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Implementation of LIM sea ice model in the CMCC global ocean high-resolution configuration

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SUMMARY This report describes the sea ice model LIM2 and how it is coupled to the ocean component of the NEMO system in the global sea ice-ocean 1/16^o configuration under development at CMCC.

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CONTENT

1. INTRODUCTION

Sea ice is both an important actor of the global climate system and an early indicator of short-term climate change. Seasonal and interannual variations of the sea ice cover represent some of the most pronounced signals of variability in the Earth's climate system. Sea ice reaches in surface up to 10% of the world ocean during the Austral winter. In the Northern Hemisphere, the ice reaches a maximum extent of 15 $\times$ 10^6 km^2 in late winter, covering the entire Arctic Ocean and many peripheral seas; the minimum extent occurs in September, when ice is confined to the central Arctic. In the Southern Hemisphere, most ice that forms in winter disappeares in summer; the total extent ranges from about 18 $\times$ 10^6 km^2 in September to 3$\times$$10^6$ km^2 at the end of February (US National Snow and Ice Data Center, http://nsidc.org/data/).
Sea ice largely modifies the direct air-ocean exchanges of heat and momentum, thereby inhibitinging the convective cooling in the ocean, and lessens and delays the spring warming through its high albedo [1, 2] and low thermal conductivity [3]. The presence of sea ice keeps the temperature of the surface ocean close to the freezing point of sea water. In addition, the rejection of salty water during sea ice growth and the release of nearly freshwater during sea ice melting induce sea surface salinity variations that alter the density structure of the ocean, which is critical to the circulation of the high-latitude oceans and control the amount and location of dense water formation. The salt-enriched waters spread off the ice edge to adjacent unfrozen waters by horizontal advection and mixing. In some specific sites, the resulting cold and saline water sinks convectively, spreads throughout the ocean depth,maintaining the thermohaline circulation.

The large sensitivity of sea ice to external forcing is due, first, to positive atmosphere-ice-ocean feedbacks and, second, to the fact that ice is much thinner (1-10 m) than horizontally wide (>1000 km). The main sea ice-related feedback mechanism is the ice-albedo feedback (e.g., [4]). A small positive perturbation in the surface energy budget triggers a decrease in ice coverage. This enhances the absorption of solar radiation in the upper ocean, which further increases melting at the base of the ice cover. Therefore small climate warming perturbations are amplified by sea ice.
In the context of the assessment of ongoing climate changes in polar regions and at a global scale, understanding the sea ice mass balance is therefore of crucial importance. The importance of sea ice processes, which are difficult to observe directly, provides a direct motivation of the numerous modelling studies on ocean circulation, sea ice-ocean interactions and the variability of these processes on seasonal and interannual time scales. Scientists started to realize the importance of sea ice for global ocean circulation since the 1980's.
In the last decades, dynamic-thermodynamic sea ice models have been coupled, first, to regional oceanic general circulation models in the Arctic (e.g. [5]) and the Antarctic (e.g. [6]. Early models had a coarse, typically 2^o, resolution (e.g. [7, 8]) and often had difficulties to achieve an overall realistic simulation of sea ice coverage and hydrography. Recently, global coupled sea ice-ocean models with eddy-permitting (typically $1/4^o$) and eddy-resolving resolutions have been developed and used for regional studies in the high latitudes (e.g. [9, 10]).
Here we describe the physical processes acting on sea ice (thermodynamic processes, which involve the transfer of heat or radiation, and dynamic processes, which move and

deform the ice) as represented in the version 2 of the Louvain-la-Neuve sea ice model (LIM2, [11, 7]), which is coupled to NEMO (Nucleus for European Modeling of the Ocean) [12] in the framework of the CMCC eddy-resolving global oceanic configuration.

