MODELS OF AIR-ICE-SEA INTERACTION

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Summary

Sea ice forms on the sea surface once the atmosphere has cooled sufficiently to lower the surface of the water to its freezing point. It often acts as an obstacle for our activity, while we feel a strong connection to nature when we view the ice-covered ocean. From a physics point of view, sea ice provides unique properties such as plastic behavior and salt rejection from freezing sea water during sea ice formation. These characteristics are explained first, followed by a description of ice-ocean interactions considering both thermodynamic and dynamic aspects. Sea ice is a distinct indicator for climate change, as its growth/decay is strongly linked to a colder/warmer climate. Therefore, we pay attention to its variability, both in thickness and extent, while particularly noting changes to sea ice cover due to the current warming phase, and discuss whether the sudden sea ice...
reduction in the Arctic might be a consequence of global warming or indicator of year-to-year variability. Overall information on the ice-covered ocean, together with an introduction to air-sea-ice modeling, are key components of the information presented herein.

1. Introduction

Sea ice is a concern for both off-shore oil operations in the Polar Regions and transportation through an ice-covered ocean. The role sea ice will play under future climate change scenarios is uncertain. Since sea ice floats on the sea surface, interactions between sea ice and ocean are naturally included in sea ice modeling. The atmosphere applies a significant dynamic force to a sea ice cover in the form of wind stress, while feedback from the ice-ocean system to the atmosphere has a significant effect on both local weather and global-scale climate. Here we first introduce thermodynamic and dynamic concepts with respect to sea ice, and then describe ice-ocean coupling and air-ice-ocean interactions.

In the last thirty to forty years, since the 1960s and 1970s, the Arctic has experienced a reduction in sea ice which is considered to be an indicator of global warming. On the other hand, climate variability has various time scales; hence we should not take all reduction in sea ice cover to be a consequence of global warming. In particular, climate variability at time scales of up to a few years can be significant, which could lead to a false alarm if we attribute the current reduction in ice cover to anthropogenic climate change alone. This document discusses recent climate variability and considers future projections into the 21st century.

2. Ice Modeling

2.1. Principles of Sea Ice Modeling

Sea-ice models have been discussed in numerous papers (e.g. Hibler 1979, 1980). The mechanisms included in the models are categorized into the areas of thermodynamics and dynamics. The thermodynamics are mainly controlled by heat flux between the atmosphere and the ice-ocean system, in addition to solar radiation. A positive heat flux from water to sea ice immediately initiates ice melting, and water cooled below its freezing point yields ice formation. Sea ice dynamics are typically driven by wind forcing, while ice-water stress, which is usually in balance with air-ice stress, typically counters this forcing. Under a high ice concentration near a coastal boundary, the internal stress of the ice becomes a crucial parameter and also acts to resist wind forcing. A minor addition to these forcing terms also comes from the Coriolis force and gravity associated with sea surface slope.

2.2. Thermodynamics of Sea Ice

Let us consider an idealistic sea ice cover with a uniform thickness. A schematic view is shown in Figure1. Heat flux occurs from the ice surface to the atmosphere and from the underlying seawater to the bottom of the sea ice. The surface heat flux consists of incoming short-wave radiation, net long-wave radiation, and sensible and latent heat...
fluxes. Sea ice is often covered by snow, which tends to reflect solar radiation. Conductive heat flux occurs in the ice due to the temperature gradient between the ocean and atmosphere. As the bottom of the ice cover is kept nearly at the in situ freezing point of the underlying seawater (about -1.8°C), the conductive heat flux is inversely proportional to ice thickness. Sea ice typically grows (melts) in response to upward (downward) heat flux. Ice formation, which occurs at the interface of the ice cover and the ocean, occurs faster under thinner ice due to a relatively high upward conductive heat flux, and conversely occurs more slowly under a thicker ice cover where the conductive heat flux is reduced. Snow cover has an insulative effect which results in a small thermal conductive coefficient and hence tends to reduce the rate of ice formation. An interested reader is directed to a typical modeling paper on sea ice thermodynamics (Maykut and Untersteiner, 1971) for further information on this process.

Each component in the surface heat flux is now described more in detail. The components are parameterized from bulk formulae given by Smith and Dobson (1984). Incoming short-wave radiation increases with higher solar angle and is zero during night.

