Sea ice modelling for climate applications

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1. Introduction

The aims of this review are:

a) to discuss what models require to represent climate and climate change

b) to review the state of the art is in sea ice modelling for climate

On average, sea ice covers 1/20 of the area of the global ocean. The maximum and minimum sea ice extents occur in Northern (southern) hemisphere in March and September (vice-versa). Arctic ice area ranges from 5-13 x10^6 km^2 while Antarctic ice area ranges from 3-16 x10^6 km^2.

Sea ice has an interface with atmosphere and ocean. Sea ice growth and melt is determined by the surface heat and freshwater fluxes and ocean surface temperatures. Sea ice motion is driven by windstresses and ocean currents

Sea ice is important to climate for three reasons; it determines the surface albedo, it insulates the ocean and it determines the salinity of the ocean. High surface albedo reflects radiation back to the atmosphere. If sea ice begins to melt, lowering albedo will warm the ocean and lead to faster ice melt. The presence of sea ice insulates the ocean from losing heat and cools surface air temperature. Brine rejection occurs when ice forms and surface freshening occurs when ice melts. Ocean salinity in high latitudes modifies water mass formation and strength of thermohaline circulation

What does a climate model need to represent in terms of sea ice? We can assume that for climate change simulations a good representation of present day sea ice (mean state and variability) is necessary in addition to modelling sea ice processes correctly. The properties which are likely to be most important for climate are:

1. Ice extents as they determine albedo and insulation
2. Surface properties (snow/ice/meltponds) as they determine the albedo
3. Ice thicknesses (especially thin ice) as they determine albedo, insulation and brine rejection

Although ice velocities in themselves may not be so important, the interaction between dynamics and thermodynamics means that it is also important to simulate ice velocities well and they help to measure the overall sea ice simulation.

This review is laid out as follows; we will first discuss horizontal representation, dynamics and thermodynamics in sea ice models. We will then discuss the design of coupled models and the sea ice interfaces. Sea ice simulations from IPCC models will be discussed before concluding with some thoughts on future model development and evaluation.

2. Horizontal representation

Early sea ice models (Semtner, 1976) defined the surface as being in two categories; sea ice or open water (figure 1). More recently, this has been modified to include multiple categories of sea ice (Bitz et al., 2001) (figure 1).

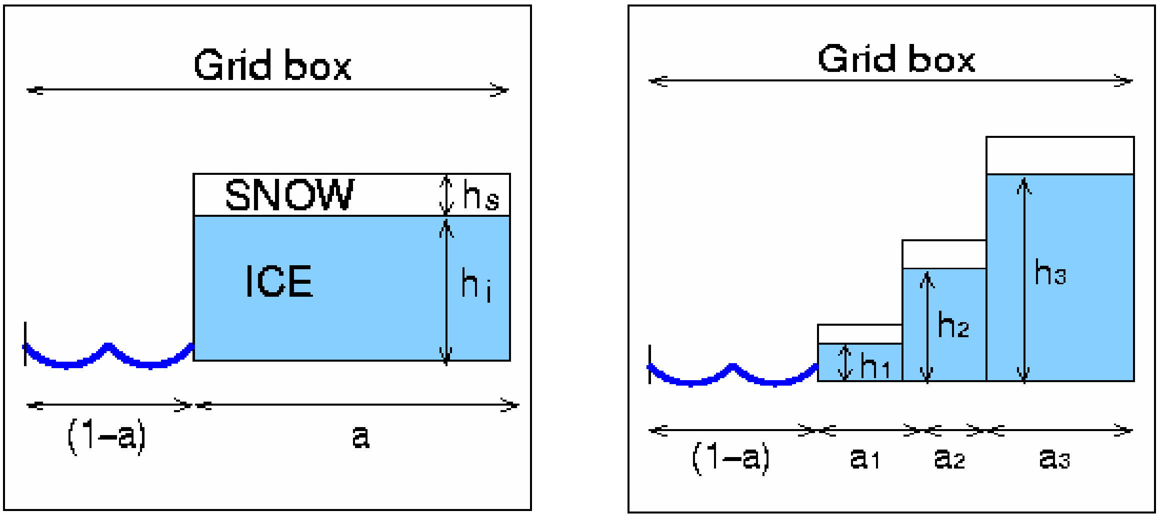

Figure 1 Horizontal representations: simple and sub-gridscale ice thickness distribution.

