

## Description of Sea-Ice Component of Coupled Ocean–Sea-Ice Model for the Earth Simulator (OIFES)

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(Received May 31, 2005; Revised manuscript accepted August 17, 2005)

**Abstract** In this report, sea-ice component of OIFES, Coupled Ocean–Sea-Ice Model for the Earth Simulator, is briefly described. It employs elastic–viscous–plastic rheology for dynamics considering massively parallel computation on the Earth Simulator, and simple zero-layer model for thermodynamics. Some early results are also presented.

**Keywords:** sea-ice model, elastic–viscous–plastic rheology, parallel computation

### 1. Introduction

Coupled Ocean–Sea-Ice Model for the Earth Simulator, named OIFES, is an extended version of OFES, Ocean General Circulation Model (GCM) for the Earth Simulator [1]. OFES is developed mainly to study the ocean surface circulations in tropical and subtropical regions and therefore the computational domain used by Masumoto et al. [1] excludes polar regions. OIFES contains a sea-ice model to be used for global ocean simulations including the Arctic and Antarctic regions and for coupled atmosphere–ocean simulations as an oceanic part of CFES, Coupled GCM for the Earth Simulator.

The sea-ice component of OIFES is implemented as a set of subroutines instead of an independent program from an ocean model. The ocean component of OIFES is almost identical to OFES, which is based on Modular Ocean Model version 3 (MOM3) [2] developed at Geophysical Fluid Dynamics Laboratory (GFDL). Therefore, the ocean component of OIFES is not touched on in this report.

The sea-ice model is based on the model developed at International Arctic Research Center (IARC) of University of Alaska [3], which is coupled with GFDL MOM2 [4] and employs dynamics with viscous–plastic (VP) rheology of Hibler [5] and zero-layer thermodynamics of Parkinson and Washington [6] with snow effects of

Oberhuber et al. [7]. Their coupled ocean–sea-ice model is used for regional study of the Arctic Ocean.

However, momentum equation of sea-ice with VP rheology is not suitable for massively parallel computation on the Earth Simulator, because it is usually solved implicitly and needs many iterations to get the solution of Poisson’s equation. Therefore, we have extended their sea-ice dynamics from VP rheology to elastic–viscous–plastic (EVP) rheology adopted from Hunke and Dukowicz [8], in which momentum equation of sea-ice is solved explicitly. We also modified the coordinate system from Cartesian coordinates to spherical coordinates following Hunke and Dukowicz [9] to cover the global ocean. In order to respond to the improvement in surface treatment of the ocean component from rigid-lid approximation (MOM2) to free-surface (MOM3), fresh water flux is adopted for boundary condition between sea-ice and ocean instead of salinity flux. Additionally, some modifications are applied to sea-ice thermodynamics for coupling with AFES, Atmospheric GCM for the Earth Simulator [10], and the source code is entirely parallelized using MPI (message passing interface). Incidentally, one may find the word “IARC” in some compile options described below, but note that many of them (such as `iarc_ice_evp`) are implemented within the Earth Simulator Center.

In this report, subscripts  $a$ ,  $i$ ,  $s$ , and  $w$  denote air

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(atmosphere), sea-ice, snow, and seawater (ocean), respectively, and superscript *eff* means an “effective” variable, which is multiplied by sea-ice concentration or, in other words, averaged within a model grid. Subroutine names are written in *italic* and option names are expressed in sans serif.

## 2. Fundamentals

### 2.1 Model Variables, Constants, and Namelist Parameters

Sea-ice is assumed to exist over sea surface with the thickness of  $h_i$  and the concentration of  $A_i$ , and to move with the velocity  $\mathbf{u}_i = (u_i, v_i)$ . In addition, snow lies on sea-ice with the depth of  $h_s$ . Figure 1 shows a schematic of sea-ice model. Actually, effective sea-ice thickness  $h_i^{eff} = A_i h_i$  and effective snow depth  $h_s^{eff} = A_i h_s$  are prognostic variables instead of  $h_i$  and  $h_s$ , respectively. We call  $A_i$ ,  $h_i^{eff}$ , and  $h_s^{eff}$  “tracers” in this report for convenience.

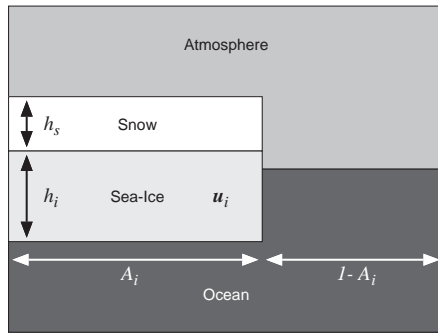


Fig. 1 Schematics of sea-ice model.

Table 1 Prognostic variables in OIFES.

Symbol	Description	Unit
$A_i$	sea-ice concentration	[-]
$h_i^{eff}$	effective sea-ice thickness	[m]
$h_s^{eff}$	effective snow depth	[m]
$u_i$	zonal component of sea-ice velocity	[m s <sup>-1</sup> ]
$v_i$	meridional component of sea-ice velocity	[m s <sup>-1</sup> ]

Model variables, constants, and namelist parameters are summarized in Tables 1, 2, and 3, respectively.

### 2.2 Coordinate System

The sea-ice component of OIFES employs a 2-dimensional spherical coordinate system, so the position is

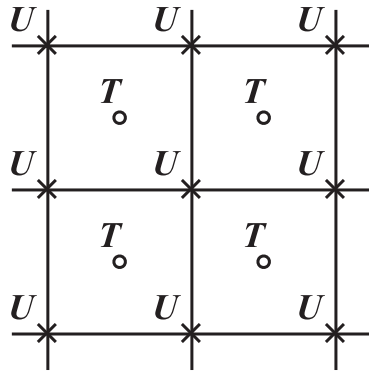
Table 2 Constants used in OIFES.

Symbol	Description	Value
$a$	radius of the earth	$6,371 \times 10^3$ [m]
$c_{pa}$	specific heat of air	1005.0 [J kg <sup>-1</sup> K <sup>-1</sup> ]
$c_{pw}$	specific heat of seawater	4218.0 [J kg <sup>-1</sup> K <sup>-1</sup> ]
$g$	acceleration due to gravity	9.806 [m s <sup>-2</sup> ]
$k_i$	thermal conductivity of ice	2.0344 [W m <sup>-1</sup> K <sup>-1</sup> ]
$k_s$	thermal conductivity of snow	0.3098 [W m <sup>-1</sup> K <sup>-1</sup> ]
$L_{melt}$	latent heat of fusion	$3.4 \times 10^5$ [J kg <sup>-1</sup> ]
$L_{vap}$	latent heat of vaporization	$2.5 \times 10^6$ [J kg <sup>-1</sup> ]
$T_0$	freezing point of fresh water	273.15 [K]
$T_w^{frz}$	freezing point of seawater *	271.20 [K]
$T_i^{melt}$	melting point of sea-ice	273.05 [K]
$T_s^{melt}$	melting point of snow	273.15 [K]
$\alpha_i^{dry}$	albedo of dry sea-ice	0.60
$\alpha_i^{wet}$	albedo of wet sea-ice	0.50
$\alpha_s^{dry}$	albedo of dry snow	0.84
$\alpha_s^{wet}$	albedo of wet snow	0.70
$\alpha_w$	albedo of seawater	0.10
$\gamma$	adiabatic lapse rate	0.01 [K m <sup>-1</sup> ]
$\epsilon_i$	sea-ice emissivity	0.97
$\epsilon_s$	snow emissivity	0.99
$\epsilon_w$	seawater emissivity	0.97
$\kappa$	von Kármán's constant	0.4
$\rho_a$	air density	1.2 [kg m <sup>-3</sup> ]
$\rho_f$	fresh water density	1000 [kg m <sup>-3</sup> ]
$\rho_i$	sea-ice density	920 [kg m <sup>-3</sup> ]
$\rho_s$	snow density	330 [kg m <sup>-3</sup> ]
$\rho_w$	seawater density	1026 [kg m <sup>-3</sup> ]
$\sigma$	Stefan-Boltzman constant	$5.67 \times 10^{-8}$ [W m <sup>-2</sup> K <sup>-1</sup> ]

\*When the option `variable_ticb` is enabled,  $T_w^{frz}$  becomes a diagnostic variable.

Table 3 Namelist parameters in OIFES. Test values are used for test run described in Section 8.

Symbol	Description	Default Value	Test Value
$A_{i\_min}$	minimum sea-ice concentration	0.15	0.01
$C_t$	bulk transfer coefficient	$9.65 \times 10^{-6}$ [m s <sup>-1</sup> ]	$9.65 \times 10^{-5}$ [m s <sup>-1</sup> ]
$d_1$	diffusivity for harmonic-type operator	0.004 [m s <sup>-1</sup> ]	0.004 [m s <sup>-1</sup> ]
$d_2$	diffusivity for biharmonic-type operator	0.004 [m s <sup>-1</sup> ]	0.004 [m s <sup>-1</sup> ]
$h_{i\_min}^{eff}$	minimum sea-ice thickness	0.01 [m]	0.01 [m]
$N_{evp}$	number of elastic subcycle	72	10
$P$	coefficient for sea-ice pressure	$2.75 \times 10^4$ [Pa]	$10.0 \times 10^4$ [Pa]
$\rho_w C_{Dw}$	seawater density $\times$ bulk drag coefficient	5.5 [kg m <sup>-3</sup> ]	5.5 [kg m <sup>-3</sup> ]



**Fig. 2** Spatial arrangement of variables. The letters T and U refer tracer point and velocity point, respectively.  $A_i$ ,  $h_i^{eff}$ , and  $h_s^{eff}$  are defined at tracer points and  $u_i$  and  $v_i$  are at velocity points.

expressed by longitude  $\lambda$  and latitude  $\phi$ . The Arakawa B-grid is used for spatial discretization, and  $A_i$ ,  $h_i^{eff}$ , and  $h_s^{eff}$  are arranged to so-called tracer points and  $u_i$  and  $v_i$  are to velocity points (Fig. 2). The grid spacing is set to the same value as in the ocean component in subroutine *setice*.