2. SEA ICE MODEL

LIM is a numerical model of sea ice specifically designed for climate studies and operational oceanography. LIM is originally a 3D dynamic-thermodynamic model developed by [11]. The model component that determines the vertical growth and decay of ice due to themodynamic processes is based on the Semtner model [13] and the dynamic component is originally based on the Hibler viscous-plastic rheology [14]. LIM includes various parameterizations (open water, snow, snow ice formation, increased heat conduction through thin ice). The basis of the coupling with ocean models was introduced in the CLIO global coupled sea ice-ocean model [7]. LIM was then coupled to the ocean general circulation model OPA [15], leading to the ice-ocean model configuration ORCA2-LIM [8]. The code was then rewritten formally without significant physical changes, which resulted in LIM2 that is presently used in the reference version of the NEMO modeling framework. The code is freely available under a software licensing agreement (http://www.nemo-ocean.eu).
The physical processes governing the evolution of sea ice can be conceptually divided into two parts. The first category of processes regards the freezing and melting of sea ice (ice thermodynamics), which can be considered to depend only on the vertical response of the ice layer to the exchanges with the atmosphere and the ocean. Lateral growth and melt of sea ice are also described as vertical processes, being mainly a function of the vertical fluxes from the atmosphere and the ocean. The horizontal component of thermodynamical processes (e.g. heat conduction) can be safely neglected because of the much larger horizontal scales with respect to the vertical ones.
Since the ice is relatively thin and does not significantly move vertically, ice dynamics and transport are represented as 2D horizontal processes. This distinction between thermodynamics and dynamics is purely formal since they are intrinsically coupled. The ice growth is a function of the ice thickness and concentration, which in turns depends strongly on the advection pattern. Conversely, the motion of sea ice is to a great extent affected by the ice-thickness distribution which roughly controls the amount of stress the ice can sustain [14].
Each grid cell is assumed to be covered by a fraction of sea ice (ice concentration), the rest being ice-free, which provides a representation for the presence of open water in the pack. The ice-covered fraction of the cell is vertically divided in two layers of sea ice and one layer of snow. Each ice or snow layer is characterized by a temperature value. Brine inclusions within the sea ice are represented using a heat reservoir storing solar radiation in spring and releasing it in fall. This description of the sea ice in LIM2 relies on 7 state variables: ice concentration A, ice thickness h_i, snow depth h_s, the three layer temperatures, and the heat content of brine inclusions Q_{st}. Ice dynamics compute the ice drift as a function of wind, ocean and internal stresses.

3. SEA ICE THERMODYNAMICS

Ice thermodynamics corresponds to all the processes that involve energy transfer through and storage into the ice, and hence may lead to net growth and melt of ice. The purpose of LIM thermodynamics is to compute changes in snow depth (h_s), ice thickness (h_i), surface temperature (T_{su}) and vertical temperature profiles inside snow (T_s) and ice (T_i), given the changes in the atmospheric and oceanic forcings.

3.1 VERTICAL GROWTH AND DECAY

Vertical and lateral sea ice growth/decay rates are obtained from energy budgets at the upper and lower surfaces of the snow-ice cover, and at the surface of leads present within the ice pack. The representation of the vertical growth and decay of sea ice is an improved version of the Semtner's [13] three-layer model (one layer for snow and two layers for ice). All processes are assumed purely vertical.
Within the ice-covered portion of each grid cell, sea ice is considered as a homogeneous slab of ice (divided into two layers of equal thickness) covered by a snow layer when the surface temperature is below the melting point. The evolution of temperature inside the snow-ice system is governed by the one dimensional heat-diffusion equation

$$\rho c_p \frac{\partial T}{\partial t} = \left[1 + \frac{1}{2} ln\left(\frac{2h_e}{e\epsilon}\right)\right] k \frac{\partial^2 T}{\partial z^2} \qquad (1)$$

where ρ, c_p, and k are the density, the specific heat and thermal conductivity of ice or snow; T is the temperature, t is the time and z is the vertical coordinate. The correction term between square brackets represents the effect of the subgrid-scale snow and ice thickness distributions on the conductive heat flux [16, 17]. It varies both in time and space and is diagnosed by assuming that the snow and ice thicknesses are uniformly distributed between zero and twice their mean value over the ice-covered portion of the grid cell [11]. e is Euler's number (the base for natural logarithms). h_e is an effective thickness for heat conduction defined as

$$h_e = \frac{k_s k_i}{k_s + k_i}\left(\frac{h_s}{k_s} + \frac{h_i}{k_i}\right) \qquad (2)$$

where h_s and h_i are the thicknesses of snow or ice, respectively. ϵ is a threshold thickness that determines the limit of validity of (1). For $h_e < e\epsilon/2$, the correction term is taken equal to 1.
The surface temperature T^{ice}_{su} is determined by considering the budget of a thin layer of thickness h_{su} at the air-snow/ice interface:

$$h_{su}(\rho c_p)\frac{\partial T^{ice}_{su}}{\partial t} = Q_{sr}(1 - i_0)(1 - \alpha^{ice}) + Q^{ice}_{ns} + F_{cs} \qquad (3)$$

