Appendix

A. Sea ice model

The thermodynamics of the sea-ice model used in the coupled model are basically similar to those in the model of Mellor and Kantha (1989). The prognostic variables of the present model are sea-ice thickness (h_I) and sea-ice concentration (A). The model consists of the equations for conservation of mass:

$$\partial(Ah_I)/\partial t + \pounds(Ah_I)$$
$$= \rho_O\{A(W_{IO} + W_{AI}) + (1 - A)W_{AO} + W_{FR}\} /\rho_I + \text{Diffusion} \qquad (A1)$$

and concentration:

$$\partial A/\partial t + \pounds(A)$$
$$= \rho_O\{\Phi(1 - A)W_{AO} + \Psi A W_{IO} H(-W_{IO})$$
$$+ (1 - A)W_{FR}\} /(\rho_I h_I) + \text{Diffusion} \qquad (A2)$$

where ρ_O and ρ_I are the density of seawater and sea ice respectively, H is the Heaviside step function, W_{IO} is the freezing or melting rate at the bottom of the sea ice, W_{AO} is the freezing rate in leads (open water), W_{FR} is the rate of sea-ice accretion at the sea surface due to frazil ice formation in the ocean, W_{AI} is the melting rate at the top of the sea ice. Figure A1 shows the location of W_{IO}, W_{AO}, W_{FR} and W_{AI}. Φ and Ψ are the empirical constants and are set at 1.0 and 0.7, respectively. £ is the advection operator, represented by

$$\pounds(\mu) = ma^{-1}\{\partial(u_{eff}\mu)/\partial\lambda + \partial(v_{eff}\mu m^{-1})/\partial\phi\} \qquad (A3)$$

where μ is an arbitrary variable, λ is longitude, ϕ is latitude, m is secϕ, and a is the radius of the earth. The effective velocity (u_{eff}, v_{eff}) for the advection of sea ice is set in the present study at one third the velocity of the uppermost layer of the ocean model. W_{AS} (see Fig. A1) is the rate of snow melt or accumulation due to precipitation. W_{AS} and W_{AI} are estimated in the atmospheric component of the coupled model.

W_{IO} is calculated from the heat and water balance and equilibrium condition for freezing at the bottom of sea ice:

