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The Role of Sea Ice Dynamics in Global Climate Change

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Introduction

The response of the polar regions to climatic change is significantly affected by the presence of sea ice. This sea ice cover is very dynamic and hence contains a variety of thicknesses, ranging from open water to pressure ridges tens of meters thick. In addition to undergoing deformation, the ice pack is typically transported from one region to another, with melting and freezing occurring at different locations. This transport tends to create net imbalances in salt fluxes into the ocean. On the large scale, an important factor here is the amount of ice drifting out of the Arctic Basin through the Fram Strait. Information on the variability of this export can be estimated from satellite data on ice drift. However, estimates of the thickness distribution of the ice are difficult to obtain. Moreover the spatial and temporal variability of the transport is substantial. There is also the issue that the mechanical characteristics of the ice cover, which dictate the amount of ice flowing out of the Arctic Basin, may vary in response to climate change, and hence may cause feedback effects that could affect the response of the system.

As a consequence it seems clear that while statistical models and observations of present-day ice extent are important for validating physically based models of sea ice growth, drift, and decay, relying solely on such observations to ascertain the sensitivity of the high latitudes to climatic variations may result in leaving out important feedbacks that could affect the response. Instead, it appears important to develop physically based models to successfully explain observed features of sea ice growth drift and decay and then to

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make use of some version of these models in numerically based climate studies.

In this regard it is important to develop models that include ice drift and dynamics in climate studies and begin to examine the response of such models to simulated rather than observed atmospheric forcing. These comments need to be viewed in light of the fact that present climatic studies usually only include thermodynamic sea ice models which do not even come close to including the major sea ice processes relevant to climatic change. As a consequence, inclusion of any level of data-verified ice dynamics would appear to be an improvement provided the forcing wind fields of the atmospheric circulation models have an acceptable level of correctness. Some simple, robust sea ice dynamic models developed by the author and co-workers are discussed below.

Overall, it appears that there are three broad areas where determining the physical mechanisms is important for developing a physically based understanding of the response of the high latitudes to climate change. They are, broadly, sea ice dynamics and thermodynamics, the thickness distribution of sea ice and its evolution, and the coupling of sea ice with the ocean. Aspects of these features relevant to climatic change are discussed below. A more detailed pedagogical discussion of this material is given in Hibler and Flato (in press).

Sea Ice Dynamics

General Characteristics of Sea Ice Drift

The overall characteristic of ice drift is that on short time scales, it tends to follow the wind, with the drift approximately following the geostrophic wind with about one-fiftieth of the magnitude. This general feature of ice drift has been known for many years, beginning with Fridtjof Nansen's expeditions to the Arctic at the end of the 19th century. Since wind variations are larger than those of currents on a short time scale, this also means that, except for shallow regions, fluctuations in ice drift will be dominated more by wind than by currents. However, on a long time scale, the more steady currents can play a significant role.

These characteristics are illustrated in Figure 1, which shows short-term and long-term drift rates using a linear ice drift model. As can be seen, including current effects has only a minor effect on short-term variations. In examining cumulative drift over several years, however, significant differences occur if currents and ocean tilt are neglected. Basically, while smaller, current effects are steady. On the other hand, wind effects, while large, tend to fluctuate out over a long time period, leaving a smaller constant value.

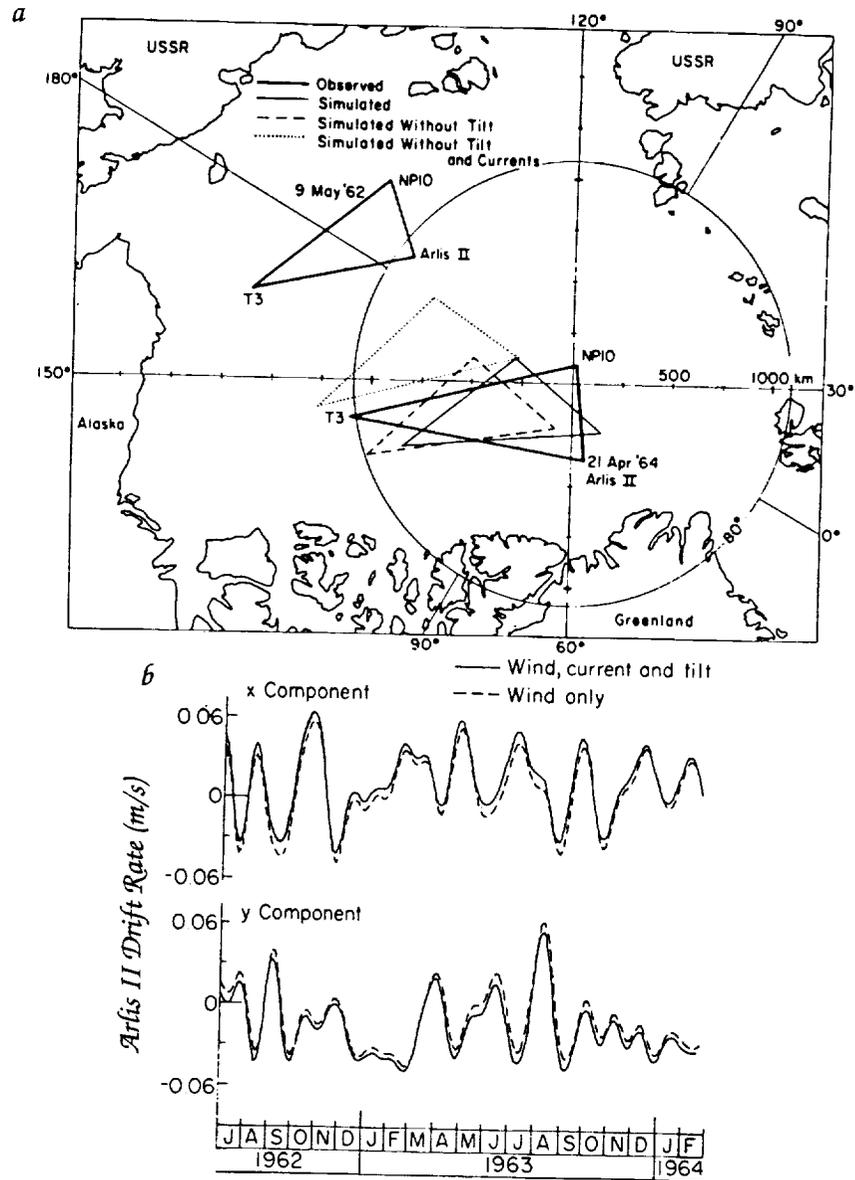


Figure 1. (a) Predicted and observed net drift of three drift stations, with and without current and ocean tilt effects; (b) effect of currents and ocean tilt on the 25-day smoothed predicted drift rate of the Arlis II ice island (from Hibler and Tucker, 1979; reproduced by courtesy of the International Glaciological Society).

Sea Ice Rheology

Sea ice motion results from a balance of forces that include wind and water drag, sea surface tilt, and ice interaction. The latter is a consequence of the large-scale mechanical properties of sea ice which include shear and compressive strengths but little tensile resistance.

In many cases the net effect of the ice interaction is to cause a force opposing the wind stress roughly in the manner shown in Figure 2. As a consequence, to achieve the same ice velocity under these circumstances, a larger wind stress more nearly parallel to the ice velocity is required. This feature is consistent with observations by Thorndike and Colony (1982), which show ice motion to be very highly correlated with geostrophic winds, with the ice drift rate decreasing somewhat in winter for the same geostrophic wind. However, examination of drift statistics shows higher winds producing an almost discontinuous shear near the coast. Moreover, observations show that while ice drift against the shore increases the ice thickness, the buildup is not unlimited.