where Q_{sr} is the downelling shortwave radiation that reaches the snow/ice surface, i_0 is the fraction of net shortwave radiation that penetrates within the ice, i.e. that does not contribute to the surface energy balance ($i_0 = 0$ for snow-covered ice), the snow-ice surface albedo[1] α^{ice} is the fraction reflected directly into the atmosphere. Q^{ice}_{ns} is the non-solar net flux at the snow/ice surface (summing the longwave, sensible and latent heat fluxes contributions). F_{cs} is the conductive flux from the sea ice/snow interior towards the top surface, and it is parameterized by $2k_{i/s}(T_{i/s} - T_{su})/h_{i/s}$.
When this balance results in a temperature of the ice surface above the melting point, T_{su} is forced to 0^{o}C and the excess of energy B_{su} is compensated by melting ice/snow:

$$\left(\frac{\partial h_{i/s}}{\partial t}\right)_{su} = -\frac{B_{su}}{L_{i/s}} \qquad (4)$$

where L_i and L_s are the volumetric latent heat of fusion of ice and snow, respectively. At the bottom of the ice slab, any imbalance between the conductive heat flux within the ice, $F_{cb} = 2k_i(T_b - T_i)/h_i$, and the heat flux from the ocean to the ice, F_w, is compensated by creation or melting of ice, following

$$\left(\frac{\partial h_i}{\partial t}\right)_b = \frac{F_{cb} - F_w}{L_i} \qquad (5)$$

[1] In the CORE formulation for atm ospheric heat fluxes, only the clear-sky albedo is used.

where L_i is the volumetric latent heat of fusion of ice. Storage of latent heat inside the ice resulting from the trapping of shortwave radiation by brine pockets is taken into account.

So called *snow-ice* forms at the ice surface when the load of snow is large enough to depress the snow-ice interface under the sea level. It is assumed that snow-ice forms by the refreezing of seawater from the upper ocean level in the submerged snow layer. The change in ice thickness $(\Delta h_i)_{si}$ and snow depth $-(\Delta h_s)_{si}$ due to snow ice formation are equal in magnitude and opposite in sign, and given by:

$$(\Delta h_i)_{si} = -\frac{(\Delta h_s)_{si}}{\beta_{si}} = \frac{\rho_s h_s - (\rho_0 - \rho_i)h_i}{\beta_{si}\rho_s + \rho_0 - \rho_i}, \tag{6}$$

with β_{si} an empirical parameter taking into account the compaction of the soaked snow, and ρ_0 the density of seawater. The total mass of newly formed ice is supposed to consist of the mass of the soaked snow and the mass of the infiltrated seawater. The latent heat and brine release occurring during the freezing of the water component are transferred to the ocean.

3.2 LATER GROWTH AND DECAY

The sea ice pack is subject to atmospheric and oceanic momentum forcing, which can create openings in the cover. These areas of open water, leads and polynyas, are very important since they allow direct heat exchange between the ocean and the atmosphere. This fraction of open water is represented in the model by the introduction of the variable A, the ice concentration, defined as the fraction of each grid cell covered by ice. By increasing or decreasing its concentration, the sea ice grows or decays laterally. The thermodynamic changes of A are function of the heat budget of the open water area, B_l. If $B_l < 0$, ice with a characteristic thickness h_0 is formed in the open water portion of the grid cell, changing the ice concentration by:

$$\left(\frac{\partial A}{\partial t}\right)_{acc} = \sqrt{1 - A^2}\left[\frac{B_l(1-A)}{L_i}\right]\frac{1}{h_0} \tag{7}$$

The term between square brackets is the amount of ice created, replacing open water in the grid cell. The term inside the square root distributes the energy over the lateral and the vertical dimensions: only a fraction of the ice formed in the freezing open water area contributes to increase the ice concentration, the remaining energy leads to an increase in the thickness of the preexisting ice [18, 11]. Whenever lateral ice accretion occurs, the thickness of newly formed ice is averaged with that of pre-existing ice to obtain a single value and snow is distributed over the new ice-covered area.

The decrease of sea ice concentration is considered to be caused only by the vertical melting of thin ice. If $B_l > 0$, the entire heat gain in open water is supposed to contribute to basal melting. Assuming that the ice is uniformly distributed in thickness between 0 and $2h_i$ over the ice-covered fraction of the grid cell, and that the melting rate does not depend on the local ice thickness, the vertical melting necessarily decreases sea ice concentration by:

$$\left(\frac{\partial A}{\partial t}\right)_{abl} = \frac{A}{2h_i}\left(\frac{\partial h_i}{\partial t}\right)_{abl-acc} \tag{8}$$

when ice ablation $\geq$ accumulation, otherwise A does not change. If the ice concentration decreases, the snow present at the surface of the removed ice is piled up onto the remaining ice.

