Modeling sea-ice and its interactions with the ocean and the atmosphere

H. Goosse, T. Fichefet, R. Timmermann, M. Vancoppenolle

Institut d'Astronomie et de Géophysique G. Lemaître. UCL, Louvain-la-Neuve, Belgium
e-mail : hgs@astr.ucl.ac.be
http://www.astr.ucl.ac.be/users/hgs/index.html

1. Introduction

Sea ice is a key constituent of the climate system as it profoundly modifies the surface momentum, heat, and mass exchanges at high latitudes. Due to its high albedo and low thermal conductivity, sea ice alters the radiative and turbulent components of the surface heat balance, cutting the absorption of shortwave radiation by as much as 80 % in the case of snow-covered ice and reducing the turbulent heat fluxes by 1 to 2 orders of magnitude. In addition, sea ice intercepts most of the snow falling during winter, thus preventing it from immediately contributing to the ocean freshwater balance. Finally, a thick and compact ice pack also hinders the free exchange of momentum between atmosphere and ocean.

Another facet of the influence of sea ice on polar climate relates to the fact that it alters the seasonal cycle of atmospheric and oceanic fields. In the first place, the release and absorption of latent heat that respectively accompany the growth and decay phases of the ice cover tend to delay the seasonal surface temperature extremes. In the second place, the influx of salt into the ocean during ice formation and the corresponding freshwater input during ice melting tend to alter the density structure of the upper ocean and hence the ocean thermohaline circulation.

The coupling of the ice drift to the aforementioned processes introduces new important phenomena. On the one hand, ice motion is responsible for the formation within the ice cover of open water areas through which air-sea exchanges are much enhanced compared to ice-covered regions. On the other hand, ice motion combines with ice production and destruction to create a net annual poleward transport of heat in the atmosphere and of salt in the ocean.

It is believed that a number of positive and negative feedbacks linking sea ice to variations in surface albedo, cloud cover, atmospheric water vapor, and ocean thermohaline circulation play an important part in determining the Earth's climate and its sensitivity to external and internal perturbations. Indeed, atmosphere-ocean general circulation model (AOGCM) simulations of the climate transient response to increasing concentrations of greenhouse gases predict that the largest surface warmings will occur over the Arctic Ocean and sub-Arctic seas (e.g., IPCC2002).

The sea ice properties and processes that need to be taken into account in a comprehensive sea ice model can be divided into three categories: (1) dynamic processes, (2), thermodynamic processes and (3) processes that couple ice thermodynamics and dynamics. The dynamic component of the model consists of a momentum balance equation, which determines the ice drift as a function of the atmospheric and oceanic dynamical forcings, and an ice rheology, which prescribes the dependence of internal ice stresses on ice deformation and ice thickness characteristics. The thermodynamic component of the model determines the freezing/melting rate of ice given the atmospheric and oceanic heat and mass fluxes. Ice thermodynamics can be sub-divided into two components, one associated with vertical processes and one dealing with lateral processes. The former aims at calculating the accretion/ablation rates at the top and bottom boundaries of the ice, whereas the latter determines the changes in ice concentration that result from ice-ocean heat fluxes in leads. The thermodynamic-dynamic coupling is done via (1) an advection scheme, which works out the transport of volumetric ice variables, such as ice mass and sensible and latent heat contents, and (2) the ice thickness distribution formulation. The ice thickness distribution describes the evolution of the different thickness categories in which ice breaks as a result of dynamical processes.

2. Sea ice dynamics

At large scale, sea ice can be considered as a two-dimensional continuum moving in response to atmospheric and oceanic dynamical forcings. The ice momentum balance is established between the forces generated by the winds, the oceanic currents, and the tilt of the sea surface, on the one hand, and the ice inertia, Coriolis force, and internal ice stresses, on the other hand. The ice-ocean stress and atmosphere-ice are usually defined by a quadratic function of the ice-ocean velocity and of the wind speed, respectively. The internal ice forces arise from the interaction between ice floes in relative motion, and are ultimately determined by the mechanical behavior of individual floes. Ice-ice interactions act as integrators of the external air and water forcings, and are a decisive factor in determining the ice drift characteristics (Hibler, 1979). Furthermore, they are also responsible for the creation of material failures in the ice which are at the origin of the opening of leads within the ice cover.