As noted by Hibler (1979), both of these characteristics of discontinuous ice drift and a limit on near-shore ice buildup may be explained by nonlinear plastic ice rheologies. These rheologies have yield stresses that are relatively independent of strain rate. Hence far from shore, even though the ice is interacting strongly, there may be very low stress gradients since the stresses are relatively

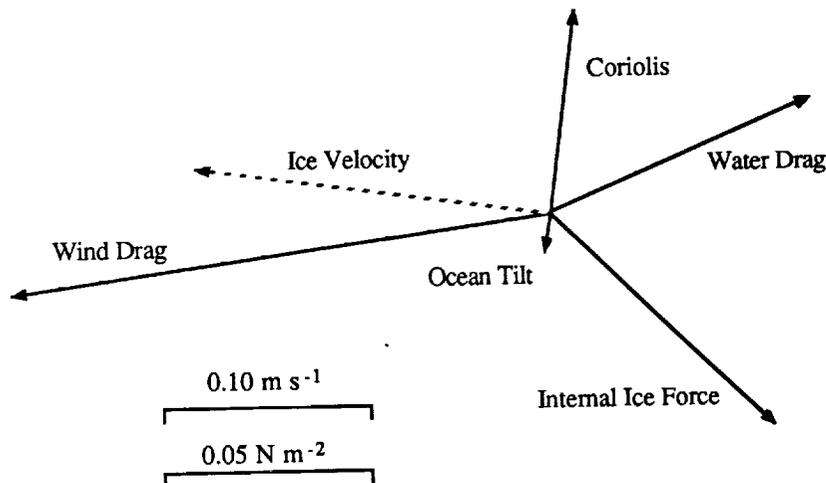


Figure 2. An estimate of the force balance on sea ice for winter conditions based on wind and water stress measurements (from Hunken, 1975). In this balance the force due to internal ice stress is determined as a residual and the dashed line shows the ice velocity.

constant. Also this fixed yield stress will cause a discontinuous slippage at coastal points and prevent the ice from building up without bound. Without such a nonlinear rheology it is very difficult to obtain these features.

For climatic purposes probably the most important feature of the ice interaction is some type of resistance to ice buildup that does not drastically affect the ice drift. While this can be accomplished by a nonlinear plastic rheology including a shear strength, reasonable results may also be obtained by considering the ice to have only resistance to compression and no resistance to dilation or shearing. This "cavitating fluid" model has been compared by Flato and Hibler (1990) to a more complete plastic rheology with good success. The ice interaction term for the cavitating fluid model is characterized by the following stress tensor:

$$\sigma_{ij} = -P \delta_{ij} \quad (1)$$

where σ_{ij} represents the i and j components of the two-dimensional Cartesian stress tensor, P is ice pressure, and δ_{ij} is the Kronecker delta which is equal to 1 if $i = j$ and 0 otherwise.

This rheology can be simulated using the viscous-plastic scheme of Hibler (1979) or an iterative method, in which the free drift velocity field is calculated (i.e., by a simple analytic solution of the momentum equation neglecting the ice interaction term) and then this estimated field is corrected in an iterative manner to account for the compressive strength of ice. This correction can be accomplished by adding a small outward velocity component to each computational grid cell such that after correction all convergence is removed. After a number of interactions all convergence will be removed from the velocity field, simulating an incompressible ice cover. This correction procedure can also be formulated in terms of an internal ice pressure field, allowing specification of a failure strength beyond which plastic flow ensues. A particularly simple approximate solution to the cavitating fluid applicable to climate modeling was presented earlier by Flato and Hibler (1990).

The appropriate yield stress depends on how one is characterizing the ice thickness distribution as discussed below. However, if one is using a simple two-level sea ice model where H is the mean thickness and A is the compactness (i.e., fraction of area covered by ice), a reasonable yield stress P is given by

$$P = P^* H e^{-K(1-A)} \quad (2)$$

where K is 20, P^* is a fixed strength constant (say 2.75×10^4 N/m²), and e is the exponential function. This strength formulation was originally proposed by Hibler (1979) and has been utilized by Flato

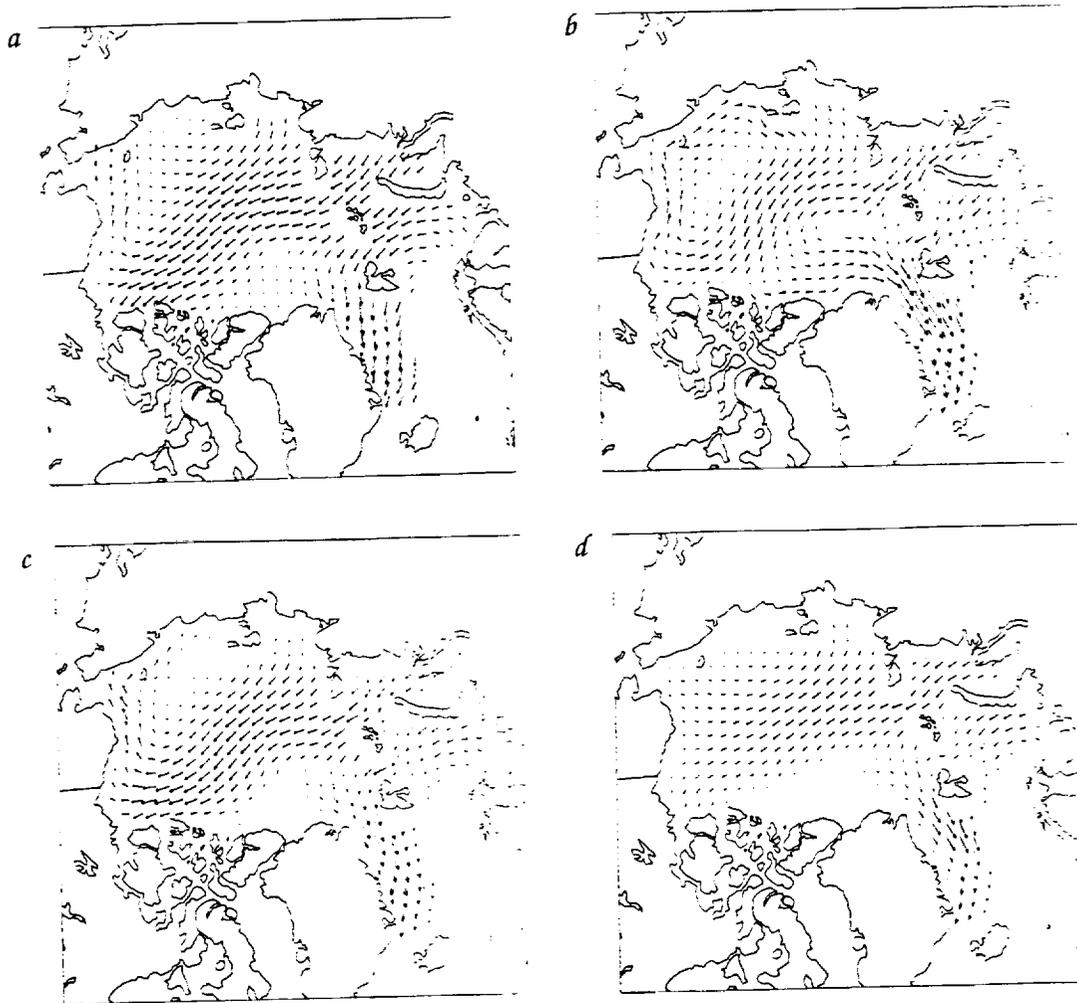


Figure 3. Average ice velocity fields calculated using forcing from March 1983: (a) free drift, (b) incompressible cavitating fluid, (c) cavitating fluid model with realistic compressive strength, (d) viscous-plastic, elliptical yield curve model with shear and compressive strength.

and Hibler (1990, in press). This procedure is easily extended to spherical coordinates as discussed by Flato and Hibler (in press).