4. SEA ICE DYNAMICS

In order to simulate ice drift, sea-ice models generally include a momentum equation to compute the drift velocity. This involves a rheology (a constitutive law that describes the material properties of the ice). The state variables must resultingly be transported, which is done by adding a transport equation. Finally, ice motion may also involve mechanical deformation (opening and ridging/rafting of the ice). In LIM2, the ice pack (consolidated ice and leads) is treated as a two-dimensional compressible fluid driven by wind and oceanic stresses. Sea ice resists to deformation with a strength which increases monotonically with ice thickness and concentration. The ice velocity $\mathbf{u}$ is determined from the conservation of linear momentum [19]

$$m\frac{\partial \mathbf{u}}{\partial t} = A(\tau_a+\tau_w)-mf\mathbf{e}_z\times\mathbf{u}-mg\nabla\xi+\tau_i \quad (9)$$

where m is the combined mass of ice and snow per unit area, and $\mathbf{u}$ is the sea ice velocity. The term in bracket on the right-hand side represents the exchanges of momentum between atmosphere or ocean and sea ice (τ_a is the wind stress on the ice and τ_w is the ocean drag on the ice). The two following stresses are, respectively, due to the Coriolis effect (f, e_z are respectively the Coriolis parameter and the upward unit vector) and pressure gradients associated with the tilt of the ocean surface (g is the gravity acceleration, ξ the surface dynamic height). τ_i represents the internal stress of the ice and can be written in term of the divergence of the two dimensional stress tensor, $\nabla \cdot \sigma$. A constitutive equation, which describes the rheology of the material, is given for computing σ within the ice for different states of deformation. There are now two different rheologies available in LIM2. In the traditional method proposed by [14], sea ice is assumed to have a pure viscous-plastic (VP) constitutive law in which the internal ice stress is computed diagnostically in combination with (9). The VP constitutive law relates the internal ice stress σ and the rates of strain $\dot{\epsilon}$ through an internal ice pressure P and non-linear bulk and shear viscosities, ζ and η, such that the principal components of stress lie on an elliptical yield curve with e_c being the ratio of the principal axes of the ellipse. The internal stress is given by:

$$\sigma = 2\eta\dot{\epsilon} + \left[(\zeta-\eta)\,T(\dot{\epsilon}) - \frac{P}{2}\right]\mathbf{I} \qquad (10)$$

where $\mathbf{I}$ is the two-dimensional unity tensor. The non-linear viscosities increase with pressure and with decreasing strain rates: $\zeta = P/(2\Delta)$ and $\eta = \zeta/e_c^2$ with the deformation rate given by

$$\Delta = \left[D_D^2 + \frac{1}{e^2}(D_T^2 + D_S^2)\right]^{1/2} \qquad (11)$$

D_D, D_T and D_S are the the divergence, horizontal tension and shearing strain rate. The ice strength P depends on both ice thickness and fraction: ice strength decreases strongly if the ice concentration decreases and increases slowly (linearly) with ice thickness. It is parameterised as

$$P = P^* A h_i e^{-C(1-A)} \qquad (12)$$

where P^* and C are model constants. This rheology allows sea ice to flow plastically for high enough strain rates and to deform in a linear viscous manner for very small rates.

Sea ice tends to move as large plates with long faults related to shearing. Close examination using modern satellite data, however, shows that the VP model does not reproduce these features (e.g. [20, 9]). The enormous range of effective VP viscosities severely limits the time-step for stability of an explicit numerical scheme, especially in regions where the ice is relatively rigid. When run on parallel computers, implicit solution methods such as successive over-relaxation and line relaxation entail a great deal of communication between processors and are difficult to parallelize efficiently, unlike explicit methods.

The most popular alternative for the calculation of the VP dynamics is the elastic-viscous-plastic (EVP) formulation by [21], who modified the VP model by adding elastic behavior, thus realizing large gains in numerical efficiency by permitting a fully explicit implementation with an acceptably long time-step. This EVP model reduces to the original VP model on long timescales and was shown to be more accurate for transient behavior. An EVP rheology was introduced in LIM by [22].