### 2.3 Governing Equations

Momentum equation of sea-ice is expressed as follows<sup>1</sup> [5, Eq. (1)]:

$$m_i^{eff} \frac{\partial \mathbf{u}_i}{\partial t} = -m_i^{eff} f \mathbf{k} \times \mathbf{u}_i + \boldsymbol{\tau}_{ai}^{eff} + \boldsymbol{\tau}_{iw}^{eff} - m_i^{eff} g \nabla h_w + \mathbf{F}, \quad (1)$$

where  $m_i^{eff} = \rho_i h_i^{eff} + \rho_s h_s^{eff}$  is total mass of sea-ice and snow per unit area,  $\rho_i$  sea-ice density,  $\rho_s$  snow density,  $f$  the Coriolis parameter,  $\mathbf{k}$  a unit vector normal to the surface,  $\boldsymbol{\tau}_{ai}^{eff} = A_i \boldsymbol{\tau}_{ai}$  effective wind stress,  $\boldsymbol{\tau}_{iw}^{eff} = A_i \boldsymbol{\tau}_{iw}$  effective ocean stress,  $g$  acceleration due to gravity,  $h_w$  sea surface height, and  $\mathbf{F}$  internal ice force. Note that sea surface height  $h_w$  is now a prognostic variable calculated in the ocean component of OIFES, so  $\nabla h_w$  is evaluated directly from  $h_w$  rather than from the geostrophic ocean current. Detail about internal ice force  $\mathbf{F}$  is described in section 4.1.

For sea-ice concentration  $A_i$ , effective sea-ice thickness  $h_i^{eff}$ , and effective snow depth  $h_s^{eff}$ , tracer equations are given as follows [5, Eqs. (13) and (14)]:

$$\frac{\partial A_i}{\partial t} = -\mathcal{L}(A_i) + \mathcal{D}(A_i) + F_A, \quad (2)$$

$$\frac{\partial h_i^{eff}}{\partial t} = -\mathcal{L}(h_i^{eff}) + \mathcal{D}(h_i^{eff}) + F_i^{eff}, \quad (3)$$

$$\frac{\partial h_s^{eff}}{\partial t} = -\mathcal{L}(h_s^{eff}) + \mathcal{D}(h_s^{eff}) + F_s^{eff}, \quad (4)$$

where  $\mathcal{L}$  and  $\mathcal{D}$  represent advection and diffusion operators, respectively (see section 4.2), and  $F_A$ ,  $F_i^{eff}$ , and  $F_s^{eff}$  are source terms determined from thermodynamic process described in section 5.

## 3. Connection with Other Components

### 3.1 Atmosphere

In order to drive OIFES, wind stress  $\boldsymbol{\tau}_a$ , downward short- and long-wave radiation fluxes,  $R_{SW} \downarrow$  and  $R_{LW} \downarrow$ , sensible and latent heat fluxes,  $H_{sen} \downarrow$  and  $H_{lat} \downarrow$ , rainfall  $W_{rain}$ , snowfall  $W_{snow}$ , river runoff  $W_{river}$ , and sea-ice/snow surface temperature  $T_{sfc}$  are needed. In the case of stand-alone simulation, some of them are calculated in OIFES using surface atmospheric data, while in the case of coupled simulation, all of them are calculated in AFES.

If the option *iarc\_flg* is enabled, forcing data dependent on ocean/sea-ice variables (sea-ice/snow surface temperature  $T_{sfc}$ , sensible heat flux  $H_{sen} \downarrow$ , and latent heat flux  $H_{lat} \downarrow$ ) are calculated within OIFES using bulk formula from surface air temperature  $T_a$ , surface specific humidity  $q_a$ , surface wind speed  $\|\mathbf{V}_a\|$ , and surface air pressure  $p_{sfc}$ . Additionally, rainfall  $W_{rain}$  and snowfall  $W_{snow}$  are parameterized from precipitation  $W_{prcp}$  and surface air temperature  $T_a$ . In detail, see section 5. Other forcing data are provided from some data sets such as NCEP/NCAR reanalysis.

When the option *cfes* is enabled, OIFES is coupled with AFES and all forcing data are provided from the atmospheric component. In this case, following variables are transferred from the ocean/sea-ice component to the atmospheric component in order to calculate surface fluxes: ocean surface temperature  $T_w$ , sea-ice concentration  $A_i$ , effective sea-ice thickness  $h_i^{eff}$ , effective snow depth  $h_s^{eff}$ , and freezing point of seawater  $T_w^{fz}$  (only if the option *variable\_ticb* is enabled).

### 3.2 Ocean

From the ocean component, temperature  $T_w$ , salinity  $S_w$ , and velocity  $\mathbf{u}_w = (u_w, v_w)$  of the top level and sea surface height  $h_w$  are input to the sea-ice component.

If the option *variable\_ticb* is enabled, freezing point of seawater,  $T_w^{fz}$ , is evaluated as a linear function<sup>2</sup> of surface salinity  $S_w$  as

$$T_w^{fz} = T_0 - \mu S_w, \quad (5)$$

where  $T_0$  is freezing point of freshwater and  $\mu = 0.0543$

<sup>1</sup> Advection term is neglected in the current version of OIFES.

<sup>2</sup> This should be extended to a nonlinear function of  $S_w$  in future.

K psu<sup>-1</sup> [11]. Otherwise,  $T_w^{frz}$  is taken as constant value (271.2 K).

After the sea-ice variables are updated, resulting momentum flux  $\boldsymbol{\tau}_{surf}$ , heat flux  $Q_{surf}$ , and fresh water flux  $W_{surf}$  are provided as surface boundary conditions for the ocean component (see section 6).

## 4 Dynamics

Sea-ice dynamics is a two-dimensional (horizontal) process and consists of momentum equation (section 4.1) and tracer advection/diffusion (section 4.2).

### 4.1 Momentum Equation

Again, momentum equation of sea-ice is expressed as follows:

$$m_i^{eff} \frac{\partial u_i}{\partial t} = +m_i^{eff} f v_i + \tau_{ai,\lambda}^{eff} + \tau_{iw,\lambda}^{eff} - m_i^{eff} g \frac{1}{a \cos \phi} \frac{\partial h_w}{\partial \lambda} + F_\lambda. \quad (1a)$$

$$m_i^{eff} \frac{\partial v_i}{\partial t} = -m_i^{eff} f u_i + \tau_{ai,\phi}^{eff} + \tau_{iw,\phi}^{eff} - m_i^{eff} g \frac{1}{a} \frac{\partial h_w}{\partial \phi} + F_\phi. \quad (1b)$$

Given the ocean surface velocity  $\mathbf{u}_w$ , ocean stress  $\boldsymbol{\tau}_{iw}$  is expressed as a function of sea-ice velocity  $\mathbf{u}_i$  as [5, Eq. (2)]:

$$\boldsymbol{\tau}_{iw} = \rho_w C_{Dw} \|\mathbf{u}_w - \mathbf{u}_i\| \left[ (\mathbf{u}_w - \mathbf{u}_i) \cos \theta + \mathbf{k} \times (\mathbf{u}_w - \mathbf{u}_i) \sin \theta \right], \quad (6)$$

where  $\rho_w$  is seawater density,  $C_{Dw}$  the bulk drag coefficient,  $\theta$  water turning angle. Note that previous studies [5, 8] used the geostrophic ocean current  $\mathbf{U}_w$  of the interior region (rather than  $\mathbf{u}_w$ ) and  $\theta$  was set to be 25°, but our OGCM usually has enough vertical resolution of several meters at surface and therefore we set  $\theta$  to be zero. The ocean stress term and the Coriolis term are treated implicitly.

Internal ice force  $\mathbf{F} = (F_\lambda, F_\phi)$  is calculated as the divergence of stress tensor  $\boldsymbol{\sigma}$  [12]:

$$F_\lambda = \frac{1}{a \cos \phi} \frac{\partial \sigma_{\lambda\lambda}}{\partial \lambda} + \frac{1}{a} \frac{\partial \sigma_{\lambda\phi}}{\partial \phi} - \frac{2\sigma_{\lambda\phi} \tan \phi}{a}, \quad (7a)$$

$$F_\phi = \frac{1}{a \cos \phi} \frac{\partial \sigma_{\lambda\phi}}{\partial \lambda} + \frac{1}{a} \frac{\partial \sigma_{\phi\phi}}{\partial \phi} + \frac{(\sigma_{\lambda\lambda} - \sigma_{\phi\phi}) \tan \phi}{a}. \quad (7b)$$

Stress tensor  $\boldsymbol{\sigma}$  is related to strain rate  $\dot{\boldsymbol{\epsilon}}$  and pressure  $P$  via a constitutive law, which depends on the treatment of sea-ice rheology (see subsection 4.1.1 for VP model and 4.1.2 for EVP model).