![Figure 1. Schematic view of heat fluxes around sea ice.](image)

The values of albedo (a measure of reflectivity) are about 0.05 for water, and in a range of 0.5 for bare ice to 0.8 for snow. The effective solar heat flux reaches a mean value in summer of 100 Wm⁻². Long-wave radiation emanating upward from the ice surface is partly cancelled by downward radiation from the atmosphere. Clouds contribute to the downward radiation and reduce net long-wave radiation from the ice. The long-wave
radiation is typically stable throughout the year.

Sensible and latent heat fluxes are expressed as follows:

\[ SH = \rho_a C_a C_h (T_i - T_a) U_a \]  \hspace{1cm} (1.1)
\[ LH = L C_e (q_s - q_a) U_a \]  \hspace{1cm} (1.2)

The upward fluxes are defined to be positive in both equations. In (1.1), \( \rho_a \) (1.3 kg m\(^{-3}\)) is air density, \( C_a \) \((10^3 \text{ J kg}^{-1} \text{ K}^{-1})\) is the specific heat of air, \( C_h \) is the sensible heat flux coefficient, \( T_i \) is air temperature, \( T_a \) is ice surface temperature, and \( U_a \) is wind speed at 10 m above the surface. In (1.2), \( L \) \((2.5 \cdot 10^6 \text{ J kg}^{-1})\) is the latent heat of water vapor, \( C_e \) is the latent heat flux coefficient, \( q_s \) is the saturation water vapor density, and \( q_a \) is the water vapor density in air. Both fluxes increase under larger wind speed. The sensible heat flux is proportional to a temperature difference between the ice surface and the atmosphere. The latent heat flux becomes larger with a dryer atmosphere. The sensible and latent heat flux coefficients, \( C_h \) and \( C_e \), are on the order of \( 10^{-3} \), based on extensive studies of air-open water interactions by the likes of Smith and Dobson (1984). The coefficients are found to increase with higher wind speed, while the values are comparable for open water and sea ice. For polar to subpolar regions, the sensible heat flux is the largest during winter (a few hundreds of Wm\(^{-2}\)). In contrast, solar radiation becomes a major component in summer. These thermodynamic calculations require input fields of wind, air temperature, dew point and cloudiness.

Snow cover tends to insulate ice in winter and reduces ice formation; hence we need precipitation data in order to accurately include its effects on ice growth. Snow reflects more short-wave radiation in spring than ice and slows down ice melting. Snow on sea ice can transition into what is known as “snow ice” due to sea water flooding and subsequent freezing, hence it can be difficult to estimate snow depth in spring. Omission of the snow cover in models yields more ice formation in winter and faster melting in spring.

Sea ice exhibits unique properties due to the fact that it is made up of solid ice crystals interspersed with pockets of liquid brine. As ice forms, salt is rejected from the solid part to the liquid part. Salinity concentrations are about 35 \( \% \) in sea water and 5 \( \% \) in sea ice. Salt is partially accumulated in the liquid fraction of the sea ice. The liquid fraction is called a brine pocket and it increases the effective heat capacity of the sea ice, as latent heat is required to maintain brine pocket expansion under warming. This extra heat capacity in brine pockets tends to maintain ice volume in spring and results in less short-wave radiation absorption by open water.

**2.3. Thermodynamics in a Partial Ice Cover**

Near the ice edge, it is more appropriate from a modeling perspective to consider open water interspersed with uniform sea ice. This two-level representation of ice has been used as in Hibler (1979). As long as sea ice is present, water temperature in a surface
water layer is set at the in situ freezing point. This assumption implies an infinitesimal time scale of heat transfer between ice and water, although this condition could be easily relaxed by incorporating a time-derivative term with a finite time scale.

In most cases, a numerical model is used to simulate ice growth and decay. A model typically represents the region of concern using a horizontal grid system. Within each grid cell, thickness and concentration of ice are variables used to describe ice distribution in time and space. Within one cell, ice formation or decay is calculated separately in open water and ice-covered portions, i.e. heat deficit in open water is used to form thin ice, while heat removed through the ice-covered portion is converted to an increased thickness of (thick) ice. The thin ice is immediately merged into the thick ice via conservation of mass. Under a warming condition, heat flux into an open water portion of the surface layer melts the thick ice in an adjacent grid cell. Heat flux into the ice-covered portion decreases the thickness of the thick ice as long as the ice is thicker than a given minimum thickness (e.g. 0.5 m). Once the thickness is reduced to the minimum thickness, all remaining heat is used to melt the ice from the side and form open water. Ice is assumed to have no heat storage.