An example of the use of the cavitating fluid model is shown in Figure 3, where we have applied the cavitating fluid correction to the free drift velocity field shown in Figure 3a. Two types of corrections are applied: In Figure 3b a totally incompressible sea ice drift field is shown, while in 3c we have assumed a constant 3-m-thick ice everywhere with a 2% fraction of open water. This, together with a yield strength constant P^* of $2.75 \times 10^4 \text{ N/m}^2$, gives a maximum allowable two-dimensional pressure of $5.53 \times 10^4 \text{ N/m}$. As can be seen,

including a maximum yield pressure modifies the velocity field somewhat inasmuch as some convergence is allowed. The main characteristic of both fields, however, is that the cavitating fluid approximation does not damp out the ice velocity field but rather simply modifies it to prevent convergence.

A more complete sea ice rheology, the viscous-plastic elliptical yield curve rheology, was presented by Hibler (1979). In this scheme plastic failure and rate-independent flow are assumed when the stresses reach the yield curve values represented by an ellipse in principal stress space. Here the compressive strength is given by the length of the ellipse while the shear strength is given by its width. Stress states inside the yield curve correspond to slow viscous creep deformation. This model has been widely used and produces realistic thickness and velocity patterns; however, its relative complexity and the dramatic slowdown in ice drift it produces when the wind fields are temporally smoothed (Flato and Hibler, 1990) make it somewhat less desirable for long-term climate studies. As a comparison, the velocity field calculated using this model with a constant ice strength of 5.53×10^4 N/m is shown in Figure 3d. Readily apparent here is the less robust velocity field that results from the increased resistance afforded by the shear strength. For a more complete comparison of a variety of plastic rheologies for the Arctic Basin, the interested reader is referred to Ipp et al. (in press), where a variety of nonlinear rheologies are compared, including a more exact solution to the cavitating fluid than presented above. It should be noted that the approximate solution discussed here compares very closely with exact, but less computationally efficient, solution methods (see, e.g., Flato and Hibler, in press).

It should be emphasized that the main reason for using some model such as the cavitating fluid in ice drift is to ensure that realistic ice transport occurs. When ice forms, heat is transferred to the atmosphere (i.e., by the latent heat of fusion). When the ice melts later at a different location, it absorbs this latent heat. Consequently, in some sense, the net effect of ice transport is to transfer heat from one location in the atmosphere to a different location. Ice transport effects are even more pronounced for the oceanic circulation in that where the ice freezes, most of the salt is expelled into the ocean. The ice is then transported to a different location, where it melts and produces a surface fresh water flux. These flux imbalances play a critical role in the salt budget and circulation of the Arctic Basin and may affect the global-scale ocean circulation by producing fluctuations in effective surface precipitation in the North Atlantic. Because of such considerations, climate studies must include ice motion as well as ice thermodynamics.

Ice Thickness Distribution

A key coupling between sea ice thermodynamics and ice dynamics is the ice thickness distribution. Ice thickness is an important factor in controlling deformation, which causes pressure ridging and creates open water. When combined with ice transport, these factors change the spatial and temporal growth patterns of the sea ice and, when coupled with mechanical properties of ice, can modify its response to climatic change.

Many features of the thickness distribution may be approximated by a two-level sea ice model (Hibler, 1979) where the ice thickness distribution is approximated by two categories: thick and thin. In this two-level approach the ice cover is broken down into an area A (often called the compactness), which is covered by ice with mean thickness H , and a remaining area $1 - A$, which is covered by thin ice, which, for computational convenience, is always taken to be of zero thickness (i.e., open water).

For the mean thickness H and compactness A the following continuity equations are used:

$$\partial H/\partial t = -\partial(uH)/\partial x - \partial(vH)/\partial y + S_h \quad (3)$$

$$\partial A/\partial t = -\partial(uA)/\partial x - \partial(vA)/\partial y + S_A \quad (4)$$

where $A < 1$, u is the x component of the ice velocity vector, v the y component of the ice velocity vector, and S_h and S_A are thermodynamic terms and are described in Hibler (1979).

The thermodynamic terms in Equations (3) and (4) represent the total ice growth (S_h) and the rate at which ice-covered area is created by melting or freezing (S_A). The parameterization of the S_A term is particularly difficult to do precisely within the two-level model and generally represents one of the weaknesses of the model.

To allow this model to be integrated over a seasonal cycle, it is necessary to include some type of oceanic boundary layer or ocean model. The simplest approach (used for example by Hibler and Walsh, 1982) is to include a motionless fixed depth mixed layer (usually 30 m in depth). With this model, any heat remaining after all the ice is melted is used to warm the mixed layer above freezing. Under ice growth conditions, on the other hand, the mixed layer is cooled to freezing before the ice forms. Another approach that treats vertical penetrative convection processes much better is to include some type of one-dimensional mixed layer. Such an approach, using a Kraus-Turner-like mixed layer, was carried out by Lemke et. al (1990) for the Weddell Sea. Another approach is to utilize a complete oceanic circulation model which also allows lateral heat transport in the ocean. The latter approach was used by Hibler and Bryan (1987)

in a numerical investigation of the circulation of an ice-covered Arctic Ocean. In this study, a two-level dynamic-thermodynamic sea ice model was coupled to a fixed-level baroclinic ocean circulation model. In this ice-ocean model the upper level was taken to be 30 m thick and, as in the motionless case, was not allowed to drop below freezing if ice was present. Inclusion of some type of penetrative convection in a similar ice-ocean circulation model is an item of high priority for future research and is presently being pursued.

A more precise theory of ice thickness distribution may be formulated by postulating an areal ice thickness distribution function and developing equations for the dynamic and thermodynamic evolution of this distribution. In this case, a probability density $g(H)dH$ is defined to be the fraction of area (in a region centered at position x at time t) covered by ice with thickness between H and $H + dH$. This distribution evolves in response to deformation, advection, growth, and decay. Neglecting lateral melting effects, Thorndike et al. (1975) derived the following governing equation for the thickness distribution:

$$\frac{\partial g(H)}{\partial t} + \nabla \cdot [\underline{u}g(H)] + \frac{\partial [f_g g(H)]}{\partial H} = \psi \quad (5)$$

where \underline{u} is the vector ice velocity with x and y components u and v , ∇ is the gradient operator, f_g is the vertical growth (or decay) rate of ice of thickness H , and ψ is a redistribution function (depending on H and g) that describes the creation of open water and the transfer of ice from one thickness to another by rafting and ridging. Except for the last term on the left-hand side and ψ on the right-hand side, Equation (5) is a normal continuity equation for $g(H)$. The last term on the left-hand side can also be considered a continuity requirement in thickness space since it represents a transfer of ice from one thickness category to another by the growth rates. An important feature of this theory is that it presents an "Eulerian" description in thickness space. In particular, growth occurs by rearranging the relative areal magnitudes of different thickness categories.