Strain rate  $\dot{\boldsymbol{\epsilon}}$  is expressed by means of sea-ice velocity  $\mathbf{u}_i$  as follows [12]:

$$\dot{\epsilon}_{\lambda\lambda} = \frac{1}{a \cos \phi} \frac{\partial u_i}{\partial \lambda} - \frac{v_i \tan \phi}{a}, \quad (8a)$$

$$\dot{\epsilon}_{\phi\phi} = \frac{1}{a} \frac{\partial v_i}{\partial \phi}, \quad (8b)$$

$$\dot{\epsilon}_{\lambda\phi} = \frac{1}{2} \left( \frac{1}{a} \frac{\partial u_i}{\partial \phi} + \frac{u_i \tan \phi}{a} + \frac{1}{a \cos \phi} \frac{\partial v_i}{\partial \lambda} \right), \quad (8c)$$

The pressure  $P$ , which is a measure of sea-ice strength, is parameterized as a function of sea-ice thickness and concentration as follows [5, Eq. (17)]:

$$P = P^* h_i^{eff} \exp[-C(1 - A_i)], \quad (9)$$

where  $P^*$  (a namelist parameter) and  $C = 20$  are empirical constants. Nonlinear shear viscosity  $\eta = \eta(\dot{\boldsymbol{\epsilon}}, P)$  and nonlinear bulk viscosity  $\zeta = \zeta(\dot{\boldsymbol{\epsilon}}, P)$ , which are used to calculate stress tensor  $\boldsymbol{\sigma}$ , are parameterized as [5]:

$$\zeta = P/2\Delta, \quad (10)$$

$$\eta = \zeta/e^2, \quad (11)$$

where

$$\Delta = \left[ \left( \dot{\epsilon}_{\lambda\lambda}^2 + \dot{\epsilon}_{\phi\phi}^2 \right) (1 + e^{-2}) + 4e^{-2} \dot{\epsilon}_{\lambda\phi}^2 + 2\dot{\epsilon}_{\lambda\lambda} \dot{\epsilon}_{\phi\phi} (1 - e^{-2}) \right]^{1/2} \quad (12)$$

and  $e$  is the ratio of principal axes of the ellipse and equal to 2. In order to avoid the viscosities become infinite in the limit of zero strain rate, we set upper bounds of  $\zeta$  and  $\eta$  as

$$\zeta_{max} = (2.5 \times 10^8) P, \quad (13)$$

$$\eta_{max} = \zeta_{max}/e^2. \quad (14)$$

We also set lower bound as  $\zeta_{min} = 4.0 \times 10^8$  kg s<sup>-1</sup>, and if  $\zeta_{max} < \zeta_{min}$  then  $\zeta = \zeta_{min}$ .

#### 4.1.1 Viscous–Plastic Model

If the option `iarc_ice_hib` is enabled, VP type constitutive law [5] is adopted.

In this case, stress tensor  $\boldsymbol{\sigma}$  is given by

$$\sigma_{\lambda\lambda} = (\zeta + \eta) \dot{\epsilon}_{\lambda\lambda} + (\zeta - \eta) \dot{\epsilon}_{\phi\phi} - P/2, \quad (15a)$$

$$\sigma_{\phi\phi} = (\zeta - \eta) \dot{\epsilon}_{\lambda\lambda} + (\zeta + \eta) \dot{\epsilon}_{\phi\phi} - P/2, \quad (15b)$$

$$\sigma_{\lambda\phi} = 2\eta \dot{\epsilon}_{\lambda\phi}. \quad (15c)$$

Thus, stress tensor  $\boldsymbol{\sigma}$  is diagnosed from sea-ice velocity  $\mathbf{u}_i$  through strain rate  $\dot{\boldsymbol{\epsilon}}$  using Eqs. (8) and (15), and momentum equation (1) is solved iteratively using the following relationship:

$$\begin{aligned} F_\lambda = & \frac{1}{a \cos \phi} \frac{\partial}{\partial \lambda} \left[ (\zeta + \eta) \left( \frac{1}{a \cos \phi} \frac{\partial u_i}{\partial \lambda} - \frac{v_i \tan \phi}{a} \right) \right. \\ & \left. + (\zeta - \eta) \left( \frac{1}{a} \frac{\partial v_i}{\partial \phi} \right) - \frac{P}{2} \right] \\ & + \frac{1}{a} \frac{\partial}{\partial \phi} \left[ \eta \left( \frac{1}{a} \frac{\partial u_i}{\partial \phi} + \frac{u_i \tan \phi}{a} + \frac{1}{a \cos \phi} \frac{\partial v_i}{\partial \lambda} \right) \right] \\ & - \frac{2 \tan \phi}{a} \left[ \eta \left( \frac{1}{a} \frac{\partial u_i}{\partial \phi} + \frac{u_i \tan \phi}{a} + \frac{1}{a \cos \phi} \frac{\partial v_i}{\partial \lambda} \right) \right], \quad (16a) \end{aligned}$$

$$\begin{aligned} F_\phi = & \frac{1}{a \cos \phi} \frac{\partial}{\partial \lambda} \left[ \eta \left( \frac{1}{a} \frac{\partial u_i}{\partial \phi} + \frac{u_i \tan \phi}{a} \right. \right. \\ & \left. \left. + \frac{1}{a \cos \phi} \frac{\partial v_i}{\partial \lambda} \right) \right] \\ & + \frac{1}{a} \frac{\partial}{\partial \phi} \left[ (\zeta - \eta) \left( \frac{1}{a \cos \phi} \frac{\partial u_i}{\partial \lambda} - \frac{v_i \tan \phi}{a} \right) \right. \\ & \left. + (\zeta + \eta) \left( \frac{1}{a} \frac{\partial v_i}{\partial \phi} \right) - \frac{P}{2} \right] \\ & + \frac{2 \tan \phi}{a} \left[ \eta \left( \frac{1}{a \cos \phi} \frac{\partial u_i}{\partial \lambda} - \frac{v_i \tan \phi}{a} - \frac{1}{a} \frac{\partial v_i}{\partial \phi} \right) \right]. \quad (16b) \end{aligned}$$

#### 4.1.2 Elastic–Viscous–Plastic Model

If the option `iarc_ice_evp` is enabled, EVP type constitutive law [8] is adopted.

Constitutive law of VP model, Eq. (15), can be rewritten in the form:

$$\frac{\zeta + \eta}{4\zeta\eta} \sigma_{\lambda\lambda} - \frac{\zeta - \eta}{4\zeta\eta} \sigma_{\phi\phi} + \frac{P}{4\zeta} = \dot{\epsilon}_{\lambda\lambda}, \quad (17a)$$

$$-\frac{\zeta - \eta}{4\zeta\eta} \sigma_{\lambda\lambda} + \frac{\zeta + \eta}{4\zeta\eta} \sigma_{\phi\phi} + \frac{P}{4\zeta} = \dot{\epsilon}_{\phi\phi}, \quad (17b)$$

$$\frac{1}{2\eta} \sigma_{\lambda\phi} = \dot{\epsilon}_{\lambda\phi}. \quad (17c)$$

By introducing “pseudo” elastic contribution to the con-

stitutive law, we obtain

$$\frac{1}{E} \frac{\partial \sigma_{\lambda\lambda}}{\partial t} + \frac{\zeta + \eta}{4\zeta\eta} \sigma_{\lambda\lambda} - \frac{\zeta - \eta}{4\zeta\eta} \sigma_{\phi\phi} + \frac{P}{4\zeta} = \dot{\epsilon}_{\lambda\lambda}, \quad (18a)$$

$$\frac{1}{E} \frac{\partial \sigma_{\phi\phi}}{\partial t} - \frac{\zeta - \eta}{4\zeta\eta} \sigma_{\lambda\lambda} + \frac{\zeta + \eta}{4\zeta\eta} \sigma_{\phi\phi} + \frac{P}{4\zeta} = \dot{\epsilon}_{\phi\phi}, \quad (18b)$$

$$\frac{1}{E} \frac{\partial \sigma_{\lambda\phi}}{\partial t} + \frac{1}{2\eta} \sigma_{\lambda\phi} = \dot{\epsilon}_{\lambda\phi}, \quad (18c)$$

where  $E$  corresponds to Young’s modulus. The parameter  $E$  is determined to satisfy the CFL condition as:

$$E = \frac{2E_0 m_i^{eff}}{(\Delta t_{evp})^2} \times \min \left[ (a \cos \phi \Delta \lambda)^2, (a \Delta \phi)^2 \right], \quad (19)$$

where  $E_0$  is a constant (0.25). Thus, temporal changes of stress tensor  $\boldsymbol{\sigma}$  and sea-ice velocity  $\mathbf{u}_i$  are solved simultaneously using Eqs. (1), (7), (8), and (18) with an elastic subcycling time step  $\Delta t_{evp} = \Delta t / N_{evp}$ , where  $N_{evp}$  is a number of elastic subcycle (a namelist parameter).

#### 4.2 Tracer Advection and Diffusion

In the subroutine `advdif`, temporal changes of tracers due to advection and diffusion are calculated.

Advection operator  $\mathcal{L}$  is given as follows:

$$\begin{aligned} \mathcal{L}(\alpha) &= \nabla \cdot (\mathbf{u}_i \alpha) \\ &= \frac{1}{a \cos \phi} \frac{\partial}{\partial \lambda} (u_i \alpha) + \frac{1}{a \cos \phi} \frac{\partial}{\partial \phi} (v_i \alpha \cos \phi), \quad (20) \end{aligned}$$

where  $\alpha$  corresponds to  $A_i$ ,  $h_i^{eff}$ , or  $h_s^{eff}$  and  $\mathbf{u}_i$  is sea-ice velocity. Simple second-order finite volume method is implemented as advection scheme.