$$F_{TIO} = \rho_O C_{PO} C_{TZ}(T_I - T_{OI}) \qquad (A4)$$

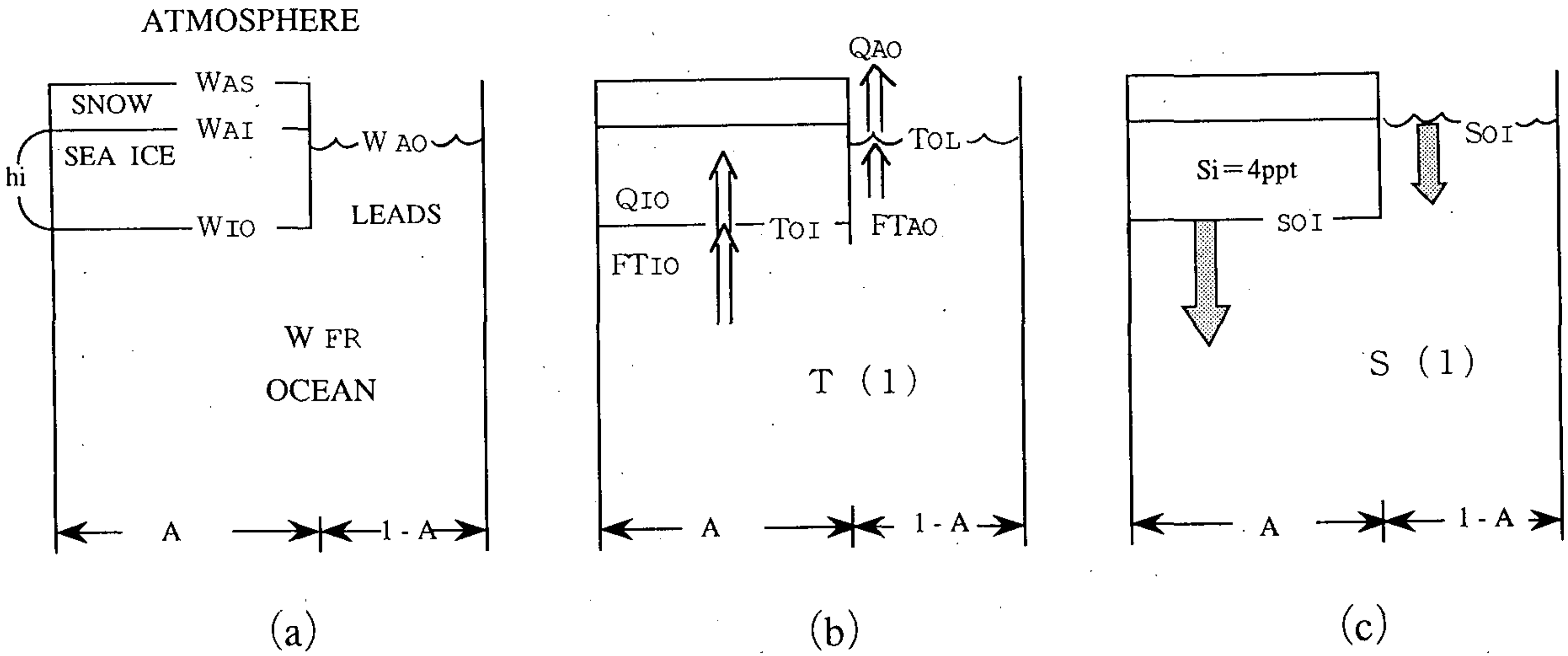

Fig.A1 Schematic diagrams of the sea-ice model. (a) Volume fluxes. (b) Heat fluxes. (c) Freshwater fluxes.

$$T_{OI} = -0.0543S_{OI} \quad \text{(A5)}$$

$$W_{IO} = (Q_{IO} - F_{TIO})/\rho_O L \quad \text{(A6)}$$

$$C_{SZ}(S_1 - S_{OI}) = W_{IO}(S_I - S_1) \quad \text{(A7)}$$

where C_{PO} is the specific heat of seawater, L is the latent heat of fusion, T_{OI} and S_{OI} are the temperature and salinity at the bottom of sea ice, T_1 and S_1 are the temperature and salinity at the uppermost layer of the ocean model, S_I is the salinity of sea ice, F_{TIO} is the heat flux to the bottom of sea ice in the ocean (oceanic heat flux beneath sea ice) and Q_{IO} is conductive heat flux through sea ice (see Fig.A1). The formulations of C_{TZ} and C_{SZ} follow those of Mellor and Kantha (1989) :

$$C_{TZ} = u_{\tau I}/(P_{rt}k^{-1}\ln(-z/z_0) + B_T) \qquad z \to 0 \quad \text{(A8)}$$

$$B_T = b(z_0 u_{\tau I}/\nu)^{1/2}P_r^{2/3} \quad \text{(A9)}$$

$$C_{SZ} = u_{\tau I}/(P_{rt}k^{-1}\ln(-z/z_0) + B_S) \qquad z \to 0 \quad \text{(A10)}$$

$$B_S = b(z_0 u_{\tau I}/\nu)^{1/2}S_c^{2/3} \quad \text{(A11)}$$

where $u_{\tau I}$ is the friction velocity, k is the von Karman's constant, z_0 is the roughness parameter, P_{rt} is the turbulent Prandtl number, b is an empirical factor, ν is the kinematic viscosity of seawater, Pr is the molecular Prandtl number and Sc is the Schmidt number.

W_{AO} is predicted from the heat and water balance and equilibrium condition for freezing at the surface of leads:

$$F_{TAO} = \rho_O C_{PO} C_{TZ}(T_1 - T_{OL}) \quad \text{(A12)}$$

$$T_{OL} = -0.0543S_{OL} \quad \text{(A13)}$$

$$W_{AO} = (Q_{AO} - F_{TAO})/\rho_O L \quad \text{(A14)}$$

$$C_{SZ}(S_1 - S_{OL}) = W_{AO}(S_I - S_1) \quad \text{(A15)}$$

where T_{OL} and S_{OL} are the temperature and salinity at the lead surface, F_{TAO} is the heat flux to the lead surface in the ocean (oceanic heat flux to the lead), and Q_{AO} is the heat flux from the ocean to the atmosphere through leads (see Fig.A1).

The conductive heat flux through sea ice (Q_{IO}) and the heat flux from the ocean to the atmosphere through leads (Q_{AO}), which play a thermal forcing role in the sea-ice model, are calculated in the atmospheric component of the coupled model.