Ice thickness within the ice pack is variable (figure 2). Many sea ice properties are very dependent on its thickness; for example, growth rate, compressive strength. Properties of ice pack are especially sensitive to amount of thin ice. This suggests that it is important to accurately model the ice thickness distribution (ITD).

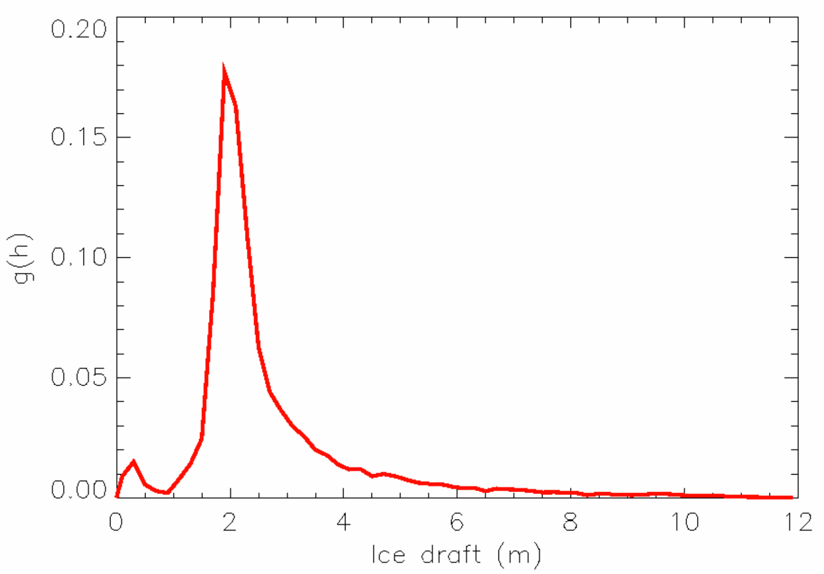

Figure 2: Ice thickness distribution from a submarine track in the central Arctic in September 1992.

The ice thickness distribution is represented by $g(x,h,t)$. $g(x,h,t)$dh is the area covered by ice in range $(h,h+dh)$. The equation for the ITD is given by:

$$\frac{\partial g}{\partial t} = -\nabla \cdot \left(g\,\mathbf{u} \right) - \frac{\partial}{\partial h}\left(fg \right) + \psi$$

where the first term on the right-hand side represents horizontal transport in (x,y) space, the second term represents transport in thickness space due to thermodynamics and the third term represents transport in thickness spec due to ridging.

Bitz et al. (2001) conclude that five thickness categories are sufficient as long as the thin ice is adequately represented within those categories. In thickness categories, the thin ice grows first in Autumn with the volume transferred to thicker classes as ice grows. The thick ice categories generally show only a small seasonal cycle.

Haapala (2000) developed a scheme for ice categories based on ice type (lead, level, rafted, rubble and ridged ice) rather than thickness categories. This type of scheme has the advantage that it can be compared directly with satellite images of sea ice. This scheme has not (at least to the authors' knowledge) been used in a climate model and would be difficult to evaluate against climatology.

3. Dynamics

The governing equation for ice dynamics is given by:

$$\rho h \frac{\partial \mathbf{u}_i}{\partial t} = \rho h f\,\mathbf{k}\,x\,\mathbf{u}_i + a_i\left(\tau_a - \tau_w\right) + \nabla\cdot\sigma - \rho h g \nabla\,\mathbf{H}$$

where the terms on the right-hand side represent Coriolis, wind stress, ocean drag, internal ice stress and surface tilt.

Acceleration is negligible on timescales of a few days. If ice rheology is also assumed to be unimportant, this is known as freedrift. Freedrift is a reasonable approximation for Antarctic sea ice. If ice is thin (i.e., h is small) there is an approximate balance between wind stress and ocean drag which implies that the ice velocity is approximately 2% of the wind magnitude.