When the option `iarc_ice_biharmonic` is NOT enabled, harmonic-type horizontal diffusion operator is used:

$$\begin{aligned} \mathcal{D}(\alpha) &= \nabla \cdot (D_1 \nabla \alpha) \\ &= \frac{1}{a \cos \phi} \frac{\partial}{\partial \lambda} \left( \frac{D_1}{a \cos \phi} \frac{\partial \alpha}{\partial \lambda} \right) \\ &+ \frac{1}{a \cos \phi} \frac{\partial}{\partial \phi} \left( \cos \phi \frac{D_1}{a} \frac{\partial \alpha}{\partial \phi} \right), \quad (21) \end{aligned}$$

For default setting, diffusivity  $D_1$  is

$$D_1 = d_1 (a \Delta \lambda),$$

while in the case that the option `iarc_ice_am_cosine` is enabled,

$$D_1 = d_1 (a \Delta \lambda \cos \phi),$$

where  $d_1$  is a namelist parameter.

When the option `iarc_ice_biharmonic` is enabled, har-

monic + biharmonic-type horizontal diffusion operator is used:

$$\mathcal{D}(\alpha) = \nabla \cdot (D_1 \nabla \alpha) - \nabla \cdot (D_2 \nabla^3 \alpha). \quad (22)$$

For default setting, diffusivities  $D_1$  and  $D_2$  are

$$D_1 = d_1 (a\Delta\lambda), \quad D_2 = d_2 (a\Delta\lambda)^3,$$

while in the case that the option `iarc_ice_am_cosine` is enabled,

$$D_1 = d_1 (a\Delta\lambda \cos \phi), \quad D_2 = d_2 (a\Delta\lambda \cos \phi)^3,$$

where  $d_1$  and  $d_2$  are namelist parameters.

### 4.3 Polar Filter

If the options `iarc_ice_filter` and `firfil` are enabled, a finite impulse response filter is applied to sea-ice velocity  $\mathbf{u}_i$  at each elastic subcycle in a similar way to that applied to ocean velocity in order to relax the time step constraint imposed by convergence of meridians. In detail, refer to MOM3 manual [2, Chapter 27]. The same reference latitudes (the filtering is applied poleward of these latitudes) as set for the oceanic variables are used.

## 5 Thermodynamics

Thermodynamic model of sea-ice is based on that proposed by Parkinson and Washington [6], which is a Semtner-type [13] zero-layer model. Open ocean (or lead) and sea-ice region within a model grid are treated separately (sections 5.1 and 5.2, respectively) and resulting growth rates of sea-ice are combined using sea-ice concentration (section 5.3). Figure 3 shows a schematic diagram of our thermodynamic model. Note that sea-ice thermodynamics is a one-dimensional (vertical) process.

### 5.1 Open Ocean

#### 5.1.1 Surface Fluxes

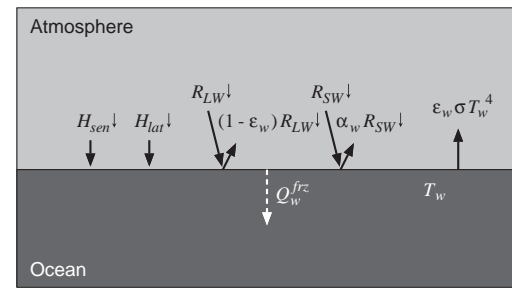
When the option `iarc_flg` is enabled, sensible and latent heat fluxes,  $H_{sen} \downarrow$  and  $H_{lat} \downarrow$ , are evaluated using surface air temperature  $T_a$ , surface specific humidity  $q_a$ , surface wind speed  $\|\mathbf{V}_a\|$ , and surface air pressure  $p_{sfc}$  in subroutine `asflx` following procedures and coefficients of Large and Pond [14, 15].

Using bulk formulae with bulk transfer coefficients,  $C_H$  and  $C_E$ , turbulent heat fluxes are expressed as:

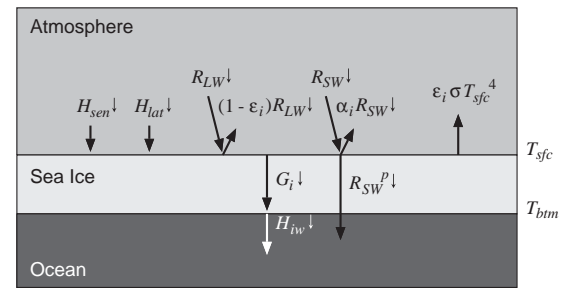
$$H_{sen} \downarrow = \rho_a c_{pa} C_H \|\mathbf{V}_a\| \Delta \theta, \quad (23)$$

$$H_{lat} \downarrow = \rho_a L_{vap} C_E \|\mathbf{V}_a\| \Delta q, \quad (24)$$

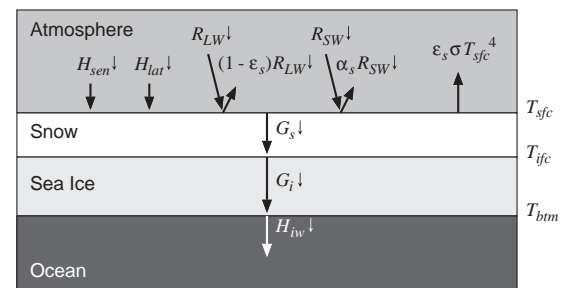
(a) open ocean (lead)



(b) snow-free sea-ice



(c) snow-covered sea-ice



**Fig. 3** Schematic diagrams of the energy fluxes involves for (a) open ocean, (b) snow-free sea-ice, and (c) snow-covered sea-ice. Adapted from Parkinson and Washington [6].

where  $\rho_a$  is air density,  $c_{pa}$  specific heat of air, and  $L_{vap}$  latent heat of vaporization. Potential temperature difference  $\Delta\theta$  and specific humidity difference  $\Delta q$  are given as:

$$\Delta \theta = T_a + \gamma z_T - T_w, \quad (25)$$

$$\Delta q = q_a - q_{sfc}, \quad (26)$$

where  $\gamma$  is adiabatic lapse rate and  $z_T = 2$  m observed height of  $T_a$ . Surface specific humidity  $q_{sfc}$  is calculated as:

$$q_{sfc} = 0.98 \times \frac{0.622 \times e_s}{p_{sfc} - 0.378 \times e_s}, \quad (27)$$

where the factor 0.98 is for seawater correction and  $e_s$  is saturation vapor pressure:

$$e_s = 6.11 \times 10^2 \exp\left(17.269 \times \frac{T_w - T_0}{T_w - T_0 + 237.3}\right). \quad (28)$$

Bulk coefficients,  $C_D = c_D^2$ ,  $C_H = c_H c_D$ , and  $C_E = c_E c_D$ , are estimated iteratively as follows. Turbulent scales of velocity  $u^*$ , temperature  $T^*$ , and humidity  $q^*$  are evaluated by means of  $c_D$ ,  $c_H$ , and  $c_E$  as:

$$u^* = c_D \|V_a\|, \quad (29)$$

$$T^* = c_H \Delta \theta, \quad (30)$$

$$q^* = c_E \Delta q, \quad (31)$$

and inverse of Monin-Obukhov length:

$$L^{-1} = \frac{\kappa g}{u^{*2}} \left( \frac{T^*}{T_v} + \frac{q^*}{1/0.606 + q_a} \right), \quad (32)$$

where  $\kappa$  is von Kármán's constant,  $g$  acceleration due to gravity, and  $T_v$  virtual temperature:

$$T_v = (1 + 0.606 \times q_a) T_a. \quad (33)$$

In turn, bulk coefficients are

$$c_D = \frac{c_{DN}}{1 + \kappa^{-1} c_{DN} [\ln(z_u/z_0) - \psi_u(z_u/L)]}, \quad (34)$$

$$c_H = \frac{c_{HN}}{1 + \kappa^{-1} c_{HN} [\ln(z_T/z_0) - \psi_T(z_T/L)]}, \quad (35)$$

$$c_E = \frac{c_{EN}}{1 + \kappa^{-1} c_{EN} [\ln(z_q/z_0) - \psi_q(z_q/L)]}, \quad (36)$$

where  $z_u = 10$  m and  $z_q = 2$  m are observed heights of  $\|V_a\|$  and  $q_a$ , respectively, and  $z_0 = 10$  m is a reference height. Bulk coefficients at  $z_0$  with neutral stability,  $c_{DN}$ ,  $c_{HN}$ , and  $c_{EN}$ , are given as:

$$c_{DN} = (0.0027 \times \|V_a\|^{-1} + 0.000142 + 0.0000764 \times \|V_a\|^{1/2}), \quad (37)$$

$$c_{HN} = \begin{cases} 0.0180 & \text{stable} \\ 0.0327 & \text{unstable} \end{cases} \quad (38)$$

$$c_{EN} = 0.0346, \quad (39)$$

and stability functions,  $\psi_u$ ,  $\psi_T$ , and  $\psi_q$ , are

$$\psi_u(z/L) = \begin{cases} -5z/L & \text{stable} \\ 2 \ln [(1+X)/2] + \ln [(1+X^2)/2] & \\ -2 \tan^{-1} X + \pi/2 & \text{unstable} \end{cases} \quad (40)$$

$$\psi_T(z/L) = \psi_q(z/L) = \begin{cases} -5z/L & \text{stable} \\ 2 \ln [(1+X^2)/2] & \text{unstable} \end{cases} \quad (41)$$

with  $X = (1 - 16 z/L)^{1/4}$  where ‘‘stable’’ means the case that  $z/L$  is positive. This procedure is iterated twice from the first guess  $c_D = c_{DN}$ ,  $c_H = c_{HN}$ , and  $c_E = c_{EN}$ .