Heat flux is larger in open water both in winter and summer. In winter, water is warmer than sea ice and hence emits larger sensible and latent heat fluxes. In summer, larger solar radiation is absorbed into the open water portion due to its lower albedo (~0.05) than sea ice (0.5 – 0.8). Once the percentage of open water increases, more heat is absorbed, and sea ice reduces. This mechanism is called ice-albedo feedback and it is active in the summer months.

2.4. Multi-Category Ice Model

A two-level formulation of ice thickness contains only thick ice and open water categories, i.e. the predicted outputs are concentration (a ratio of the area of uniform ice to the total area) and effective thickness (thickness multiplied by concentration). This configuration is categorized into a two-category model, which was later extended to a multi-category model (Walsh et al., 1985). In a multi-category model one category is converted to the adjacent category due to both thermal effects (growth/decay) and kinematic effects (convergence). The conversion is parameterized as a function of these thermal and kinematic processes.

The simplest version is a three-category (3C) model (Ikeda et al., 2004). The new ice category, which is very thin ice with 0.1 m thickness, is added to the two-category model (Figure 2). The advantage of this approach is to reproduce ice concentration more appropriately in the seasonal ice region. The thermodynamics of the thin ice are dealt with independently of the thick ice, but under identical formulations. Heat loss through the thin ice and the open water portion is converted to an increase in thin ice concentration, with its thickness kept to the specified value (0.1 m). The thin ice portion contributes to ice concentration, but has little effect on ice volume. The thin ice can prevent air-sea heat flux during the ice formation season and reflect short-wave radiation during the ice melting season. Thus, a seasonal ice cover is simulated better in the three-category model than in the two-category model.
Once the thin ice is included, the thick ice increases its concentration by rafting of thin ice. This assumption is justified by the observation that thin ice often rafts over itself in the ocean. The process is represented in the model by deformation, which is a process that converts thin ice into thick ice. Ice velocities are calculated from the momentum equations including internal stresses, as described in the next section. The ice rheology is identical to the viscous-plastic formulation proposed by Hibler (1979). It is assumed that the bulk ice strength, which is a key parameter in ice deformation, can be computed from taking the areal average of both categories of ice. The bulk strength approaches the strength of either category under the condition that the other category diminishes. When advection forces the ice concentration to exceed 100%, thin ice is converted into thick ice within one time step. Once all the thin ice has been converted to thick ice, the thick ice is assumed to deform. In addition to dynamic forcing, thick ice also increases in thickness due to thermodynamics forcing as described earlier.

2.5. Dynamics of Sea Ice

Ice dynamics in models are governed by a momentum balance between air and water stresses, Coriolis force, gravity due to sea surface slope and internal ice stress. The internal ice stress is typically determined using a plastic constitutive law, which is replaced with a viscous law for very small rates of strain. Although the ice rheology used in models has not been confirmed in field (Rothrock, 1979), it has been found to adequately represent alongshore ice movement, which is not sensitive to the internal shear stress except for within the vicinity of a coastal boundary. Furthermore, it also adequately represents the case where the ice resists onshore wind forcing and becomes solid like. Sea ice is often fixed to a coastal boundary, and it is known as “fast ice” when this occurs. The case of free-drifting ice, where the internal stress of the ice is considered to be negligible, is taken to represent fundamental sea ice dynamics and is thus examined here first.

Free-drifting sea ice receives air-ice drag and ice-water drag as its main forcing:
\[ \tau_a = \rho_a C_d U_a^2 \quad (1.3) \]
\[ \tau_w = \rho C_w (U_i - U_w)^2 \quad (1.4) \]

Here, \( \rho_a \) is air density, \( \rho \) is water density, \( U_i \) is ice velocity and \( U_w \) is water velocity. The velocity of free-drifting ice is almost completely determined by the wind velocity and a ratio of the air-ice drag coefficient (\( C_d = \sim 2 \times 10^{-3} \)) to the ice-water drag coefficient (\( C_w = \sim 7 \times 10^{-3} \)), with a minor contribution from Coriolis force along with gravity due to sea surface slope. Water density is roughly 1000 times larger than air density, hence ice velocity is about 2% of the wind velocity (e.g., Pease et al., 1983).