This multilevel ice thickness distribution theory represents a very precise way of handling the thermodynamic evolution of a continuum composed of a number of ice thicknesses. However, the price paid for this precision is the introduction of a complex mechanical redistributor. In particular, to describe the redistribution one must specify what portion of the ice distribution is removed by ridging, how the ridged ice is redistributed over the thick end of the thickness distribution, and how much ridging and open water creation occur for an arbitrary two-dimensional strain field, including shearing as well as convergence or divergence. In selecting a redistributor

one can be guided by the conservation conditions that ψ renormalizes the g distribution to unity due to changes in area and that ψ does not create or destroy ice but merely changes its distribution. An additional consistency condition can be imposed if one asserts that all the energy lost in deformation goes into pressure ridging and that the energy dissipated in pressure ridges is proportional to the gravitational energy.

A redistribution function that satisfies these constraints may be constructed (for an explicit form see Hibler, 1980) by allowing open water to be created under divergence and ridging to occur under convergence. Within this formalism ridging occurs by the transfer of thin ice to thicker categories, assuming a certain amount of ridging and hence open water created under pure shear or more generally under an arbitrary deformation state. This "energetic consistency" condition will affect the thermodynamic growth via open water fractions and hence the total ice created.

The main relevance of the variable thickness distribution to climate modeling is its more precise treatment of the growth of ice. A comparison of the two-level and multilevel approaches to modeling the dynamic-thermodynamic evolution of an ice cover will be presented below. However, here we note that the ice thickness is considerably greater with the multilevel model due to its more realistic treatment of the thickness distribution.

Sea Ice Thermodynamic Models

Sea ice grows and decays in response to long- and shortwave radiation forcing, to air temperature and humidity via turbulent sensible and latent heat exchanges, and to heat conduction through the ice. The heat conduction is significantly affected by the amount of snow cover on the ice and by the brine remaining in the ice after it has frozen. These internal brine pockets cause the thermodynamic characteristics of sea ice to be very much different than those of fresh water ice of the same thickness.

Many features of the thermal processes responsible for sea ice growth and decay can be identified from semiempirical studies of ice breakup and formation of relatively motionless lake ice and sea ice (e.g., Langleben, 1971, 1972; Zubov, 1943). Overall, the two dominant components of the surface heat budget relevant to sea ice growth and freezing are the shortwave radiation during melting conditions and sensible and radiative heat losses during freezing. Observations of fast ice (relatively motionless ice attached to land) at the border of the Arctic Ocean indicate that once the initial stages of breakup (snow cover melt and formation of melt ponds) have

passed, the remaining decay of a stationary ice cover is almost entirely due to the shortwave radiation incident on the ice surface (Langleben, 1972). However, at the initial stages of breakup and decay, the sensible heat flux plays an important role (when the air temperature rises to $\sim 0^\circ\text{C}$) in causing rapid melting of the surface snow cover. This melting forms melt ponds (Langleben, 1971), which reduce the albedo and greatly enhance the rate of ice melt. After only a few weeks, drainage canals and vertical melt holes develop, and the characteristic appearance of a summer ice cover evolves, with melt ponds and surrounding smooth hummocks. Once these melt ponds have formed, the remaining decay is dominated by the radiation absorbed by open water.

As is clear from this discussion, the decay of sea ice in nature is rather critically affected by the amount of open water, which, because of its low albedo, can absorb much more radiation than the ice. On motionless ice sheets, the open water is present only through melt ponds or holes. However, in an actual dynamic variable-thickness ice cover, as occurs almost everywhere in the polar regions, the growth and decay of sea ice can be substantially affected by spatial thickness variations. Perhaps the most obvious example is the effect of open water on the adjacent pack ice. During melting conditions the radiation absorbed by leads can contribute to lateral melting by ablation at the edges of ice floes.

A number of sea ice thermodynamic models have been used in the past, ranging from the simple empirical models of Zubov (1943) and Anderson (1961) to complex numerical models that compute the surface heat budget (Parkinson and Washington, 1979) and the conduction through an inhomogeneous ice sheet (Maykut and Untersteiner, 1971).

Equilibrium Thermodynamic Models

While it is possible to construct simpler thermodynamic models that capture the essence of sea ice growth and decay (see Thorndike, this volume), what is usually done in climate modeling is to make use of an equilibrium ice/snow system in conjunction with a complete surface heat budget. One can solve this system iteratively and come up with an ice growth rate. The basic idea is illustrated in Figure 4, in which the steady state temperature profile is shown. Assuming no melting at the snow ice interface, the amount of heat going through this interface must be the same in the snow and the ice so that

$$(T_i - T_0) \frac{\kappa_s}{H_s} = \frac{\kappa_i}{H_i} (T_B - T_i) \quad (6)$$

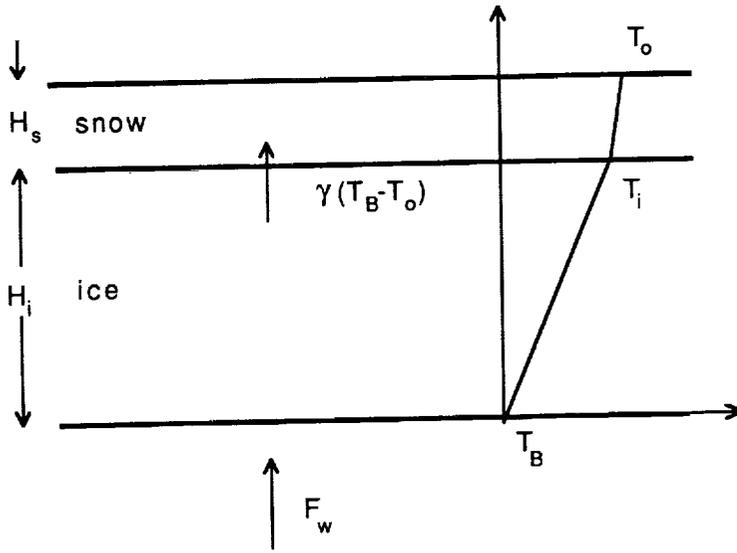


Figure 4. Sketch of combined snow and ice system used in equilibrium thermodynamic sea ice model. F_w is oceanic heat flux; other terms are defined in the text.

where T_i is the temperature of the ice at the snow-ice interface, T_0 is the surface temperature of the snow, T_B is the water temperature, κ_i and κ_s are the thermal conductivities of ice and snow, and H_i and H_s are the thicknesses of the ice and snow.

This equation allows us to solve for T_i in terms of T_B and T_0 . Substituting the resulting expression into the conductive flux through the ice, we obtain

$$(T_i - T_0) \frac{\kappa_s}{H_s} = \gamma(T_B - T_0) \quad (7)$$

where $\gamma = \frac{\kappa_i \kappa_s}{\kappa_s H_i + \kappa_i H_s}$ is the effective conductivity.