### 5.1.2 Formation Rate of Sea-Ice

Ocean surface heat flux  $Q_w$  is defined as:

$$Q_w = H_{lat} \downarrow + H_{sen} \downarrow + \epsilon_w R_{LW} \downarrow + (1 - \alpha_w) R_{SW} \downarrow - \epsilon_w R_w \uparrow, \quad (42)$$

where  $H_{lat} \downarrow$  is latent heat flux,  $H_{sen} \downarrow$  sensible heat flux,  $R_{LW} \downarrow$  downward long-wave radiation flux,  $R_{SW} \downarrow$  downward short-wave radiation flux,  $R_w \uparrow$  upward long-wave radiation flux,  $\epsilon_w$  seawater emissivity, and  $\alpha_w$  seawater albedo (Fig. 3a). Upward long-wave radiation  $R_w \uparrow$  is expressed by means of ocean surface temperature  $T_w$  as:

$$R_w \uparrow = \sigma T_w^4, \quad (43)$$

where  $\sigma$  is Stefan-Boltzman constant.

Here we evaluate  $Q_w^{frz}$  as

$$Q_w^{frz} = \rho_w c_{pw} \frac{\Delta z_w}{\Delta t} (T_w^{frz} - T_w), \quad (44)$$

where  $\rho_w$  is seawater density,  $c_{pw}$  specific heat of seawater,  $\Delta z_w$  the thickness of the top level of the ocean component including free surface,  $\Delta t$  time step,  $T_w^{frz}$  freezing point of seawater, and  $T_w$  ocean temperature of the top level.  $Q_w^{frz}$  is a negative heat flux necessary for water column with temperature  $T_w$  and thickness  $\Delta z_w$  to be frozen within a time  $\Delta t$ .

If ocean surface heat flux  $Q_w$  is greater than  $Q_w^{frz}$ , no sea-ice is formed and  $Q_w$  itself is used as a thermal boundary condition for the ocean component,  $Q_{aw}$ . When  $Q_w$  is less than  $Q_w^{frz}$ , their difference is utilized for sea-ice formation and in this case  $Q_{aw} = Q_w^{frz}$ . That is, formation rate of sea-ice,  $F_w$ , and residual heat flux into the ocean,  $Q_{aw}$ , are expressed as:

$$F_w = \max \left( 0, \frac{Q_w^{frz} - Q_w}{\rho_i L_{mlt}} \right), \quad (45)$$

$$Q_{aw} = \max (Q_w, Q_w^{frz}), \quad (46)$$

where  $\rho_i$  is sea-ice density and  $L_{mlt}$  latent heat of fusion.

## 5.2 Sea-Ice Region

### 5.2.1 Surface Fluxes

Sea-ice/snow surface heat flux  $Q_{sfc} \downarrow$  is defined as:

$$Q_{sfc} \downarrow = H_{lat} \downarrow + H_{sen} \downarrow + \epsilon_{sfc} R_{LW} \downarrow + (1 - \alpha_{sfc}) R_{SW} \downarrow - \epsilon_{sfc} R_{sfc} \uparrow, \quad (47)$$

where  $H_{lat} \downarrow$  is latent heat flux,  $H_{sen} \downarrow$  sensible heat flux,  $R_{LW} \downarrow$  downward long-wave radiation flux,  $R_{SW} \downarrow$  downward short-wave radiation flux,  $R_{sfc} \uparrow$  upward long-wave radiation flux,  $\epsilon_{sfc}$  surface emissivity, and  $\alpha_{sfc}$  surface albedo. Upward long-wave radiation  $R_{sfc} \uparrow$  is expressed as a function of surface temperature  $T_{sfc}$  as:

$$R_{sfc} \uparrow = \sigma T_{sfc}^4, \quad (48)$$

where  $\sigma$  is Stefan-Boltzman constant.

When sea-ice is not covered with snow ( $h_s = 0$ ), surface emissivity  $\epsilon_{sfc}$  is equal to sea-ice emissivity  $\epsilon_i$  and surface albedo  $\alpha_{sfc}$  is expressed with distinction between wet (melting) and dry ices as follows:

$$\alpha_{sfc} = \begin{cases} \alpha_i^{wet} & (T_{sfc} > T_i^{melt} - 0.001) \\ \alpha_i^{dry} & (T_{sfc} \leq T_i^{melt} - 0.001) \end{cases} \quad (49)$$

where  $T_i^{melt}$  is melting point of sea-ice. Otherwise ( $h_s > 0$ ), surface emissivity  $\epsilon_{sfc}$  is equal to snow emissivity  $\epsilon_s$  and surface albedo  $\alpha_{sfc}$  is expressed with distinction between wet (melting) and dry snows as follows:

$$\alpha_{sfc} = \begin{cases} \alpha_s^{wet} & (T_{sfc} > T_s^{melt} - 0.001) \\ \alpha_s^{dry} & (T_{sfc} \leq T_s^{melt} - 0.001) \end{cases} \quad (50)$$

where  $T_s^{melt}$  is melting point of snow.

When the option `iarc_flg` is enabled, sensible and latent heat fluxes,  $H_{sen} \downarrow$  and  $H_{lat} \downarrow$ , are evaluated as a function of sea-ice/snow surface temperature  $T_{sfc}$  using surface air temperature  $T_a$ , surface specific humidity  $q_a$ , and surface wind speed  $\|V_d\|$  in subroutine `budgets` as follows:

$$H_{sen} \downarrow = \rho_a c_{pa} C_H \|V_d\| (T_a - T_{sfc}), \quad (51)$$

$$H_{lat} \downarrow = \rho_a L_{sub} C_E \|V_d\| (q_a - q_{sfc}), \quad (52)$$

where  $\rho_a$  is air density,  $c_{pa}$  specific heat of air,  $L_{sub} = L_{melt} + L_{vap}$  latent heat of sublimation, and  $C_H$ ,  $C_E$  bulk transfer coefficients (constants). Saturated specific humidity  $q_{sfc}$  is calculated as:

$$q_{sfc} = \frac{0.622}{p_0} \times 6.11 \times 10^2 \times \exp\left(21.8746 \times \frac{T_{sfc} - T_0}{T_{sfc} - T_0 + 265.5}\right), \quad (53)$$

where  $p_0 = 1013 \times 10^2$  Pa is a reference air pressure. Variations of sensible and latent heat fluxes with respect to surface temperature are expressed as:

$$\frac{\partial H_{sen} \downarrow}{\partial T_{sfc}} = -\rho_a c_{pa} C_H \|V_d\|, \quad (54)$$

$$\frac{\partial H_{lat} \downarrow}{\partial T_{sfc}} = -\rho_a L_{sub} C_E \|V_d\| q_{sfc} \frac{21.8746 \times 265.5}{(T_{sfc} - T_0 + 265.5)^2}, \quad (55)$$

These values are used to solve surface temperature  $T_{sfc}$  described below.

Snowfall  $W_{snow}$  is parameterized from precipitation  $W_{prcp}$  and surface air temperature  $T_a$  [16, 3]:

$$W_{snow} = \frac{\rho_f}{\rho_s} r_s W_{prcp}, \quad (56)$$

where

$$r_s = \begin{cases} 1.0 & (T_a - T_0 < -5) \\ 1.0 - (T_a - T_0 + 5) \times 0.1 & (-5 < T_a - T_0 < 5) \\ 0.0 & (T_a - T_0 > 5) \end{cases} \quad (57)$$

is the fraction of precipitation falling as snow. Snowfall contributes to snow growth over sea-ice and the remainder, rainfall  $W_{rain} = (1 - r_s) W_{prcp}$ , is treated as surface runoff flowing into the ocean.

### 5.2.2 Snow-Free Sea-Ice

If sea-ice is not covered with snow ( $h_s = 0$ ), sea-ice surface temperature  $T_{sfc}$  is solved to satisfy the energy balance there (Fig. 3b).

A fraction  $I_0$  of the net incident short-wave radiation,  $(1 - \alpha_{sfc}) R_{SW} \downarrow$ , penetrates the bare sea-ice. In order to parameterize this process within zero-layer thermodynamics model, some part of  $I_0$  is assumed to be absorbed by sea-ice and used for surface heat budget, and remaining penetrates sea-ice. This penetration of short-wave radiation,  $R_{SW}^p \downarrow$ , is parameterized as [17, 3]:

$$R_{SW}^p \downarrow = I_0 (1 - \alpha_{sfc}) R_{SW} \downarrow \exp(-1.5 h_i) \quad (58)$$

where  $h_i$  is sea-ice thickness.

Conductive heat flux of sea-ice,  $G_i \downarrow$ , becomes

$$G_i \downarrow = \frac{k_i}{h_i} (T_{sfc} - T_{btm}), \quad (59)$$



where  $k_i$  is thermal conductivity of sea-ice,  $h_i$  sea-ice thickness, and  $T_{btm} = T_w^{frz}$  sea-ice bottom temperature.

