Energy Loan Sea Ice Model

By Rainer Bleck, 6 September 2000

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Overview

An energy loan sea ice model was developed to manage the energetics of water phase changes in a consistent yet simple manner. The model, which has much in common with the one developed by Semtner (1976, Appendix) focuses on two aspects of the influence of sea ice: (1) the stabilization of ocean temperature near the freezing point through ice formation and melting, and (2) the impact of the ice surface on ocean-atmosphere energy fluxes.

Concerning the stabilization of ocean temperature, the energy loan concept of the ice model ensures that the oceanic mixed layer temperature does not drop below the freezing point (-1.8 $^\circ$C) when the surface heat flux removes heat from the ocean. At each model grid point, the ocean borrows energy from an "energy bank" to stabilize temperature at the freezing point. The energy required to maintain this temperature comes from freezing an appropriate amount of seawater. Conversely, if the surface heat flux adds heat to the ocean, the energy loan must be repaid before the ocean temperature in a grid box is permitted to rise above freezing.

The influence of ice on surface fluxes is large, both by virtue of its high albedo compared to water and because an ice surface can be much colder than open water. In the present ice model, surface temperature is calculated based on the assumption that the system is energetically in a steady state; i.e., the heat flux through the ice matches the atmospheric heat flux.

Specifics

To illustrate this approach, the atmospheric heat flux is written as $F_{air} = a(T_i - T_a)$, and the heat flux through the ice as $F_{ice} = a(T_w - T_i)$, where T_i, T_a, and T_w represents ice, air, and water temperature while a and b are proportionality factors. Given T_a, T_w, and a first guess of T_i (the unknown in this problem), T_i is modified by an amount ΔT_i to minimize the difference between F_{air} and F_{ice}:

$$a(T_i + \Delta T_i - T_a) = b(T_w - T_i - \Delta T_i),$$

which yields

$$\Delta T_i = \frac{aT_a + bT_w}{a+b} - T_i. \tag{1}$$

To make this formula applicable in situations where F_{air} is a mixture of sensible, latent, and radiative heat fluxes, the expressions $aT_i - aT_i$ and $bT_i - bT_i$ are added to the numerator of (1), then the original definitions of F_{air} and F_{ice} are substituted:

$$\Delta T_i = \frac{F_{ice} - F_{air}}{a+b}. \tag{2}$$

The new temperature obviously must not be allowed to exceed the freezing point until the ice has melted completely.

Practical application of (2) requires knowledge of the coefficients a and b which represent the derivatives dF_{air} / dT_i and dF_{ice} / dT_i, respectively. Guidance on the magnitude of a can be obtained from the conventional heat flux bulk formula. It suggests that $a = c_t r c_p U$ where c_t is a nondimensional transfer coefficient (similar to the drag coefficient), r is the air density, c_p is the specific heat of air at constant pressure, and U is the wind speed. The formula for radiative energy loss, $s T^4$, suggests that the previous estimate for a should be increased by an amount $4s T^3$. A reasonable choice for b is the ratio of ice thermal conductivity to ice thickness, k_{ice} / H_{ice}.

In the interest of computational efficiency in coupled climate models, information exchange with the atmosphere should be minimized. The coefficient a is therefore assumed to be independent of atmospheric state variables. To avoid oscillatory behavior in (2), a should be chosen somewhat larger than "typical" values of $c_t r c_p U + 4s T^3$; in other words, a strategy of prudent under-relaxation of T_i is adopted.

Finally, a statement is needed to relate the rate of energy borrowing or repaying to the "composite" atmospheric heat flux:

$$F = cF_{ice} + (1-c) F_{opw}, \tag{3}$$

where F_{opw} is the thermal energy flux over open water and c is the fractional ice coverage. As defined in (3), F is the energy flux felt by the ocean irrespective of the presence of ice. In other words, we assume that the energy flux between the atmosphere and ice, F_{ice}, equals the energy flux between ice and ocean. This assumption is compatible with the steady state (zero heat flux divergence) made to derive (2).

Concerning implementation of the ice model in HYCOM 1.0/2.0, the surface temperature is represented by the temperature of layer 1 regardless of whether the model is run with hybrid vertical coordinates or in MICOM mode. When the model is run with hybrid vertical coordinates and a non-slab mixed layer model, no attempt is made to reduce the thickness of layer 1 at a given grid point when ice forms.

Reference

Semtner, A. J., Jr., 1976: A model for the thermodynamic growth of sea ice in numerical simulations of climate. *J. Phys. Oceanogr.*, **6**, 379-389.