3. Sea-ice thermodynamics

Ice formation, accretion, and ablation are all three ultimately determined by the heat and mass fluxes at the surface of the ocean and at the air-ice and ice-water interfaces. The surface energy budget includes shortwave and longwave radiative forcings, sensible and latent heat exchanges, and conductive heat flux at the top ice boundary. At the ice bottom and lateral walls, the heat balance is established between the heat conduction within the ice and the oceanic sensible heat flux that is computed as a function of the ocean temperature and the turbulent mixing in the ocean.

The little brine (15 ppt at most) that remains trapped in the ice after freezing has a large impact on the effective heat capacity of sea ice. This is because temperature variations inside ice cause brine pockets to either melt or freeze. The latent heat fluxes involved in these phase changes tend to retard ice heating in summer and ice cooling in autumn. In addition, brine has the important effect of decreasing heat conduction. However, the major factor in controlling heat conduction through ice is the accumulation of snowfall. Snow has a heat conductivity which is 5-6 times smaller than that of ice, and therefore the snow cover acts a thermal shield between air and ice. This is not the only important effect of snow. Snow also modifies the surface albedo, reducing the amount of absorbed solar radiation by 30-40% with respect to that of snow-free ice. Another snow-related process is the formation of snow ice, i.e., the transformation of snow into ice due to flooding of the ice floes. When the load of snow depresses the snow-ice boundary underwater, seawater infiltrates the submerged snow and freezes, forming a snow ice cap and releasing latent heat and salt.

Finally, one more physical process manifests itself during the decay phase of the ice cover: melt ponding. Melt ponds are small bodies of meltwater retained at the surface of the ice. These features substantially decrease the surface albedo, thus contributing to the acceleration of the thawing process (Ebert and Curry, 1993).

The role of leads in a thermodynamic ice cover is two-fold. In summer, they act in a similar way to melt ponds, considerably increasing, by their low albedo, the amount of solar radiation available for ice melting (Maykut and McPhee, 1995). In fall and winter, leads act as ice factories, as turbulent heat losses to the atmosphere are up to 2 orders of magnitude greater over open water than over ice.

4. Thermodynamics-dynamics coupling, ice thickness distribution

Ice thermodynamics and dynamics mutually modify each other. The ice accretion/ablation rate depends on ice thickness, with open water and thin ice regions experiencing faster ice growth in winter and more intense melting in summer than thick ice regions. Thus, at large scale, the location of regions of ice growth and decay is intimately related to the geographical distribution of ice thickness, which depends in turn on ice advection patterns. At small scale as well, dynamical opening and closing of leads and build-up of pressure ridges have similar effects in altering the thermodynamic evolution of ice. Conversely, ice motion characteristics are influenced by ice thickness ones, which largely control the amount of stress that ice can transmit.

The coupling between thermodynamics and dynamics is encapsulated in a system of equations which express the conservation of volumetric ice variables, such as snow/ice volume and sensible heat content per unit area and brine latent heat content per unit area. If ψ represents any of these quantities, the generic law is

$$\frac{\partial \psi}{\partial t} = -\nabla\left(\vec{V}\psi\right) + S_{\psi}$$

where S_{ψ} is the rate of change of ψ due to thermodynamic processes.