The energy balance at the sea-ice surface is

$$\Delta Q_{sfc} \equiv Q_{sfc} \downarrow - R_{SW}^p \downarrow - G_i \downarrow = 0, \quad (60)$$

and sea-ice surface temperature  $T_{sfc}$  is calculated iteratively using Newton-Raphson method (5 times) to maintain this energy balance:

$$\begin{aligned} \Delta Q_{sfc} [T_{sfc}^{(i+1)}] &\approx \Delta Q_{sfc} [T_{sfc}^{(i)}] \\ &+ \left. \frac{\partial \Delta Q_{sfc}}{\partial T_{sfc}} \right|_{T_{sfc}^{(i)}} (T_{sfc}^{(i+1)} - T_{sfc}^{(i)}) \\ &= 0, \end{aligned} \quad (61)$$

or sea-ice surface temperature is updated as

$$T_{sfc}^{(i+1)} = T_{sfc}^{(i)} - \Delta Q_{sfc} [T_{sfc}^{(i)}] / \left. \frac{\partial \Delta Q_{sfc}}{\partial T_{sfc}} \right|_{T_{sfc}^{(i)}}, \quad (62)$$

where  $T_{sfc}^{(i)}$  is  $i$ -th iteration result and  $\partial \Delta Q_{sfc} / \partial T_{sfc}$  is

$$\frac{\partial \Delta Q_{sfc}}{\partial T_{sfc}} \equiv \frac{\partial H_{lat} \downarrow}{\partial T_{sfc}} + \frac{\partial H_{sen} \downarrow}{\partial T_{sfc}} - 4\epsilon_{sfc} \sigma T_{sfc}^3 - \frac{k_i}{h_i}. \quad (63)$$

### 5.2.3 Snow-Covered Sea-Ice

When sea-ice is covered with snow ( $h_s > 0$ ), snow surface temperature  $T_{sfc}$  is solved to satisfy the energy balances at the snow surface and at the snow/sea-ice interface (Fig. 3c).

In this case, no penetration of short-wave radiation is considered:

$$R_{SW}^p \downarrow = 0. \quad (64)$$

Conductive heat fluxes of snow,  $G_s \downarrow$ , and sea-ice,  $G_i \downarrow$ , become

$$G_s \downarrow = \frac{k_s}{h_s} (T_{sfc} - T_{ifc}), \quad (65)$$

$$G_i \downarrow = \frac{k_i}{h_i} (T_{ifc} - T_{btm}), \quad (66)$$

where  $k_s$  is thermal conductivity of snow,  $k_i$  thermal conductivity of sea-ice,  $h_s$  snow depth,  $h_i$  sea-ice thickness,  $T_{ifc}$  snow/sea-ice interface temperature, and  $T_{btm} = T_w^{frz}$  sea-ice bottom temperature.

The energy balance at the snow surface is [6, Eq. (23)]

$$\Delta Q_{sfc} \equiv Q_{sfc} \downarrow - G_s \downarrow = 0, \quad (67)$$

and the energy balance at the snow/sea-ice interface is

$$\Delta Q_{ifc} \equiv G_s \downarrow - G_i \downarrow = 0. \quad (68)$$

From Eq. (68), interface temperature  $T_{ifc}$  is expressed as a function of snow surface temperature  $T_{sfc}$  as:

$$T_{ifc} = \frac{k_s h_i T_{sfc} + k_i h_s T_{btm}}{k_s h_i + k_i h_s}. \quad (69)$$

Therefore,

$$G_s \downarrow = G_i \downarrow = \frac{k_s k_i}{k_s h_i + k_i h_s} (T_{sfc} - T_{btm}). \quad (70)$$

Now  $\Delta Q_{sfc}$  is a nonlinear function of  $T_{sfc}$ , and Eq. (67) is solved in the same way as for snow-free sea-ice, but in this case,

$$\begin{aligned} \frac{\partial \Delta Q_{sfc}}{\partial T_{sfc}} &\equiv \frac{\partial H_{lat} \downarrow}{\partial T_{sfc}} + \frac{\partial H_{sen} \downarrow}{\partial T_{sfc}} - 4\epsilon_{sfc} \sigma T_{sfc}^3 \\ &- \frac{k_s k_i}{k_s h_i + k_i h_s}. \end{aligned} \quad (71)$$

### 5.2.4 Growth Rate of Sea-Ice and Snow

Consider the case that sea-ice is covered with snow ( $h_s > 0$ ). If the snow surface temperature  $T_{sfc}$  exceeds melting point of snow,  $T_s^{mlt}$ , then  $T_{sfc}$  is fixed to  $T_s^{mlt}$  and the residual of snow surface heat flux,  $\Delta Q_{sfc} [T_s^{mlt}]$ , is used for snow melting:

$$F_s^{sfc} = \min \left( 0, \frac{\Delta Q_{sfc} [T_s^{mlt}]}{\rho_s L_{mlt}} \right), \quad (72)$$

where  $\rho_s$  is snow density and  $L_{mlt}$  latent heat of fusion.

If sea-ice is not covered with snow ( $h_s = 0$ ) and the sea-ice surface temperature  $T_{sfc}$  exceeds melting point of sea-ice,  $T_i^{mlt}$ , then  $T_{sfc}$  is fixed to  $T_i^{mlt}$  and the residual of sea-ice surface heat flux,  $\Delta Q_{sfc} [T_i^{mlt}]$ , is used for sea-ice melting:

$$F_i^{sfc} = \min \left( 0, \frac{\Delta Q_{sfc} [T_i^{mlt}]}{\rho_i L_{mlt}} \right), \quad (73)$$

where  $\rho_i$  is sea-ice density and  $L_{mlt}$  latent heat of fusion.

Growth rate of sea-ice thickness at the sea-ice bottom is calculated from energy imbalance there. When snow lies on sea-ice ( $h_s > 0$ ), conductive heat flux of sea-ice,  $G_i \downarrow$ , is expressed as a function of surface temperature  $T_{sfc}$  as

$$G_i \downarrow = \frac{k_s k_i}{k_s h_i + k_i h_s} (T_{sfc} - T_{btm}), \quad (74)$$

where  $k_s$  and  $k_i$  are thermal conductivities of snow and sea-ice, respectively, and  $T_{btm} = T_w^{frz}$  is sea-ice bottom temperature. This expression is valid even for the case of no snow ( $h_s = 0$ ). Turbulent heat flux through ocean/sea-ice interface,  $H_{iw\downarrow}$ , is parameterized as [18, 3]:

$$H_{iw\downarrow} = \rho_w c_{pw} C_t (T_{btm} - T_w), \quad (75)$$

where  $\rho_w$  is seawater density,  $c_{pw}$  specific heat of seawater,  $C_t$  the bulk transfer coefficient (a namelist parameter with dimension of  $m s^{-1}$ ), and  $T_w$  ocean surface temperature. At the sea-ice bottom, growth rate of sea-ice thickness,  $F_i^{btm}$ , is expressed as:

$$F_i^{btm} = -\frac{G_i\downarrow - H_{iw\downarrow}}{\rho_i L_{mli}}, \quad (76)$$

where  $\rho_i$  is sea-ice density and  $L_{mli}$  latent heat of fusion. Positive (negative)  $F_i^{btm}$  represents sea-ice formation (melting).

The aging process of snow is parameterized as a rate  $F_{s2i}$  at which snow is changed into ice [7]:

$$F_{s2i} = \gamma_s h_s + \frac{1}{\Delta t} \max\left(0, h_s - \frac{\rho_w - \rho_i}{\rho_s} h_i\right). \quad (77)$$

The first term represents an assumption that conversion of snow into ice simply depends on temporal scale  $\gamma_s = 2.0 \times 10^{-7} s^{-1}$  and snow depth  $h_s$ . The second term expresses the case that accumulated snow suppress the sea-ice surface below the sea level.

Finally, growth rates of snow depth,  $F_s$ , and sea-ice thickness,  $F_i$ , at sea-ice region are calculated as follows:

$$F_s = F_s^{sc} - F_{s2i}, \quad (78)$$

$$F_i = F_i^{sc} + F_i^{btm} + \frac{\rho_s}{\rho_i} F_{s2i}. \quad (79)$$

### 5.3 Total Growth Rate of Sea-Ice and Snow

Growth rate of sea-ice thickness,  $F_i^{eff}$ , is expressed as a combination of growth rates at open ocean,  $F_w$ , and sea-ice region,  $F_i$ :

$$F_i^{eff} = (1 - A_i) F_w + A_i F_i. \quad (80)$$

Growth rate of snow depth,  $F_s^{eff}$ , is given as a sum of growth rate at sea-ice region,  $F_s$ , and snowfall  $W_{snw}$  minus sublimation  $W_{sub} = -H_{lat\downarrow} / L_{sub}$ :

$$F_s^{eff} = A_i (F_s + W_{snw} - W_{sub}). \quad (81)$$

Growth rate of sea-ice concentration,  $F_A$ , is

$$F_A = (1 - A_i) F_w / h_0 + A_i \min(0, F_i) / 2h_i^{eff}, \quad (82)$$

where  $h_0 = 0.2$  m is cut off thickness between thick and thin ice.

## 6. Boundary Conditions for the Ocean Component

Finally, momentum flux  $\tau_{surf}$ , short-wave radiation flux  $Q_{SW\downarrow}$ , total heat flux  $Q_{surf}$ , and fresh water flux  $W_{surf}$  at ocean surface are calculated in subroutine *giosbc* to drive ocean component of OIFES.

### 6.1 Momentum Flux

If there is no sea-ice, the momentum flux through the sea surface,  $\tau_{surf}$ , is composed as [2, Eq. (4.32)]:

$$\tau_{surf} = \tau_{wind} + \tau_{fresh} \quad (83)$$

where  $\tau_{wind}$  is the wind stress and  $\tau_{fresh}$  represents momentum transfer in relation with a fresh water flux. This equation is extended to the region covered with sea-ice as:

$$\tau_{surf} = (1 - A_i) \tau_{aw} + A_i \tau_{iw} + \tau_{fresh} \quad (84)$$

where  $\tau_{aw}$  is direct momentum input from the atmosphere and  $\tau_{iw}$  the stress at ocean/sea-ice interface determined from sea-ice dynamics.