If the sea water becomes supercooled, frazil ice forms and the temperature and salinity of the sea water return to the freezing temperature (T_f) and salinity (S_f), respectively. T_f, S_f and the ratio of mass increment of frazil ice to total seawater mass (γ) are estimated from the heat and water balance and equilibrium condition of freezing:

$$C_{PO}T_S + L = (1 - \gamma)(C_{PO}T_f + L) + \gamma C_{PI}T_f \tag{A16}$$

$$S_S = (1 - \gamma)S_f + \gamma S_I \tag{A17}$$

$$T_f = -0.0543S_f - 0.000759z \tag{A18}$$

where T_S and S_S are the temperature and the salinity of supercooled seawater, respectively, C_{PI} is the specific heat of sea ice and z is the ocean depth. The frazil ice is immediately removed from the ocean and added to the sea ice. The rate of frazil ice accumulation at the sea surface is then

$$W_{FR} = \Delta t^{-1} \int_D^0 \gamma \, dz \tag{A19}$$

where Δt is the numerical time step of the model and D the depth of the ocean.

B. Ocean model

B.1 Governing equations

The ocean model is a world ocean general circulation model developed at the Meteorological Research Institute. It basically follows the ocean general circulation model of Bryan(1969) and includes the Mellor and Yamada(1974) level 2 turbulence closure scheme.

The physical condition of the ocean is specified by seven variables, velocity(u, v and w), pressure(P), in-situ density(ρ), potential temperature(T), and salinity(S). The model employs the Boussinesq and rigid-lid approximations and hydrostatic assumption. A spherical coordinate system is used, with λ, ϕ and z representing longitude, latitude and height. Let m=secϕ,

$n = \sin\phi$ and a be the radius of the Earth. Then the equations of motion and the equations for conservation of heat and salt are:

$$\partial u/\partial t + £(u) - (mna^{-1}u + 2\Omega n)v = -ma^{-1}\rho_o^{-1}\partial P/\partial\lambda + \partial(K_M\partial u/\partial z)/\partial z + A_{MH}\nabla^2 u, \quad \text{(B1)}$$

$$\partial v/\partial t + £(v) + (mna^{-1}u + 2\Omega n)u = -a^{-1}\rho_o^{-1}\partial P/\partial\phi + \partial(K_M\partial v/\partial z)/\partial z + A_{MH}\nabla^2 v, \quad \text{(B2)}$$

$$\partial T/\partial t + £(T) = \partial(K_H\partial T/\partial z)/\partial z + \rho_o^{-1}C_p^{-1}\partial I/\partial z + A_{HH}\nabla^2 T + CA(T), \quad \text{(B3)}$$

$$\partial S/\partial t + £(S) = \partial(K_H\partial S/\partial z)/\partial z + A_{HH}\nabla^2 S + CA(S), \quad \text{(B4)}$$

where £ is the advection operator represented by:

$$£(\mu) = ma^{-1}[\partial(u\mu)/\partial\lambda + \partial(v\mu m^{-1})/\partial\phi] + \partial(w\mu)/\partial z, \quad \text{(B5)}$$

and ∇^2 is the horizontal Laplacian operator defined by:

$$\nabla^2\mu = ma^{-2}[m\partial^2\mu/\partial\lambda^2 + \partial(m^{-1}\partial\mu/\partial\phi)/\partial\phi], \quad \text{(B6)}$$

where μ is an arbitrary variable. The continuity equation is:

$$£(1) = 0. \quad \text{(B7)}$$

The hydrostatic equation and the equation of state are:

$$\partial P/\partial z = -\rho g, \quad \text{(B8)}$$

$$\rho = \rho(T, S, P). \quad \text{(B9)}$$

Here t is time, ρ_o is a reference value of density, C_p is specific heat, I is the downward solar irradiance, g is the acceleration of gravity and Ω is the angular speed of the Earth. The term CA in Eqs.(B3) and (B4) represents convective adjustment which mixes statically unstable water to neutral stratification. The equation of state (B9) is one for in-situ density from potential temperature, salinity and pressure. The formulae from UNESCO(1981) and the Fofonoff's (1977) procedure of calculating in-situ temperature from potential temperature are used to define the equation of state.

The lateral eddy viscosity (A_{MH}) and diffusivity (A_{HH}) are set to be 2.0×10^9 cm^2s^{-1} and 5.0×10^7 cm^2s^{-1}, respectively, between 78°S and 78°N.

Smaller values are used for latitudes higher than 78 degrees, corresponding to the zonal distance between grids which becomes smaller toward the poles.