A minor input from the Coriolis force nearly balances with gravity due to sea surface slope under a no-wind condition. In a horizontally uniform ocean with wind forcing, only the Coriolis force is at work (i.e. sea surface slope is uniformly zero) and it is responsible for vertical profiles of water speed and direction under sea ice, in a similar manner as those in an open water area. For the northern hemisphere the near-surface ocean layer, called the Ekman layer, has a vertically averaged flow which is directed towards the right when facing in the downwind direction. Its thickness is around 30 m, and its velocity diminishes at the bottom of the boundary layer. Thus, wind forcing balances with Coriolis force in the entire boundary layer. It is natural to expect sea ice motion to also turn to the right when facing the downwind direction. As shown in Figure 3, the sea ice part receives forcing from the air-ice drag, ice-water drag and the Coriolis force. The ice velocity is directed between the wind and the water velocity under sea ice so that the air-ice drag is in balance with the ice-water drag and the Coriolis force. An interested reader is directed to physical oceanography textbooks on the Ekman spiral (e.g., LeBlond and Mysak, 1978) for further information.
Figure 3. Ekman spiral under sea ice with wind forcing in the northern hemisphere.

The governing momentum equations contain time-derivative terms, however a practical modification is made to the equations in the models, i.e. that the time derivative terms are negligible and the forces acting on the ice are balanced at each time step. This modification is reasonable, because a balance between air-ice stress and ice-water stress is established within about 1000 s (Parkinson and Washington, 1979), and it also remains valid when the internal ice stress is included.

2.6. Rheology of Sea Ice

Internal ice stress only plays an important role in sea ice dynamics near a coastal boundary at time scales longer than a few days. However, it becomes more influential in any location at shorter time scales when ice forcing generated by tides and wind waves is considered. The principle of internal stress in the ice is typically expressed using a plastic constitutive law, in which stresses are related to strain rates within a horizontal two-dimensional domain. An interested reader is directed to read Coon et al. (1974) and Hibler (1979) for further information.

A more recent approach is to replace the plastic constitutive law with an elastic one, as proposed by Hunke and Dukowics (1997). The elastic constitutive law requires an extremely small time step for stable computation. However, a computer composed of parallel processors can handle this type of computation more easily than the heavy iterative computation required in the case of the plastic constitutive law. There is
essentially no difference in results between these two types of models.

Fluid rheology in numerical models is usually expressed by the Newtonian viscous law, in which shear stress is proportional to strain rate. One of the most significant differences between the plastic law and the viscous law is realized as an asymmetric response to forces inducing ice compression from those forces inducing ice expansion. Sea ice which is represented using the plastic law resists compression up to a critical pressure, but freely moves under expansion. Ice strength is on the order of $10^4 \text{ Nm}^{-2}$. Sea ice with a 1 m thickness can resist the pressure accumulated over an ice cover with a transverse length of 100 km under shoreward wind forcing of 0.1 Nm$^{-2}$. Once the applied stress reaches the ice strength, ice is crushed into deformed ice. Thinner ice is more easily deformed, even in the middle of the Arctic Basin. Most observed ice ridges are thought to be created from thin ice growing in a polynya (an ice free region within a mostly ice-covered area) under spatially variable wind forcing.

Another difference between the plastic law and the viscous law is also clear when shear stress is applied to sea ice. As derived by Ikeda (1988), sea ice becomes motionless at a coastal boundary below a critical condition, but slips at the boundary where the shear stress is acting above the critical condition, independent of alongshore velocity. The critical value is determined by a cross-shore pressure gradient, which increases under a shoreward wind stress. This critical value increases even when there is only an alongshore wind forcing, as ice velocity tends to turn shoreward due to the Coriolis force, e.g. towards a coastal boundary on the right when facing downwind in the northern hemisphere, and towards the left in the southern hemisphere.

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Bibliography


anomaly]

©Encyclopedia of Life Support Systems (EOLSS)
ice modeling]


IPCC, (2007) Intergovernmental Panel on Climate Change, Working Group I Report: The physical science basis of climate change. eds. S. Solomon and D. Qin.[Comprehensive review on climate change in the polar oceans]


**Biographical Sketch**

**Motoyoshi Ikeda** born in Tokyo, Japan, 1946. Ph.D in Aeronautics, University of Tokyo, 1974. Research activities have been made in climate variability, ocean data assimilation and socio-climate modeling.

He has taken the professional positions as follows:

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His publication includes:

Ikeda, M., and F. Dobson (Eds.), *Oceanographic Applications of Remote Sensing*, CRC Press, 1995

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