A problem that obviously arises here is that the continuity equation for ice volume does not allow, by itself, the determination of the mean ice thickness and concentration, let alone a distribution of ice thicknesses. Some hypotheses must therefore be adopted as regards the way in which ice thickness and concentration are affected by advection in order to derive these quantities from the continuity equations. To achieve this goal, a large fraction of sea ice models to date have postulated a 2-level ice thickness distribution, in which the ice cover in a given point is described by two variables: its mean thickness $\underline{h}_i$ and its concentration A. The evolutions of $A\underline{h}_i$ and A are then determined by the conservation law described above with the appropriate thermodynamic terms and with the constraint $0 \leq A \leq 1$.

A more complete approach to this problem consists of defining a thickness distribution G. The thickness distribution is a function of the ice thickness h_i, such that

$$\int_{h_0}^{h_1} G \, dh_i$$

determines the fractional area covered by ice of thickness $h_0 \leq h_i \leq h_1$ (Thorndike et al. 1975). Note that

$$\underline{h}_i = \int_{\varepsilon}^{\infty} h_i G \, dh_i \quad A = \int_{\varepsilon}^{\infty} G \, dh_i$$

where ε is an arbitrarily small positive number. The continuity equation for the ice volume can now be written as follows

$$\frac{\partial G}{\partial t} = -\nabla\left(\vec{V}G\right) - \frac{\partial\left(s\ G\right)}{\partial h_i} + \Phi$$

where s is the thermodynamic growth/melt rate for ice of thickness h_i and Φ is a redistribution function. The redistribution function depends on h_i and G itself. It describes changes in G due to deformation events (lead opening or closing and ice rafting and ridging), thus redistributing ice from one thickness category to another. Although there is flexibility in the parameterization of Φ, it must satisfy a number of physical constraints: conservation of area, volume, and energy. Ice models which adopt this approach are called multi-level models (e.g., Hibler, 1980). Although computationally expensive, multi-level models allow for a greatly improved representation of the ice cover compared to 2-level ones. For this reason, as computing power increases, they are expected to supersede 2-level models.

In addition to the ice thickness distribution, it is also necessary to make the distinction between different ice classes (or types) because the mechanical, thermal and radiative properties of sea ice can be strongly influenced by its history. For instance, Flato and Hibler (1995) have included separate distributions for the undeformed ice and for the ridged ice that has been produced by convergent ice motion. Using a more general approach, Haapala (2000) has subdivided the pack ice into open water, two different types of undeformed ice and three types of deformed ice. Evolution equations for each ice class include a redistribution between the ice classes that is depending on the ice compactness, thickness and velocity divergence. In contrast to Flato and Hibler (1995), Haapala (2000) did not include ice thickness distribution for each of his ice classes.

5. The ice-ocean model ORCA-LIM

In this section, a brief description on a particular ice model (the Louvain-la-Neuve Ice Model, LIM) is proposed. LIM is a comprehensive thermodynamic-dynamic sea-ice model (Fichefet and Morales Maqueda, 1997). Sensible heat storage and vertical heat conduction within snow and ice are determined by a three-layer model (one layer for snow and two layers for ice). The effect of the subgrid-scale snow and ice thickness distributions is accounted for through an effective thermal conductivity, which is computed by assuming that the snow and ice thickness are uniformly distributed between zero and twice their mean value over the ice-covered portion of the grid cell. The storage of latent heat inside the ice resulting from trapping of shortwave radiation by brine pockets is taken into account.

The model also allows for the presence of leads within the ice pack. Vertical and lateral growth/decay rates of the ice are obtained from prognostic energy budgets at both the bottom and surface boundaries of the snow-ice cover and in leads. When the load of snow is large enough to depress the snow-ice interface under the water level, seawater is supposed to infiltrate the entirety of the submerged snow and freeze there, forming a snow-ice cap. Also included in the model is a physically-based parameterisation of the lateral accumulation of frasil ice.

For the momentum balance, sea ice is considered as a two-dimensional continuum in dynamical interaction with the atmosphere and the ocean. The viscous-plastic constitutive law proposed by Hibler (1979) is used for computing the internal force. The physical fields that are advected are the ice concentration, the snow volume per unit area, the ice volume per unit area, the snow enthalpy per unit area, the ice enthalpy per unit area, and the brine reservoir per unit area.