The complete surface heat budget equation, with a sign convention such that fluxes *into* the surface are considered positive, then becomes

$$(1 - \alpha)F_s + F_l + D_1 V_g (T_a - T_0) + D_2 V_g [q_a(T_a) - q_s(T_0)] - D_3 T_0^4 + (\gamma)(T_B - T_0) = 0 \quad (8)$$

where α is the surface albedo, T_a is the air temperature, γ is the effective conductivity, V_g is the wind speed, q_a is the specific humidity of the air, q_s is the specific humidity of the ice surface, and F_s

and F_1 are the incoming shortwave and longwave radiation terms. The constants D_1 and D_2 are bulk sensible and latent heat transfer coefficients, and D_3 is the Stefan-Boltzmann constant times the surface emissivity. The equation is usually solved iteratively (see, e.g., Appendix B of Hibler, 1980, for details and numerical values of various constants) for the ice surface temperature. The conduction of heat through the ice is used to estimate ice growth using

$$\gamma(T_b - T_0) + F_w = \rho_i H \frac{dH_i}{dt} \quad (9)$$

where F_w is the oceanic heat flux, ρ_i is the ice density, H_i is the ice thickness, and dH_i/dt is the rate of change of ice thickness. When the calculated surface temperature of the ice is above melting, it is then set equal to melting, and the imbalance of surface flux is used to melt ice.

Effect of Internal Brine Pockets and Snow Cover

Apart from the absence of dynamics, the global thermodynamic models mentioned above are still somewhat simplified in nature, the main simplification being that no internal melting due to brine pockets in the ice has been considered. In particular, in sea ice the density, specific heat, and thermal conductivity are all functions of salinity and temperature (the dependence on temperature is indirectly also due to the salinity). These dependencies are caused by salt trapped in brine pockets that are in phase equilibrium with the surrounding ice. The equilibrium is maintained by volume changes in the brine pockets. A rise in temperature causes the ice surrounding the pocket to melt, diluting the brine and raising its freezing point to the new temperature. Because of the latent heat involved in this internal melting, the brine pockets act as a thermal reservoir, retarding the heating and cooling of the ice. Since the brine has a smaller conductivity and a greater specific heat than the ice, these parameters change with temperature.

Maykut and Untersteiner (1971) developed a time-dependent thermodynamic model for level multiyear sea ice and carried out a variety of calculations that yielded considerable insight into the growth and decay of sea ice. A simplified model that reproduced most of the Maykut-Untersteiner results was proposed by Semtner (1976). The Semtner model assumed that sea ice could be represented as a matrix of brine pockets surrounded by ice where melting can be accomplished internally by enlarging the brine pockets rather than externally by decreasing the thickness. As a consequence, for the same forcing sea ice can have a substantially greater equilibrium thickness than fresh water ice.

Based on this "brine damping" concept, Semtner (1976) proposed a simple model in which the snow and ice conductivities were fixed. In this model the salinity profile does not have to be specified. To account for internal melting, an amount of penetrating radiation was stored in a heat reservoir without causing ablation. Energy from this reservoir was used to keep the temperature near the top of the ice from dropping below freezing in the fall. Using this simplified model, Semtner was able to reproduce many aspects of Maykut and Untersteiner's model results within a few percent.

For an even simpler diagnostic model, Semtner proposed that a portion of the penetrating radiation I_0 be reflected away. The remainder of I_0 was applied as a surface energy flux. In addition, to compensate for the lack of internal melting, the conductivity was increased to allow greater winter freezing. In the simplest model, linear equilibrium temperature gradients are assumed in both the snow and ice. Since no heat is lost at the snow-ice interface, the heat flux is uniform in both snow and ice.

The results of Semtner's prognostic and diagnostic models are compared to Maykut and Untersteiner's results in Figure 5. This figure also shows the importance of the assumed penetration of radiation which causes internal melting. By allowing no radiation to penetrate, the internal melting is mitigated and the radiation instead is used to melt the ice, causing a reduced thickness. Note that while the 0-layer diagnostic model reproduces the mean thickness well, the amplitude and phase of the seasonal variation of thickness are somewhat different from those of the 3-layer prognostic model. This simplest diagnostic model has taken on special significance for numerical simulations of sea ice because almost continual ridging and deformation make it difficult to record the thermal history of a fixed ice thickness. Selected thermodynamic-only simulations using this model will be discussed in the next section.

Finally, because of the problems with an excessive seasonal cycle in the simplest Semtner model, it may be useful to employ some type of brine damping in sea ice models that are used in climate studies. However, the difficulty here is that when a full dynamic model is used, the advection of ice makes it difficult to record the internal temperature characteristics of the ice. An approach that improves the summer thermodynamic response of the ice cover is to include a brine pocket heat storage term in an equilibrium one-level model. Such an approach was carried out by Flato (1991) in a numerical investigation of a variable-thickness sea ice model. While his approach does not take into account the heat capacity of the ice, it does create a much more realistic summer cycle in that it retards spring melting and fall freezing.

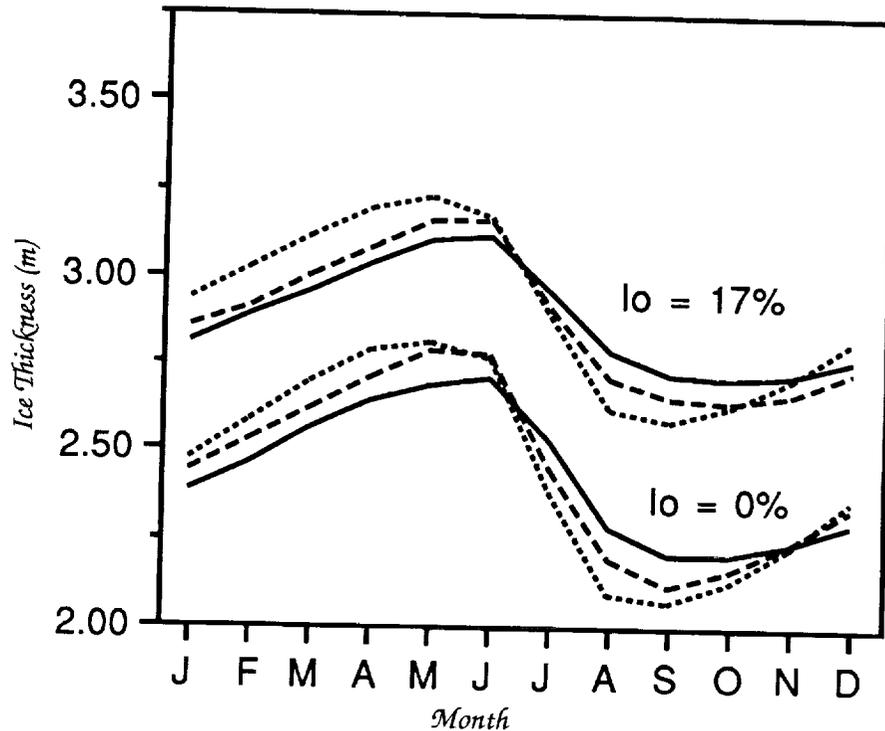


Figure 5. Annual thickness cycles of three thermodynamic sea ice models for the cases of 0 and 17% penetrating radiation. Solid line, Maykut-Untersteiner model; dashed line, Semtner 3-layer model; dotted line, Semtner 0-layer model.

The role of snow cover in modifying the thermodynamic growth of sea ice was also investigated by Maykut and Untersteiner (1971) and Semtner (1976). The thermal conductivity of snow is considerably less than that of ice; however, the albedo of snow is much higher than that of ice. Lower conductivity implies less ice growth, whereas higher albedo implies less summer melt. These conflicting processes combine to produce little change in mean annual ice thickness for snowfall rates less than about 80 cm/yr. Since this is much higher than the snowfall rate for most of the Arctic, snow cover can be expected to have little effect on the amount and distribution of sea ice. This notion is confirmed by recent dynamic-thermodynamic simulations with a multilevel sea ice model that included the effect of snowfall (Flato, 1991).

