J kg^{-1}	Latent heat of evaporation
L_v	2.834 10^6 J kg^{-1}	Latent heat of vaporization
LW↓	W m^{-2}	Downwelling long wave radiation
q_{10m}	g kg^{-1}	Water vapor mixing ratio at 10 m
q_s	g kg^{-1}	Water vapor mixing ratio at T_a
r_e	m	Equivalent drop radius
S(z)	g kg^{-1}	Salinity of the ice
S_0	1367 W m^{-2}	Solar constant
T	K	Ice/snow temperature versus depth
T_a	K	Air temperature at 10 m
T_s	K	Ice/snow surface temperature
U_{ag}	m s^{-1}	Geostrophic air velocity at 10 m
z_{atm}	m	Air - floe surface boundary
z_{si}	m	Snow - ice boundary
z_w	m	Ice - ocean boundary
α		Albedo
β		Dependence of thermal conductivity on salinity
ϵ_i	0.97	Long wave emissivity of ice
γ		Dependence of specific heat times density on salinity
μ		Cosine of the solar zenith angle
ρ_a	1.29 kg m^{-3}	Density of air
σ	5.67 10^{-8} W m^{-2} K^{-4}	Stefan-Boltzman constant
τ		Optical depth

9 Appendix C

Symbols - Mass Conservation

Symbol	Value	Parameter
c_1	0.15	
f	m s^{-1}	Freezing rate
F_L	m^{-1} s^{-1}	Increase in ice cover due to lateral freezing
g	m^{-1}	Ice concentration per unit thickness interval
h	m	Thickness
h_1	m	Thickness of thinner ice in ridging
h_2	m	Thickness of ridged ice
H*	100 m	Limiting thickness of ice in Hibler [1980] ridging
k	5	Thorndike et al [1975] ridging parameter ice ridges to form new ice k times thicker than original
P		Probability that ice of thickness h ridges
t	s	Time
U	m s^{-1}	Ice velocity
W_r	m^{-1}	Redistribution kernal under convergence
δ	m^{-1}	Delta function (units from Thorndike et al. [1975])
$\gamma(h_1, h_2)$		Probability that ice with thickness h_1 that ridges to form ice with thickness h_2
ψ	m^{-1} s^{-1}	Ridging function

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