We are currently implementing in LIM a representation of the different ice classes as well as an ice thickness distribution in each ice classes. Furthermore, a representation of thermodynamic processes including and arbitrary number of level in the snow ice-system as well as a more sophisticated representation of the brine pockets is presently tested

LIM has been used for both the Arctic and the Antarctic. It has been coupled to the UCL-ASTR, LODYC and GFDL oceanic general circulation model, to the LMD atmospheric general circulation model, and to the KNMI quasi-geostrophic atmospheric model. Below are presented some results of the ice-ocean model ORCA-LIM, which is made of LIM and ORCA (Madec et al., 1997), using the coupling technique described in Goosse and Fichefet (1999).

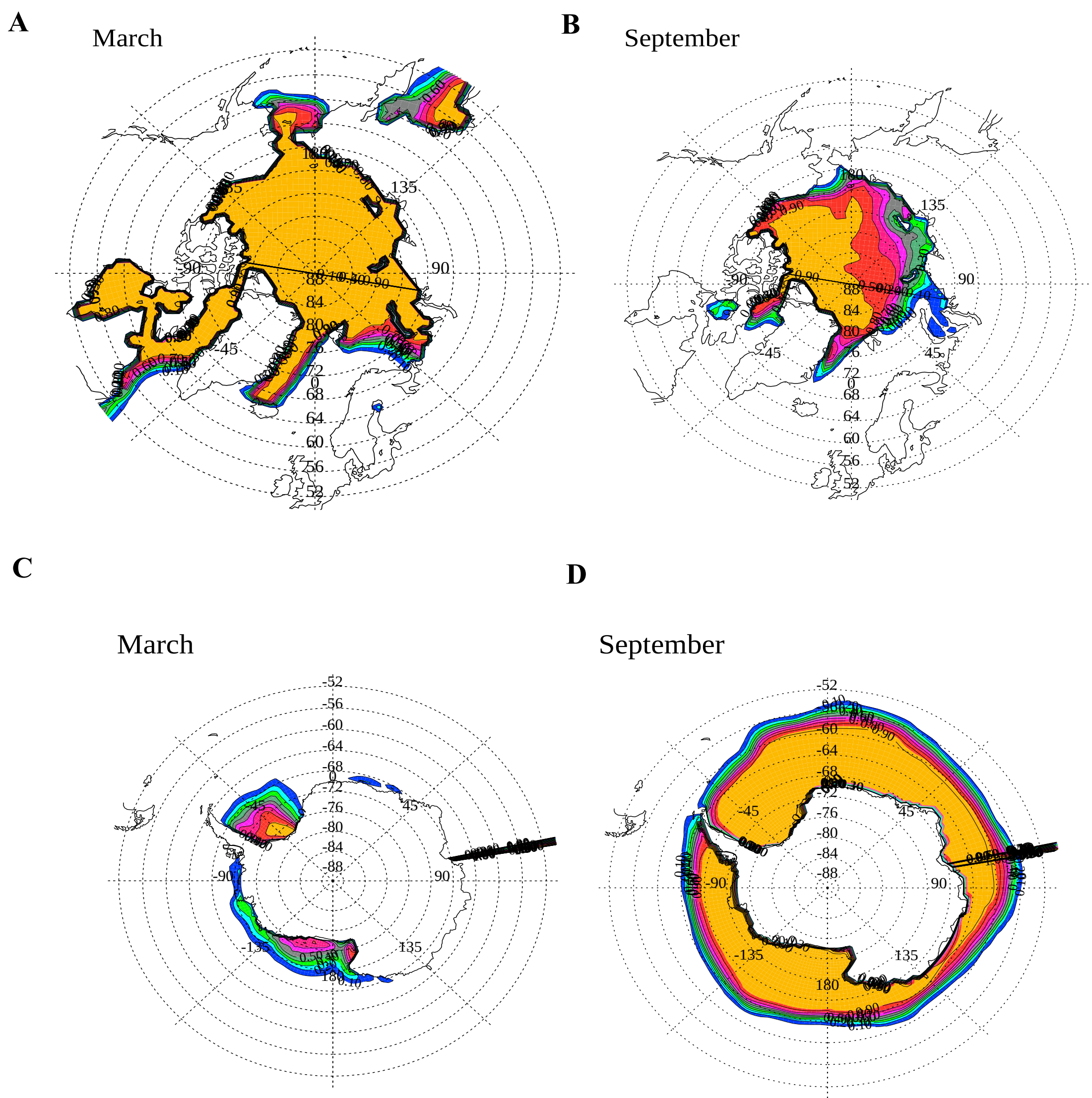

Figure 1. Mean ice concentration simulated by the model ORCA-LIM. a) Northern Hemisphere in March. b) Northern Hemisphere in September. c) Southern Hemisphere in March. d) Southern Hemisphere in September.

Figure 1 represents the mean ice concentration in summer and winter in both hemisphere obtained in a simulation driven by NCEP-NCAR (Kalnay et al., 1996) reanalysis for the period 1979-1997. It indicates that the ice extent in both hemispheres is reasonably well simulated. In the Northern Hemisphere, only some minor differences between the model and observations are noticed. For instance, in March, the ice concentration is overestimated off the north coast of Iceland, while the predicted ice extent is underestimated along the East Coast of Greenland. It seems than the ice in the East Greenland Current tend to flow directly toward Iceland rather than staying along the Greenland Coast. Considering the summer ice extent in the Northern Hemisphere, the model overestimated the ice cover at nearly all the longitudes. The largest differences between model and observations occurs in the Kara Sea and North of the coast of Alaska. In the Southern Ocean, the only significant problem is the underestimation in the ice concentration off the west coast of the Antarctic peninsula all year long