### 6.2 Heat Flux

Surface heat flux due to downward short-wave radiation,  $Q_{SW\downarrow}$ , is expressed as a combination of direct short-wave radiation flux at open ocean and penetrative flux below sea-ice (without snow cover):

$$Q_{SW\downarrow} = (1 - A_i) (1 - \alpha_w) R_{SW\downarrow} + A_i R_{SW\downarrow}^p, \quad (85)$$

where  $\alpha_w$  is seawater albedo. Total heat flux  $Q_{surf}$  is

$$Q_{surf} = (1 - A_i) (Q_{aw} + H_{snw\downarrow}) + A_i (R_{SW\downarrow}^p + H_{iw\downarrow}), \quad (86)$$

where  $Q_{aw}$  is net heat flux between atmosphere and ocean (including short-wave radiation flux),  $H_{snw\downarrow} = -\rho_s L_{mli} W_{snw}$  latent heat of snow melting<sup>3</sup>, and  $H_{iw\downarrow}$  turbulent heat flux at sea-ice/ocean interface.

<sup>3</sup> This process is considered only if the option *cfes* is enabled.

### 6.3 Fresh Water Flux

Total fresh water flux  $W_{surf}$  is expressed as follows:

$$W_{surf} = (1 - A_i) (W_{prcp} - W_{evap}) + A_i (W_{rain} + W_{ice}) + W_{rivr}, \quad (87)$$

where  $W_{prcp} = W_{rain} + W_{snw}$  is total precipitation,  $W_{evap} = -H_{lat} \downarrow / \rho_w L_{vap}$  evaporation,  $W_{rain}$  sea-ice surface runoff due to rainfall,  $W_{ice}$  fresh water flux due to sea-ice formation/melting and snow melting, and  $W_{rivr}$  river runoff.

## 7. Code Optimization

Here, some issues on code optimization are described. Generally speaking, computational cost of a sea-ice component is much smaller than that of an ocean component in a coupled ocean–sea-ice model, because calculation of sea-ice is, at most, two-dimensional while calculation of ocean is three-dimensional. However, OFES has excellent computational efficiency for the Earth Simulator, so the sea-ice component of OIFES is carefully optimized not to drag down the total computational performance.

Source code of the sea-ice component is optimized in a similar manner applied to OFES. First of all, most of DO loops are vectorized in the zonal direction. This is an important work for vector computers like the Earth Simulator. Then, they are parallelized using MPI library such as MPI\_SEND and MPI\_RECV. We employ one-dimensional domain decomposition in the meridional direction as for OFES.

Table 4 shows results obtained from “flow trace analysis” on major subroutines relating to the sea-ice component for the case in which the horizontal resolution is  $0.25^\circ$  in both longitude and latitude. We used 45 computational nodes of the Earth Simulator with 360 MPI processes. Note that in this case the numbers of grid

points in the zonal and the meridional directions are 1440 and 720, respectively, and therefore, each MPI process treats 2 zonal strips as computational domain with 4 halo regions (2 for the northern side and 2 for the southern side). Execution times of these subroutines are 1.4% at most, and the total execution time of the sea-ice component is about 6%.<sup>4</sup> Their vector operation ratios are about 99% and averaged vector lengths are about 240, which is close to the maximum value of 256. Some of them (*asflx*, *form*, and *budgets*) attain the flops values over 40% of the theoretical peak performance of 8 Gflops. On the other hand, subroutines *plast*, *ice\_slow*, and *advdif* exhibit the low flops values, because the communication costs are relatively large in these subroutines, and *ice\_slow* is the main driving routine for the sea-ice component.

## 8. Preliminary Results

In this section, some preliminary results on sea-ice obtained from a coupled atmosphere–ocean simulation using AFES and OIFES, namely CFES, are presented in order to show potential of our sea-ice model.

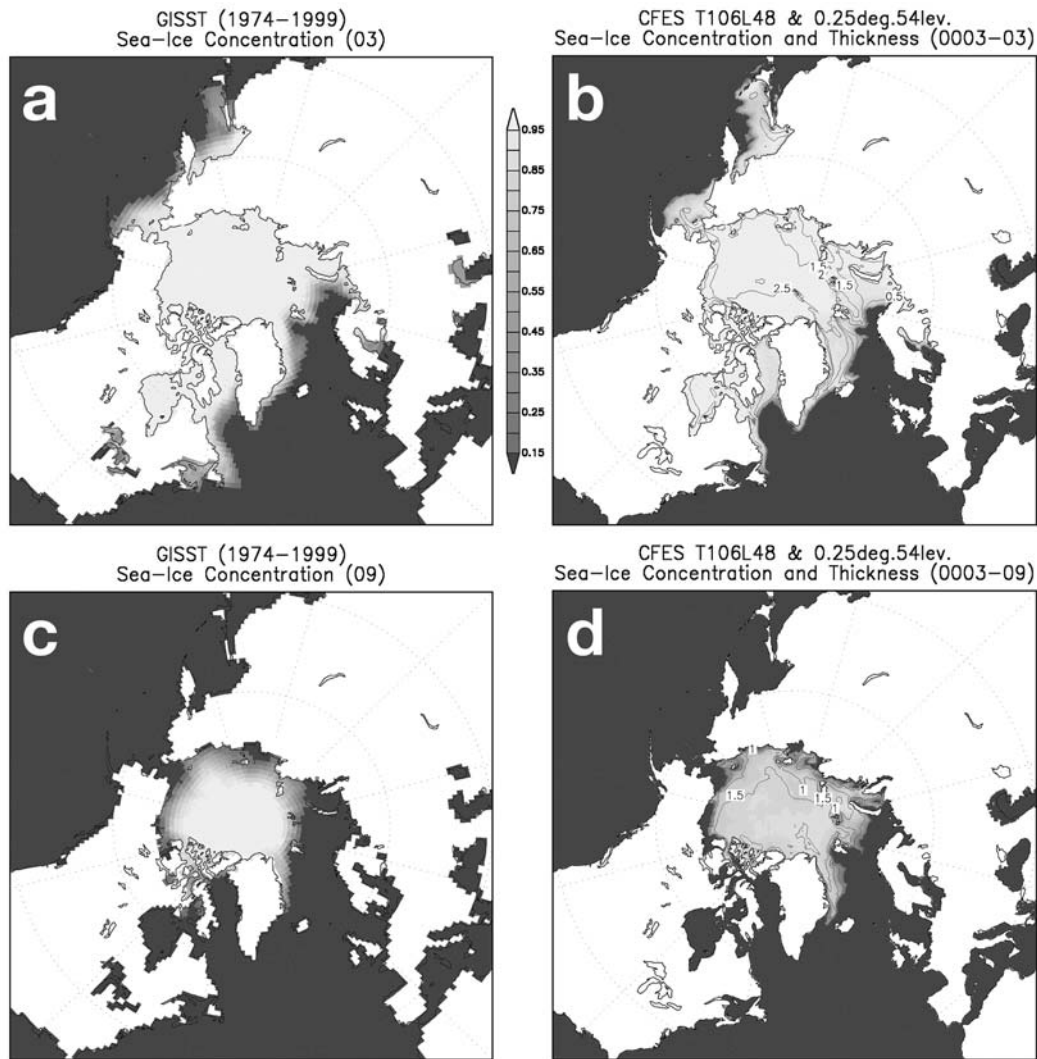
### 8.1 Setting of Simulation

Brief overview of a coupled simulation is as follows. Horizontal resolution of the ocean/sea-ice component is  $0.25^\circ$  in both longitude and latitude. It contains 54 levels in vertical and the top level has thickness of 5 m. Parameters relating to sea-ice are summarized in Table 3. Initial conditions of the ocean component are climatological temperature and salinity of the World Ocean Atlas 1998 [19, 20, 21, 22, 23, 24] with no motion. Model resolution of the atmospheric component is T106 (triangle truncation of wave number 106,  $\sim 1.1^\circ$ ) in horizontal and 48 layers in vertical. Detailed model configurations may be presented in another report. Monthly-mean simulation results of the 3rd model year rather than climatological ones are shown below due to short of integration time.

**Table 4** Results obtained from “flow trace analysis”.

Subroutine	Description	Exec. Time [%]	MFLOPS	V. Ratio [%]	V. Length
<i>plast</i>	strain rate and viscosity	1.4	741.3	98.56	240.6
<i>asflx</i>	air–sea flux	1.0	3303.1	99.60	240.9
<i>ice_slow</i>	main routine for sea-ice	0.8	44.7	92.59	223.3
<i>advdif</i>	advection and diffusion	0.7	467.7	98.03	238.1
<i>form</i>	momentum equation	0.6	3425.0	99.30	240.6
<i>budgets</i>	surface energy budget	0.6	3729.9	99.46	206.6
<i>elast</i>	elastic equation	0.5	2393.0	98.92	237.8

<sup>4</sup> These values depend on a number of elastic subcycle,  $N_{evp}$ .



**Fig. 4** (a) (c) Observed and (b) (d) simulated sea-ice concentration in the Arctic region. Effective sea-ice thickness is overplotted in (b) and (d). Contour interval is 0.5 m. (a) (b) March. (c) (d) September.

### 8.2 Arctic Region

Figure 4 shows comparison between observed and simulated sea-ice concentration in the Arctic region in March (Fig. 4a, b) and in September (Fig. 4c, d). Climatological monthly-mean observations are calculated using global sea-ice concentration data of GISST [25] from 1974 through 1999. Seasonal variation of simulated sea-ice extent well captures the observed feature such that a main portion of sea-ice remains throughout the year. Simulated effective sea-ice thickness is also shown in Fig. 4. Overall pattern that sea-ice is thicker on the Greenland side than on the Eurasian side is realistically reproduced, but the sea-ice thickness is about half of the typical observed value (3–5 m).