Model Simulations of Arctic Sea Ice

Sea ice in climate models has generally been included simply as a motionless sheet, in some cases using a thermodynamic growth model to calculate the surface temperature and ice thickness. As

discussed above, neglecting ice dynamics not only removes the lateral heat and salt transport capabilities of the ice pack, it also leads to unrealistic spatial patterns of thickness buildup and air-sea heat fluxes. In this section we will discuss some numerical model results to illustrate a few of these effects. In particular we will compare free drift (no ice interaction), the cavitating fluid rheology, and the elliptical yield curve, viscous-plastic rheology in simulations employing climatological wind and thermodynamic forcing (Flato and Hibler, 1990). For comparison we will also show the result of a thermodynamics-only simulation in which the ice cover is assumed to be a motionless, unbroken sheet.

Perhaps the most illustrative comparison of these simulations is provided by the seasonal equilibrium thickness contours (Figure 6) shown here for the end of March. The lack of ice interaction in the free drift case leads to unreasonable thickness buildup near the coasts after only a couple of months and is thus clearly undesirable for climate studies. Both the cavitating fluid and viscous-plastic models produce thickness buildup patterns which are roughly similar to observations (e.g., Bourke and Garrett, 1987), with the thickest ice (4–6 m) off the Canadian Archipelago and North Greenland coast. The similarity here reinforces the notion that, to lowest order, the role of ice interaction is to prevent excessive convergence when ice is driven against a coast. Shear strength in the viscous-plastic case acts primarily to slow the ice drift and impede outflow through the Fram Strait (Flato and Hibler, 1990), although it does modify the thickness buildup pattern somewhat. We might note, however, that including shear strength in the ice cover modifies the curl of the wind stress applied to the ocean and thus influences the barotropic ocean circulation, a modification which may be important in long-term ice-ocean simulations. It is also interesting that the net ice growth patterns (contours of net ice growth at a point over a year) are almost unaffected by shear strength in the ice cover (Figure 7), and, since net growth corresponds to a net source of salt at the ocean surface, the baroclinic ocean circulation should be impacted little. In addition to the unrealistic thickness patterns, it is this last issue which makes the thermodynamics-only model unsuitable for long-term climate studies with a coupled ice-ocean model. The problem here is that in a thermodynamics-only model the ice grows and melts in place, and hence, over an annual cycle, there is no net salt flux to the ocean. Atmospheric general circulation models can tolerate coupling to a thermodynamics-only sea ice model since the principal bottom boundary condition is temperature; however, some parameterization of leads is necessary if the fluxes of heat and water vapor are to be at least crudely included.

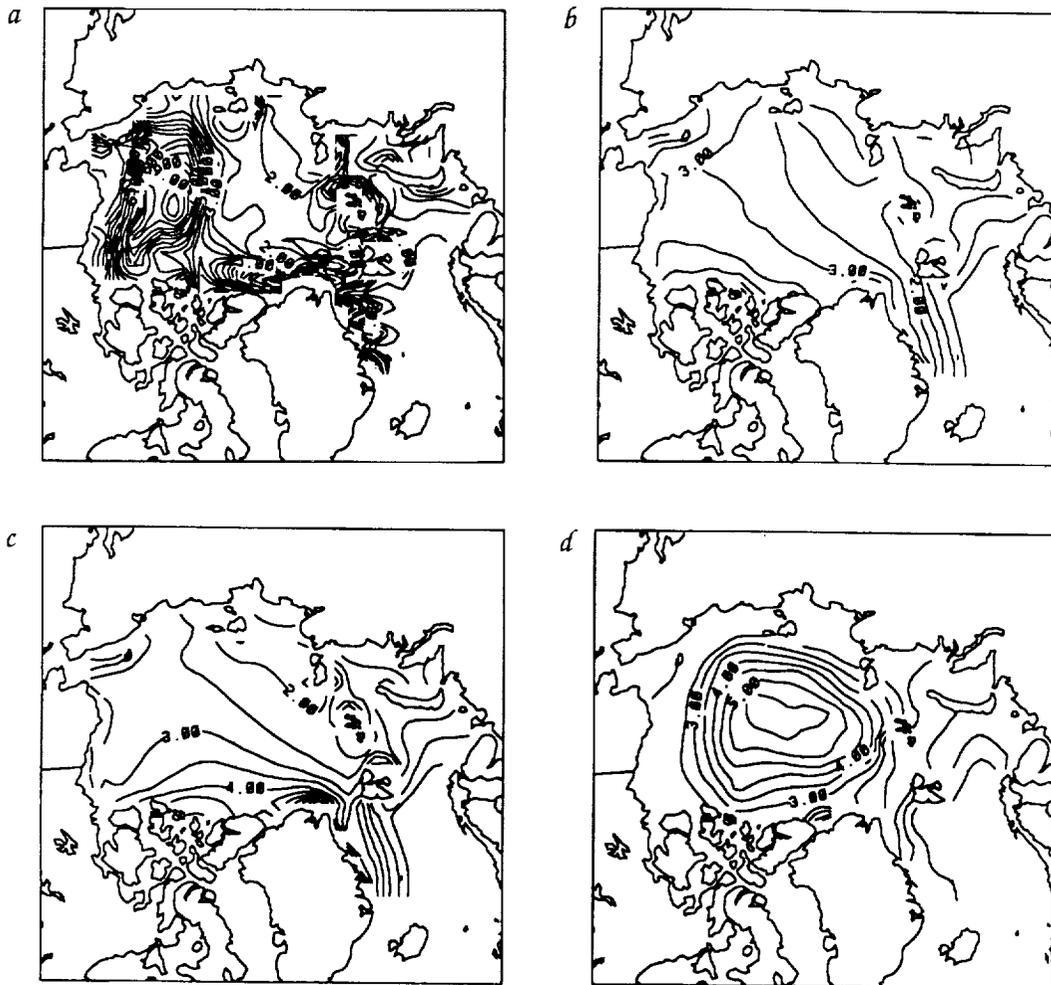


Figure 6. Equilibrium March thickness fields in the Arctic calculated using climatological forcing: (a) free drift model, (b) cavitating fluid model, (c) viscous-plastic, elliptical yield curve model, (d) thermodynamic-only model. Contours are in meters with a half-meter contour interval.