6. References

Ebert, E.E,. and J.A. Curry, 1993: An intermediate one-dimensional thermodynamic sea ice model for investigating ice-atmosphere interactions. *J. Geophys. Res*., **98**, 10085-10109.

Fichefet T., and M. A. Morales Maqueda, 1997: Sensitivity of a global sea ice model to the treatment of ice thermodynamics and dynamics. *J. Geophys. Res*., **102**, 12609-12646.

Flato G.M., and W.D. Hibler, 1995: Ridging and strength in modeling the thickness distribution of Arctic Sea-ice. *J. Geophys. Res*., **100**, 18611-18626.

Goosse H., and T. Fichefet,1999: Importance of ice-ocean interactions for the global ocean circulation: a model study. *J. Geophys. Res*., **104**, 23,337-23355.

Hibler W.D., 1979: A dynamic thermodynamic sea ice model. *J. Phys. Oceanogr*., **9**, 815-846.

Hibler, W.D., 1980: Modeling a variable thickness sea ice cover. *Mon. Weather Rev*., **108**, 1943-1973.

IPCC, 2001: *Climate Change 2001: The Scientific Basis. Contribution of Working Group I to the Third Assessment Report of the Intergovernmental Panel on Climate Change*, J.T Houghton, Y. Ding, D.J. Griggs, M. Noguer, P.J. van der Linden, X. Dai, K. Maskell, and C.A. Jonhson (eds). Cambridge University Press, Cambridge, United Kingdom and New York USA, 881pp.

Kalnay, E., and XXI others, 1996: The NCEP/NCAR 40-year reanalysis project. *Bull. Amer. Meteor. Soc.,* **77**, 437-471.

Maykut G.A., and M.G. McPhee, 1995: Solar heating of the Arctic mixed layer. *J. Geophys. Res*., **100**, 24691-24703.

Thorndike A.S., D.A. Rothrock, G.A. Maykut, and R. Colony, 1975: The thickness distribution of sea ice. *J. Geophys. Res*., **80**, 4501-4513.