The ice rheology is the parameterisation of the internal ice stress term. Ice is generally assumed to diverge freely but resist compression. The details of the relationship between stress and strain are defined by the rheology. Figure 3 shows a number of different rheologies which can be described:

- *Cavitating fluid*: this rheology has incompressible, pure shear behaviour. Convergence and divergence occurs only at endpoints and there is no shear stress.
- *Viscous-plastic*: this rheology has plastic flow on yield curve. Inside the ellipse, viscous behaviour is allowed (with strain rate proportional to stress)
- *Elastic-viscous-plastic*: this rheology is like viscous-plastic but with elastic behaviour added (where strain is proportional to stress) for numerical efficiency
- *Mohr-Coulomb*: this rheology also includes shear stress

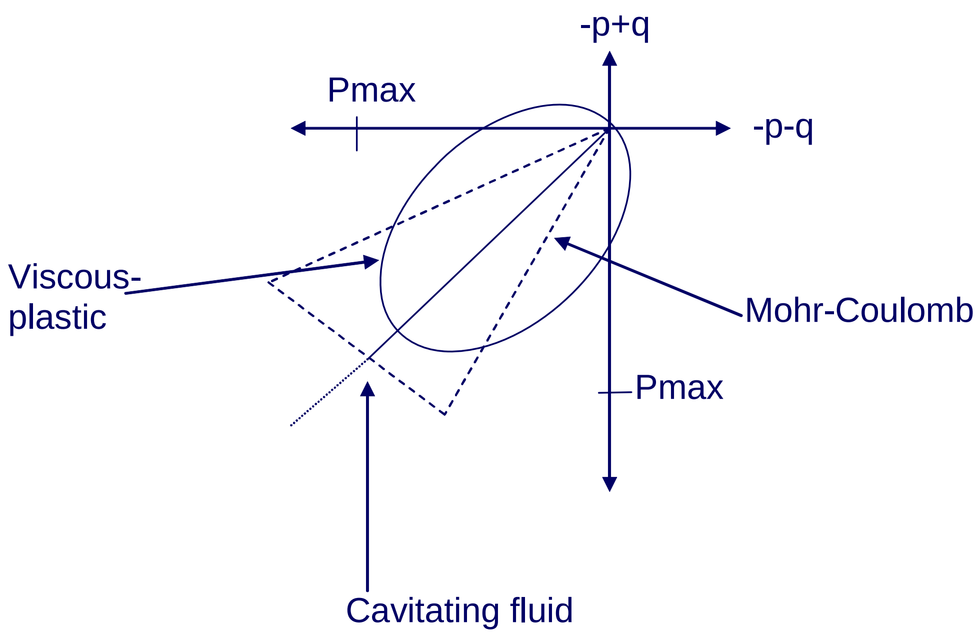

Figure 3: Representation of yield curves from rheologies in principle stress space.

Pmax can be calculated from the Hibler (1979) parameterisation which uses a p* value. Alternatively, it can be related to the ITD and ridging according to Rothrock (1975).

From Flato and Hibler (1992) and experiments at the Met Office, we can conclude that the impact of including a rheology on the Arctic ice velocity field is to strengthen the Beaufort gyre and the export of ice through Fram Strait.

When convergence or shear is present, ridging can occur. The net effect of ridging is to reduce the ice area by transferring thin ice into thicker ice and open water.

4. Thermodynamics

The total albedo is defined as the ratio of the integrated outgoing short wave (250 - 2500 nm) radiation and the integrated incoming short wave radiation. The albedo of bare dry sea ice is greatest in the visible spectrum at 500 nm and least in the near IR.

The albedo of polar regions varies from that of open water (0.05) to that of new snow (0.90), and encompasses the entire range of albedos found on the surface of the planet. It is important in the representation of the ice albedo feedback in the climate system.