The vertical eddy viscosity (K_M) and diffusivity (K_H) are estimated from a turbulence closure scheme following Mellor and Yamada (1974, 1982) and Mellor and Durbin (1975):

$$K_M = lqS_M \tag{B10}$$

$$K_H = lqS_H \tag{B11}$$

where l is the turbulent length scale, q is the square root of twice the turbulent kinetic energy and S_M and S_H are stability functions which depend on the flux Richardson number. The flux Richardson number can be calculated from the gradient Richardson number, R_i:

$$R_i = g(\partial\rho/\partial z)/[(\partial u/\partial z)^2 + (\partial v/\partial z)^2] \tag{B12}$$

The quantity q is determined from a quasi-equilibrium form of the turbulent kinetic energy equation in which shear production, buoyancy production and dissipation are locally balanced:

$$lqS_M[(\partial u/\partial z)^2 + (\partial v/\partial z)^2] - lqS_H(g\partial\rho/\partial z) - q^3/(B_1 l) = 0 \tag{B13}$$

where B_1 is an empirical constant. Following Mellor and Durbin(1975), the turbulent length scale, l, is estimated from the ratio of the first to the zero-th moment of the turbulence field:

$$l = \alpha\int_D^0 |z| q\,dz / \int_D^0 q\,dz \tag{B14}$$

where α is an empirical constant. The equations (B10) - (B14) close the turbulence parameterization. The minimum values of K_M and K_H (1.0 cm^2s^{-1} and 0.5 cm^2s^{-1}, respectively) are introduced in the model to guarantee subgrid-scale mixing.

B.2 Boundary conditions

The surface boundary conditions for the ocean with sea ice are:

$$\rho_o K_M(\partial u/\partial z, \partial v/\partial z) = A(\tau_I^\lambda, t_I^\phi) + (1 - A)(\tau_L^\lambda, \tau_L^\phi) \tag{B15}$$

$$\rho_o C_{Po} K_H \partial T/\partial z = -A\, F_{TIO} - (1 - A)\, F_{TAO} \tag{B16}$$

$$K_H\partial S/\partial z = A\{(W_{IO} + W_{AI})(S_1 - S_I) - W_{AS} S_1\}$$
$$+ (1 - A)\{W_{AO}(S_1 - S_I) - (P - E) S_1\} - W_R S_1 \qquad (B17)$$

where (τ_I^λ, t_I^ϕ) is the wind stress on the snow/sea-ice surface, (τ_L^λ, τ_L^ϕ) is the wind stress on the lead surface, P is precipitation, E is evaporation and W_R is the runoff from the land to the ocean.

The surface boundary conditions for the ocean without sea ice are:

$$\rho_O K_M(\partial u/\partial z, \partial v/\partial z) = (\tau^\lambda, \tau^\phi) \qquad (B18)$$

$$\rho_O C_{PO} K_H \partial T/\partial z = Q - Q_S \qquad (B19)$$

$$K_H\partial S/\partial z = (E - P - W_R) S_1 \qquad (B20)$$

$$I = Q_S \qquad (B21)$$

where (τ^λ, τ^ϕ) is the wind stress, Q is the net surface heat flux and Q_S is the solar irradiance at the surface.

The solar penetration is parameterized by:

$$I(z) = Q_S \exp(z/\zeta) \qquad (B22)$$

where ζ is the attenuation length. The divergence of downward irradiance is written as:

$$\partial I/\partial z, \qquad (B23)$$

which is included as a source term in Eq.(B3).

At the bottom, the vertical fluxes of heat and salt are taken to be zero:

$$K_H\partial T/\partial z = 0 \qquad (B24)$$

$$K_H\partial S/\partial z = 0 \qquad (B25)$$

also,

$$q^2 = 0\,. \qquad (B26)$$

The bottom friction is formulated based on the Ekman boundary layer theory:

$$K_M\partial u/\partial z = (\Omega A_{MVb}|\sin\phi|)^{1/2}(u+v) \qquad (B27)$$

$$K_M\partial v/\partial z = (\Omega A_{MVb}|\sin\phi|)^{1/2}(v\pm u)\,. \qquad (B28)$$

The upper and lower signs in Eqs.(B27) and (B28) refer to the northern and southern hemisphere, respectively. A_{MVb} is a coefficient of the vertical eddy viscosity in the bottom Ekman layer. The value of A_{MVb} is set at 1.0×10^2 $cm^2 s^{-1}$.

At lateral walls, a no-slip condition is imposed and no flux of heat or salt is allowed:

$$u = v = \partial T/\partial n = \partial S/\partial n = 0 \qquad \text{(B29)}$$

where $\partial(\ )/\partial n$ represents the local derivative normal to the wall.