The multilevel model approach represents an improvement over the two-level model since the proportion of ice in a number of thickness categories is kept track of and the thermodynamic growth rate of each category is calculated separately. Furthermore, the multilevel case allows the ice strength to be parameterized in terms of the deformational work done during ridging (Rothrock, 1975). The effect of the improved representation of thermodynamic growth is illustrated in Figure 8, which shows the thickness field calculated by the multilevel model using the same strength parameterization as the two-level model. The increase in overall thickness here is due in part to the improved resolution of thin ice and open water, for which the

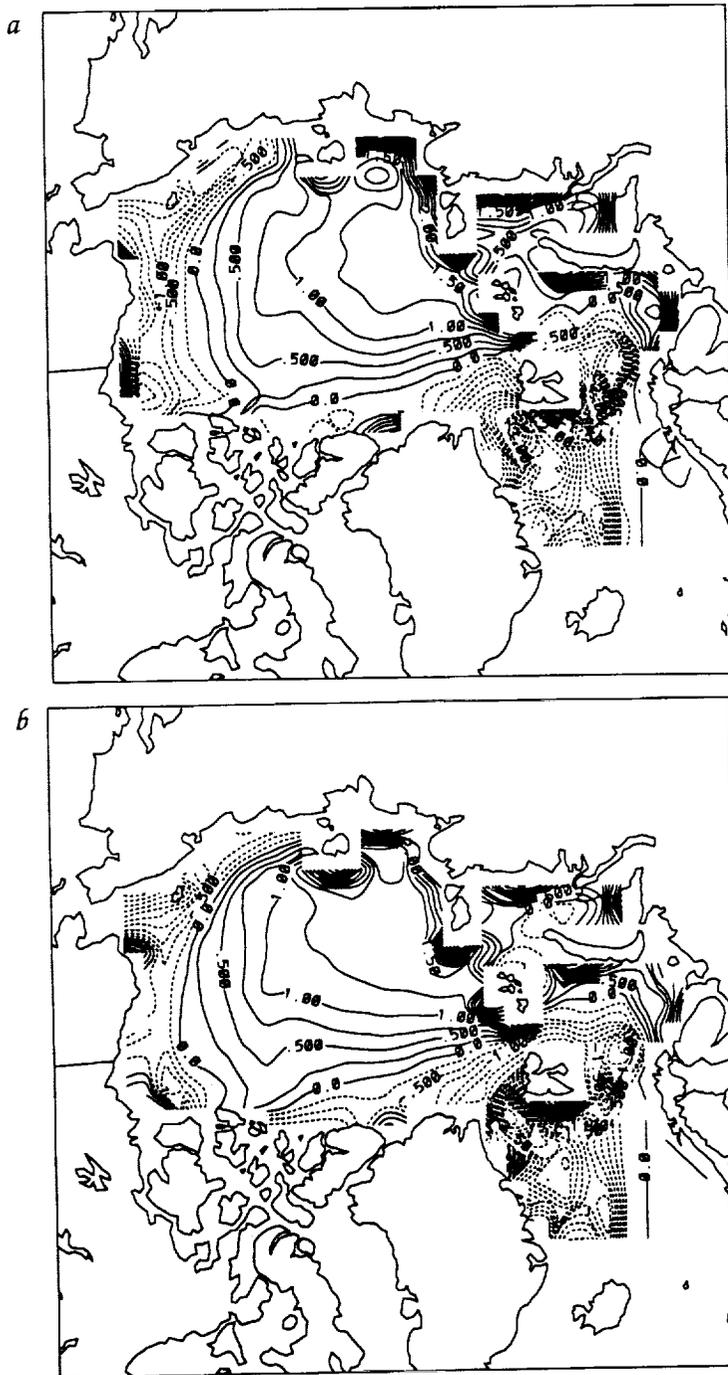


Figure 7. Net annual ice growth fields (in meters of ice per year) calculated using climatological forcing: (a) viscous-plastic, elliptical yield curve model, (b) cavitating fluid model. The contour interval is 0.25 m of ice per year.

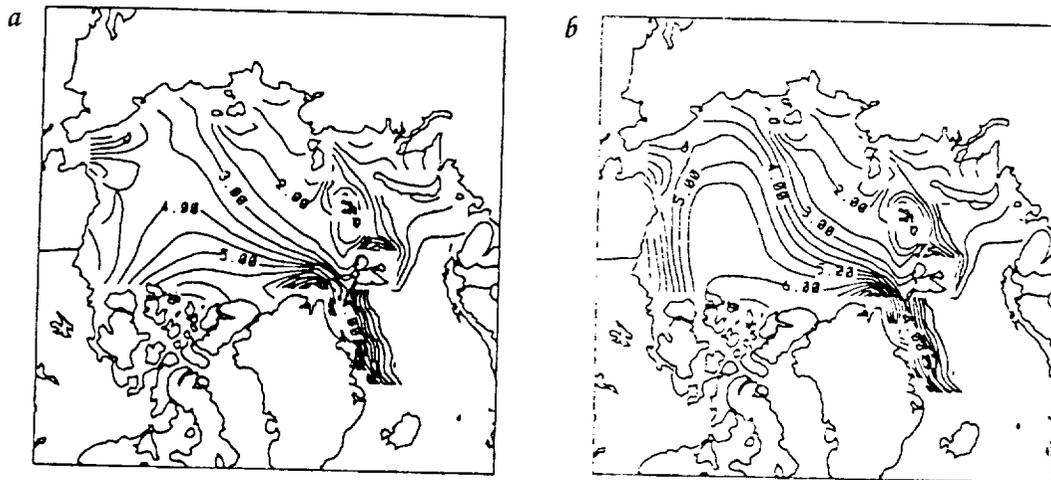


Figure 8. Equilibrium March thickness fields calculated using (a) climatological forcing, together with (b) a multilevel sea ice model with two-level model strength parameterization. Contours are in meters with a half-meter contour interval.

growth rate is very high, and redistribution from thin to thick ice representing the formation of ridges during deformation. It is likely that the complexity and computational demands of the multilevel model are unjustified for climate studies at this point, although a simpler three- or four-level model (currently under development) would be a significant improvement over the two-level scheme.

Sensitivity of Sea Ice Models to Climate Change

In this section we will demonstrate the sensitivity of the modeled sea ice cover to changes in atmospheric forcing which might accompany changes in global climate. Aside from dramatic alterations in atmospheric circulation patterns, the most significant impact on the ice cover will likely be due to changes in air temperature and cloud cover. An increase in air temperature results in greater sensible and latent heat fluxes and incident longwave radiation, all of which inhibit ice growth. An increase in the cloud cover, on the other hand, reduces the incoming shortwave radiation while increasing the longwave flux. These effects will be investigated here by examining several simulations of the Arctic ice cover using the two-level dynamic-thermodynamic sea ice model with forcing fields covering the period 1981–83.

The four simulations we will discuss here are the standard or unperturbed run, runs in which the cloud cover was increased or decreased by 20%, and a run in which the air temperature was increased by 1°C. Time series of total ice volume and ice extent (the

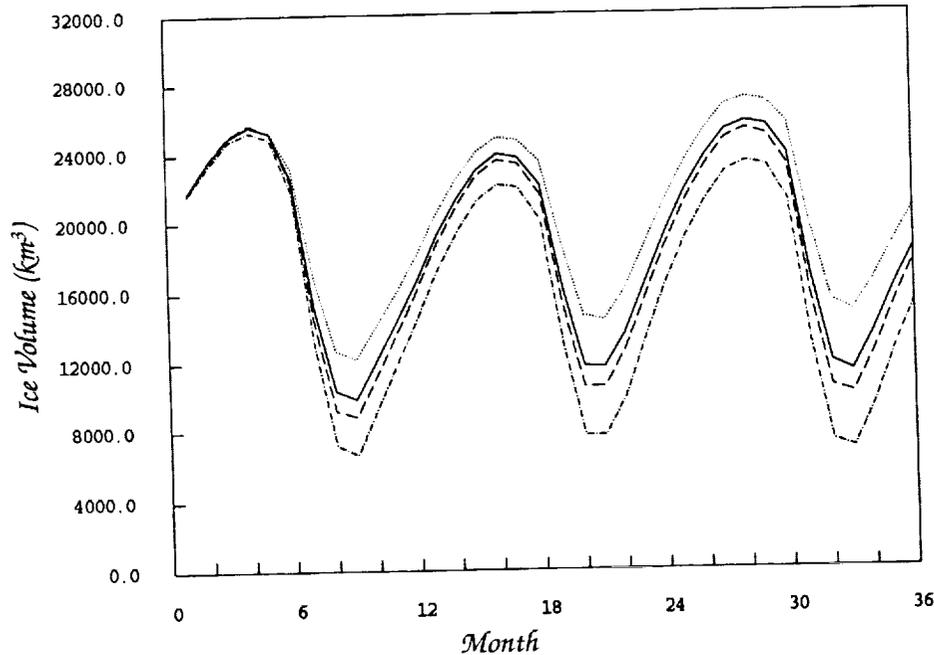


Figure 9. Total ice volume and ice extent (area enclosed by 15% compactness contour) time series calculated using two-level viscous-plastic model and forcing fields from 1981–83. Results are shown for (solid line) a standard simulation, (dotted line) a case with cloud cover increased by 20%, (dashed line) a case with cloud cover decreased by 20%, and (dot-dashed line) a case with a 1°C increase in air temperature.