As ice thickens the formation of brine drainage channels and trapped brine inclusions result in increased internal scattering and higher albedo. As the ice becomes stressed, ridged and rafted cracks form which increase albedo to 0.6 – 0.7.

Fallen snow grain size and structure changes with temperature (Antarctic sea ice always snow covered). Rainfall and fresh snow fall can significantly alter albedo on short time scales. Effect on albedo of snow is logarithmically dependant on the snow depth, representing drifting and partial coverage of the ice.

Ponds develop during continued melting once ice is snow-free. Ponds grow laterally and in depth draining through the brine channels as quickly as they melt. Albedo of the ponds is ~0.1-0.2 depending on depth and impurities (e.g. soot) which settle on the bottom of the pond. Ponds coverage of the ice surface grows to

~40% significantly reducing the structural strength of the ice. Ponds freeze-up in autumn, and, followed by fresh snowfall the albedo returns to the winter value.

Figure 4 shows the annual cycle of albedo. Models simulate the annual cycle by breaking the season into characteristic periods based on surface air temperature (dotted line). The exact representation of a single season of point observations is not possible with a climate model. Indeed, different parameterisations, as a function of air temperatures, of the summer melt cycle in models can result in significantly different sea ice characteristics. Too much summer melt can result in thinner winter Arctic ice cover and consequently increased heat fluxes to the atmosphere.

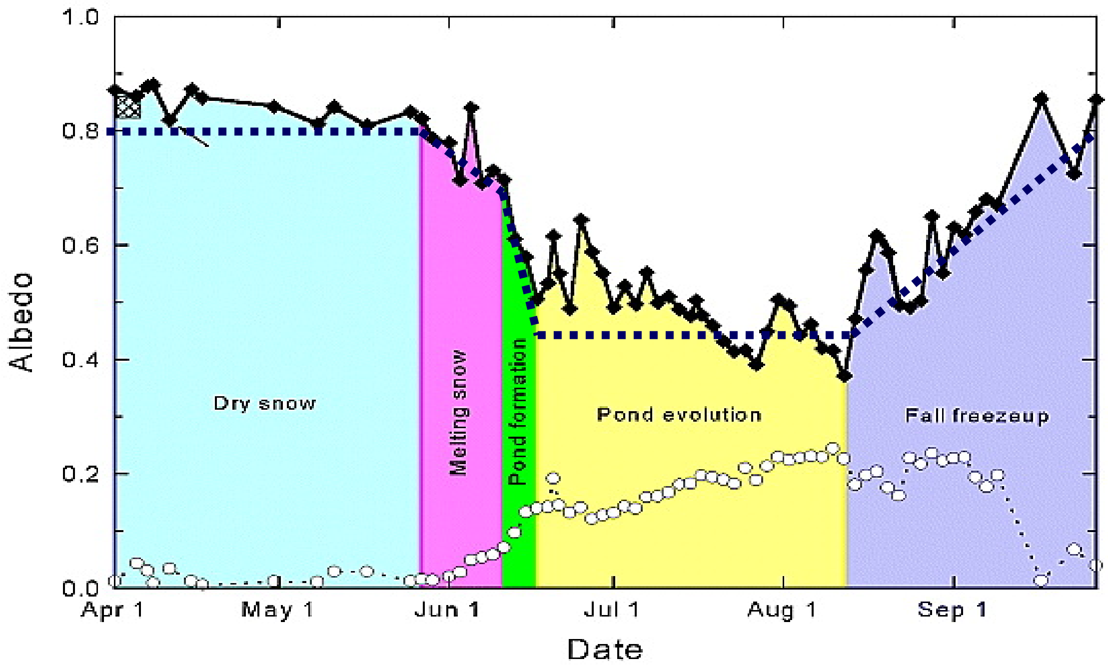

Figure 4: Observations of mean albedo along a 100m line on multiyear ice at the SHEBA field site (from Perovich et al., 2002) throughout the summer season. Open symbols are the spatial standard deviation of albedo.