In Fig. 5, simulated sea-ice velocity in the Arctic Ocean is plotted. Two major structures, the Trans-Polar Drift on the Eurasian side of the Ocean and the Beaufort

Gyre in the Canada Basin, are well reproduced in March (Fig. 5a), while reversal of the Trans-Polar Drift in September (Fig. 5b) may be unrealistic.

### 8.3 Antarctic Region

Simulated sea-ice concentration in the Antarctic region is compared with climatological observation (GISST) in Fig. 6. In both seasons (February and August) they look similar. However, absence of sea-ice around the Antarctic coast in February should be pointed out as a noticeable problem. Simulated sea-ice thickness in winter (Fig. 6d) is acceptable.

Figure 7 shows simulated sea-ice velocity around the Antarctica. Clockwise circulations corresponding to the Weddell Gyre and the Ross Gyre are clearly seen in August (Fig. 7b).

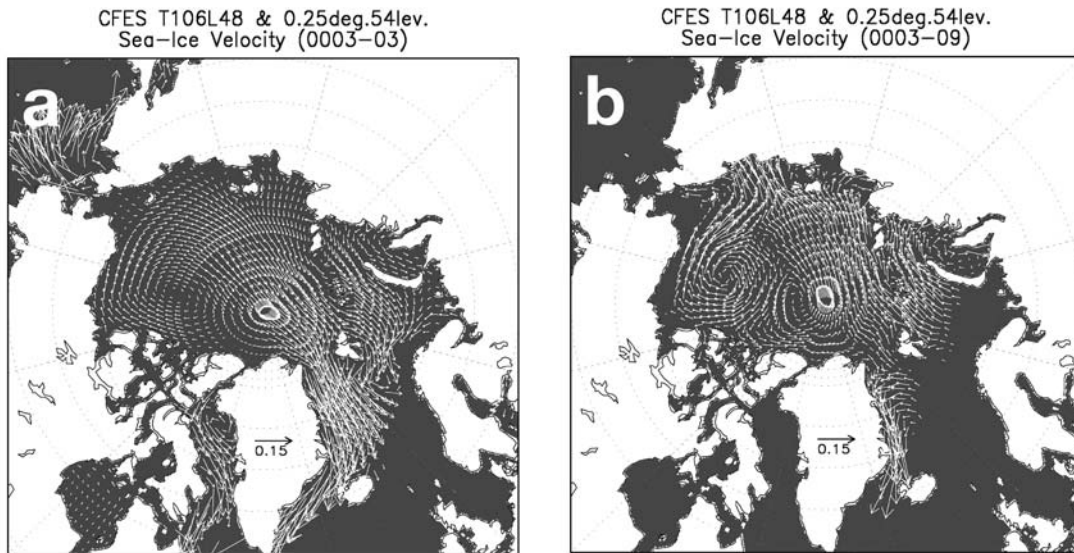


Fig. 5 Simulated sea-ice velocity [ $\text{m s}^{-1}$ ] in the Arctic region. (a) March. (b) September.

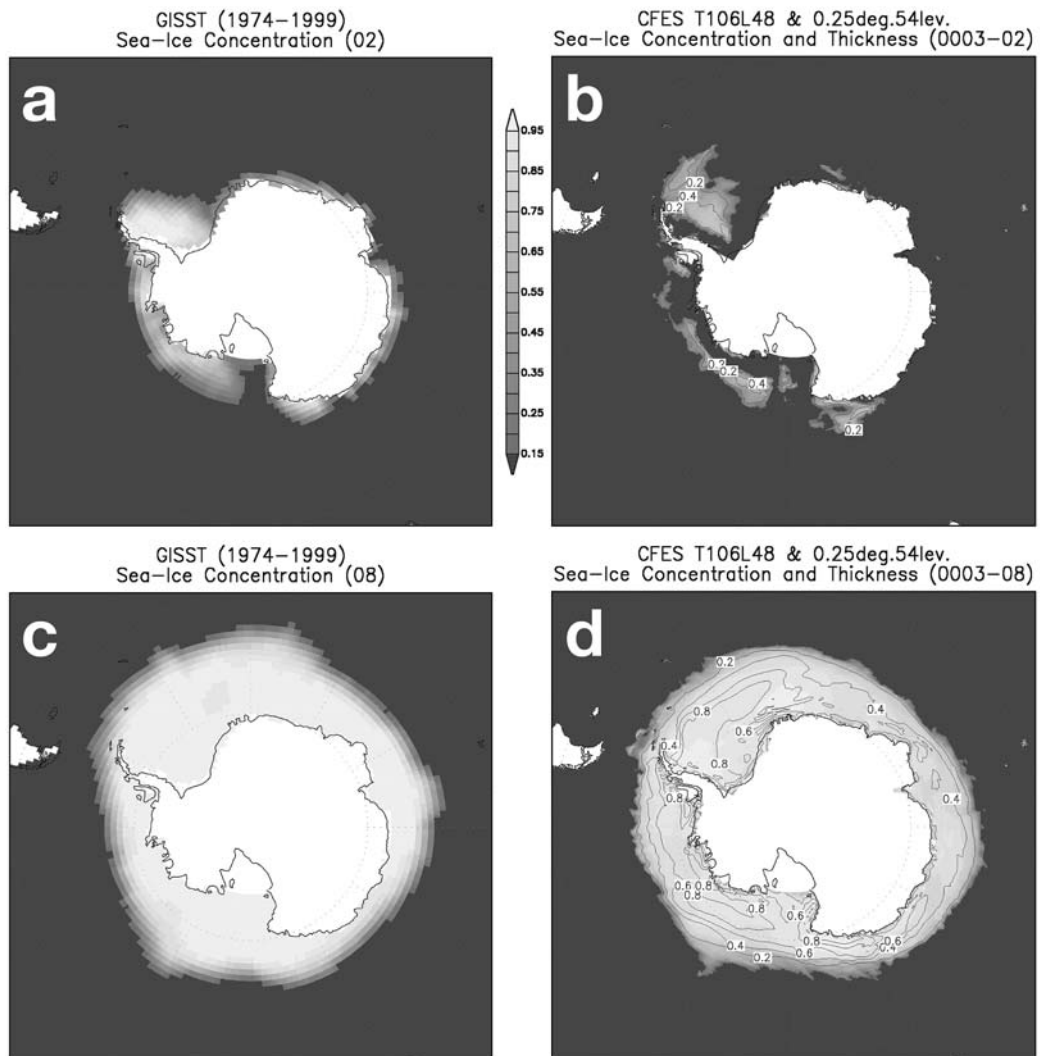


Fig. 6 Same as Fig. 4 but in the Antarctic region and contour interval is 0.2 m. (a) (b) February. (c) (d) August.

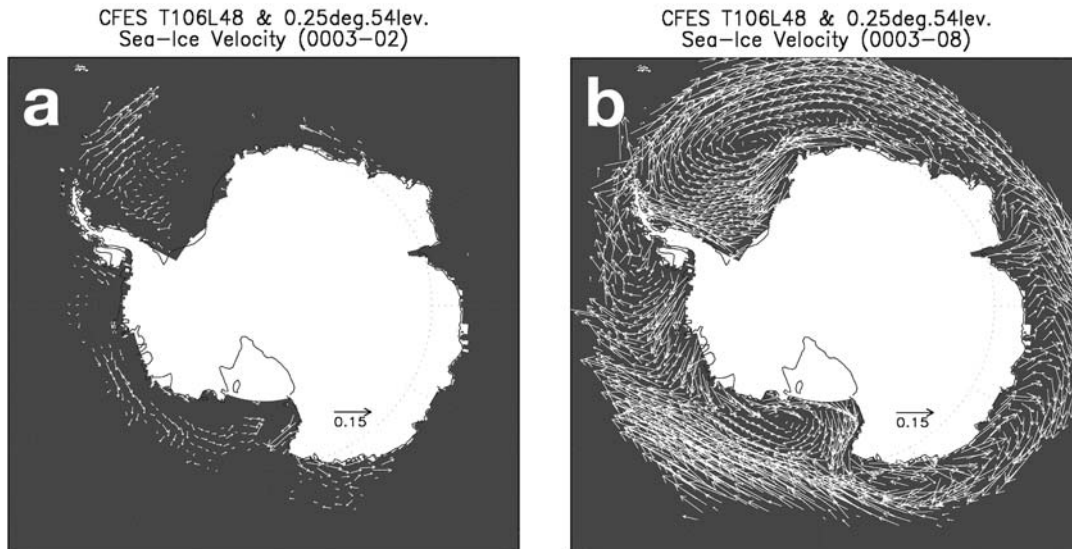


Fig. 7 Same as Fig. 5 but in the Antarctic region. (a) February. (b) August.

### 8.4 Short Summary

Our coupled ocean–sea-ice model, OIFES, seems to have the capability of simulating the global sea-ice variability: simulated sea-ice concentration, thickness, and velocity fields are fairly good in reproducing observed features in both the Arctic and the Antarctic regions and in both summer and winter. Some discrepancies found between observation and simulation may be reduced by making efforts at tuning model parameters of both ocean and sea-ice components, and longer model integration is necessary for detailed comparison with observation.

### Acknowledgements

The authors would like to thank Prof. Moto Ikeda for his valuable advice and Dr. Toru Miyama for his significant contribution to improvement of OIFES. We also thank Messrs. Takashi Abe and Akira Azami for their help in optimizing the source code of OIFES. Development of OIFES is partly supported by MEXT (Ministry of Education, Culture, Sports, Science and Technology) RR2002 Project for Sustainable Co-existence of Human, Nature and the Earth, category 7 (PI: Prof. Toshiyuki Awaji).

(This article is reviewed by Dr. Tetsuya Sato.)

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