area enclosed by the 15% compactness contour) are plotted in Figure 9 and show the sensitivity to each of these changes. The increase in total ice volume and summer ice extent in the increased cloud cover case points out that the dominant role of clouds is to control the incoming shortwave radiation—their contribution to longwave radiation being secondary. It is also apparent from the asymmetry of the response to increasing and decreasing the cloud cover that this effect is rather nonlinear. A significant decrease in both ice volume and summer ice extent is seen to accompany a 1°C rise in air temperature; in fact, the summer ice volume is reduced by almost a third. An increase in air temperature of about 4°C (a value obtained by Manabe and Stouffer, 1980, for a doubling of CO₂) is sufficient to almost eliminate the summer ice cover. What is missing here, of course, is the feedback between the increase in open water area and the amount of cloud cover. The increase in open water area, not only in the peripheral seas but also in the central pack, is illustrated by the August 1983 compactness fields for the four simulations (Figure 10). A general increase in cloud

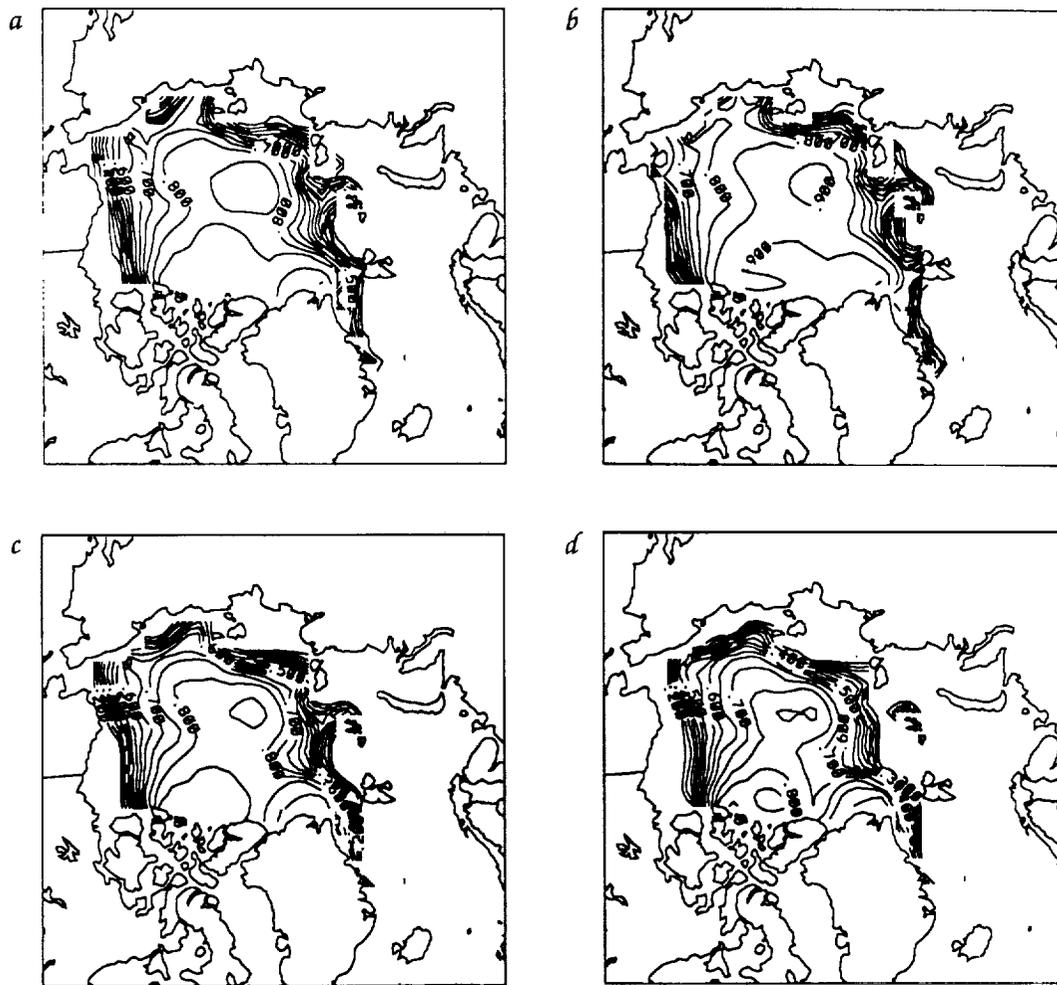


Figure 10. Compactness fields calculated by the two-level model for August 1983: (a) standard simulation, (b) 20% increase in cloud cover, (c) 20% decrease in cloud cover, (d) 1°C increase in air temperature.

cover would be expected in the warming case due to the increased availability of water vapour from leads, and this, based on the cloud sensitivity results, may counteract the effect of increased air temperature.

We might note here that the ice edge location, particularly in the eastern Arctic, is somewhat too far north in these simulations—a shortcoming of the two-level model. The ice edge position is somewhat better in the multilevel model case; however, the sensitivity to clouds and warming is about the same. We should also point out here that the shape and interannual variability of the ice edge in the

Barents and Greenland seas is controlled primarily by upwelling of heat from the ocean. To properly simulate the ice cover in this climatically important region requires a coupled ice-ocean model which not only provides realistic ocean currents (to properly represent the lateral transport of heat) but also includes a sufficiently detailed parameterization of the vertical processes to bring the heat from the deeper ocean to the surface at the correct location.

Concluding Remarks

Because of almost constant motion and deformation, the dynamics and thermodynamics of sea ice are intrinsically coupled. In addition, the ice and ocean circulation are tied together by the freezing and melting of ice, which causes salt and fresh water fluxes into the ocean, and ice transport, which yields unbalanced fluxes. As a consequence, understanding the response of the high latitudes to climatic change requires considering the coupled ice-ocean system (including ice interaction) in the polar regions. Results and theory reviewed here have indicated the complexity of different thermodynamic and dynamic effects and the role they can play in air-sea interaction. This complexity makes it difficult to guess the correct ad hoc treatment of sea ice to use in climate models. Instead, the results emphasize the importance of including a more realistic treatment of sea ice vis-à-vis a fully coupled, ice interaction-based, dynamic-thermodynamic sea ice model. These models at least contain the main first-order aspects of the sea ice system, whereas simple thermodynamics-only models clearly do not.

By coupling such models with treatments of the ocean, we may obtain quantitative insights. It also appears that due to the variety of complex dynamic processes, specifying ice fluxes and transport for use in ocean circulation modeling will leave out many major feedbacks that affect climatic change.

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