The evolution of the vertical temperature profile of ice is given by:

$$\rho c \frac{\partial T}{\partial t} = k \frac{\partial T^2}{\partial z^2} + \kappa I_0 e^{-\kappa z}$$

where the second term on the right-hand side represents penetrating radiation. Radiation penetrating ice does not immediately melt ice but warms the interior and melts patches of high salinity ice. Liquid trapped within ice is known as “brine pockets”. Brine pockets are important because they store heat; brine pockets will refreeze with onset of cooling and ice is subsequently thicker at start of autumn than if brine pockets were not present. The penetrating radiation can be stored (Semtner, 1976) which is equivalent to heat being released from refreezing of brine pockets.

Basal fluxes are very difficult to measure. Ocean-ice heat flux can be represented (McPhee, 1992) as a function of the temperature difference between ocean and ice and the ocean drag:

$$F_b = \rho c_p c_h u_* \left(T_{ml} - T_f \right)$$

where $u_* = \sqrt{\tau_w / \rho}$

Ocean-ice heat fluxes need special treatment in the marginal ice zone where ice is thin.

Ablation and accretion (ie, change in thickness) is dealt with by the balance with the surface flux $F(T)$:

$$F(T) = q(S,T)\frac{dh}{dt}$$

where $q(S,T)$ is the energy of melting.

If snow is below freeboard, snow is converted to ice until the base of the snow is at freeboard. This is known as white ice formation.

5. Coupled model design

Coupled models vary on where the sea ice interface is placed. For example, the CSIRO model has sea ice in atmosphere component, the IPSL coupled model has sea ice in ocean component and the Hadley Centre coupled model has sea ice split between ocean and atmosphere. The decisions on where to place sea ice were made for good scientific reasons, which suggests that there probably is no ‘perfect solution’!

There are a number of considerations when deciding where sea ice should be placed:

a) Boundary condition at bottom of atmosphere; Dirichlet or Neumann?

At the top of the sea ice:

Fs (~To-Ts1)=Fa=Rnet-SH-LH

where Ts1 is the first level temperature within the sea ice.

If the boundary conditions are Dirichlet, the atmospheric temperature profile is calculated with To fixed from sea ice and Fa is therefore known. If the boundary conditions are Neumann, Fs is required to solve the atmospheric temperature profile (including To) and Ts1 needs to be passed from the sea ice component to the atmosphere.

b) Capturing the diurnal cycle

To capture diurnal variations in temperature over sea ice (figure 5), frequent coupling is required between Fa, Fs, To and Ts1.

c) Computational limits of coupling frequency between component models

Passing data between model components can increase the cost of the coupled model. For example, going from daily to 3 hourly coupling in HadGEM1 increases the cost of the model by 10%. Therefore we need to design the model to minimise the computational cost (eg, if sea ice is in ocean but needs to couple with atmosphere every timestep this would be an unreasonable overhead).

d) Resolving ocean-ice processes

The ocean grid is generally finer than atmospheric grid. For example, in HadCM3, the ocean is 6 times atmosphere resolution and in HadGEM1, the ocean is 2.3 times atmosphere resolution. To resolve ocean-ice processes need to calculate sea ice properties (ice fraction, thickness, motion) on the ocean grid.

In conclusion, the most appropriate choice to meet these considerations, is to place the sea ice component within the ocean but with frequent coupling to the atmosphere.

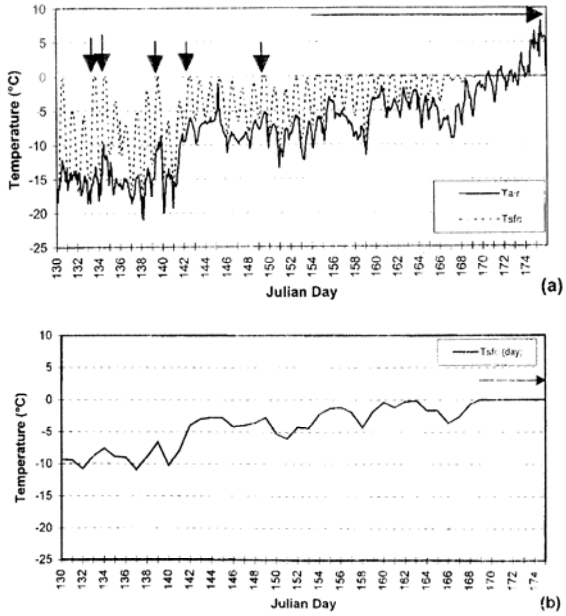

Figure 5: Time series of Arctic spring (a) hourly forced modeled surface temperature (dotted line) and ambient air temperature (solid line) and (b) daily modeled surface temperature. Arrows indicate days with surface melt (from Hanesiak et al., 1999). Inclusion of diurnal cycle increases open water duration by 21 days

6. Results from IPCC models

Data from 23 climate models has been submitted to IPCC AR4. Information is available on 20 sea ice components. Of those 20 components there are:

- 11 multilayer models, 9 single layer models
- 7 multicategory models, 13 single category
- 2 no dynamics, 1 simple dynamics, 17 rheology (viscous-plastic or elastic-viscous-plastic)

Results from sea ice simulations with coupled models are sensitive to the atmosphere and ocean forcings. For example, Arctic sea ice is highly sensitive to the details of the thermohaline circulation in the climate model.

Figure 6 shows that while the ice extents in individual models varies dramatically between the different models, the ensemble average agrees remarkably well with the observed extents in both hemispheres.

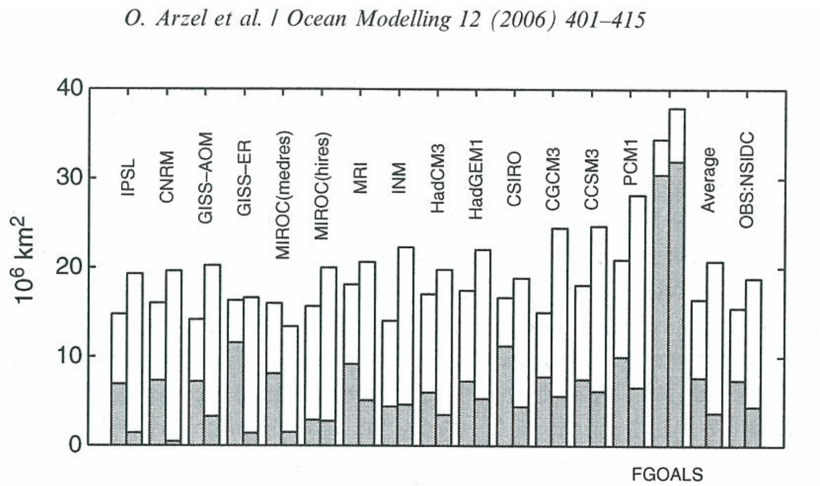

Figure 6: Arctic and Antarctic sea ice extents from Arzel et al. (2006)

7. Conclusions

State of the art in sea ice modelling for climate is:

- multilayers
- multicategories
- sophisticated ice rheology

The emerging choice for the position of the sea ice model is to place sea ice in the ocean component but tightly coupled to the atmospheric boundary layer.

Future model developments are most likely to be focussed on improving albedo and ocean-ice fluxes. To improve albedo the following are suggested:

- Antarctic albedo changes are dependant on the metamorphosis of snow, hence we need to include a good snow model (and the same snow model as used over land!)
- Arctic albedo changes are dependant on the temporal evolution of meltponds, hence we need an explicit model of the ponds to replace simplistic empirical parameterisations

To improve ocean-ice fluxes, the following are suggested:

- Improved parameterisations of ocean-ice heat fluxes in the marginal ice zone
- Improved momentum coupling to ocean, since modelled ice velocities are too high and this is believed to be related to ocean drag

Future evaluation of model simulations suggests the following requirements:

- For the mean state, models should be evaluated against an increasing range of observations; concentration, thickness, velocities, ice types
- To evaluate variability of sea ice on all timescales requires ongoing semi-operational observations
- To enhance confidence in model results, more emphasis should be placed on the ensemble means which should reduce systematic ocean and atmosphere biases and eliminate some ‘tuning’ choices’

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