The EVP model incorporates the constitutive law in the time-dependent equations. σ is computed prognostically. The strain rate is separated into the sum of a plastic contribution (given by (10) and an elastic part, approximated by

$$\frac{1}{E}\frac{\partial \sigma}{\partial t} = \dot{\epsilon} \tag{13}$$

where E, the Young's modulus, is a tunable coefficient that can be chosen to make the elastic term small compared to the other terms. The rheology equation is rewritten as

$$\frac{1}{E}\frac{\partial \sigma}{\partial t} + \frac{1}{2\eta}\sigma + \frac{1}{4\zeta}\left[\frac{\eta - \zeta}{\eta}\sigma + P\right]\mathbf{I} = \dot{\epsilon} \tag{14}$$

An explicit numerical solution of the momentum equation, taking into account (14), now involves a time step restriction. While the EVP rheology (14) becomes (10) in the steady state, static flow in the EVP rheology is prepresented by an elastic deformation and so imposing a minimum value of Δ is no longer necessary. The numerical solution of the momentum equation in combination with (14) does converge to the VP stationary solution as long as the elastic time scale is several times smaller than the time scale of variation of the external forcing [21].

While the VP case in LIM uses a B grid arrangement of the variables, [22] have introduced, in LIM2, a centred difference C-grid formulation of the EVP dynamics that allows the dynamical coupling to NEMO (also in C-grid) in a more natural way. They showed that EVP formulation implemented in LIM is able to produce realistic ice fields in both hemispheres. Because it allows to drastically reduce the numerical viscous flow limit, using EVP gives a better solution of the ice momentum equation. The EVP ice dynamics approaches plasticity better, numerically more efficiently than the previous VP implementation. EVP is explicit, which allows easier parallelization.

5. HORIZONTAL TRANSPORT

Changes of any variables, due to the large-scale transport by ice motion and to the thermodynamical processes are governed by a general conservation law:

$$\frac{\partial \Psi}{\partial t} = -\nabla \cdot (\mathbf{u}\Psi) + S_{\Psi} + D\nabla^2\Psi \tag{15}$$

where Ψ represents the physical variables that are transported in the model (ice concentration, snow volume per unit area, ice volume per unit area, snow and ice enthalpy per unit area, and latent heat content in the brine reservoir). S_{Ψ} is the rate of change of Ψ due to thermodynamics and D is a horizontal diffusivity to avoid nonlinear instabilities arising from the coupling between ice dynamics and advection. D is constant inside the ice pack and equal to 0 at the ice edge. In order to keep diffusion relatively small compared to model dynamics, D, resolution dependent, is set to 2 m^2/sec in our configuration.

The global ice state variables are transported horizontally using the advection scheme of [23], a numerical method based on the conservation of second-order moments of the spatial distribution of the advected quantities within each grid cell. It preserves the positiveness of the transported variables and yields small diffusion.

6. ICE-OCEAN COUPLING

The coupling between the sea ice and ocean follows the equations given by [7].

Momentum is exchanged at the ice-ocean interface through an ice-ocean stress deriving from a quadratic function of the difference between the ice velocity u_i and the ocean surface velocity u_w:

$$\tau_w^{ice} = \rho_0 c_w^{ice} |\mathbf{u_{ice}} - \mathbf{u_w}|(\mathbf{u_{ice}} - \mathbf{u_w}) \qquad (16)$$

with c_w^{ice} the drag coefficient. The heat fluxes from the ocean to the ice are assumed to be proportional to the difference of temperature between the surface layer T_w and the freezing point T_{freez} (which depends on the sea surface salinity), and the friction velocity at the ice-ocean interface u_{io}^* [24]:

$$F_w^{ice} = \rho_0 c_w c_h u_{io}^* (T_w - T_{freez}) \qquad (17)$$

where c_w is the seawater specific heat, c_h is a heat transfer coefficient.
The upper ocean layer salinity changes linked to surface freshwater fluxes are due to snow melting and sea ice growth or melt. The freshwater fluxes between ice and ocean are represented as

$$F_{salt} = S_w \left(\frac{\partial m_{snow}}{\partial t}\right) + (S_w - S_{ice}) \left(\frac{\partial m_{ice}}{\partial t}\right) \qquad (18)$$

where S_w is a reference seawater salinity, S_{ice} is the sea ice salinity, m_{snow} and m_{ice} are the snow and ice mass per unit area, respectively. Constant salinities are assumed for ice ($\sim$ 4-6 psu) and snow (0 psu).

7. ICE-ATMOSPHERE FORCING

The computation of sea ice surface boundary conditions is based on bulk formulae expressing the fluxes of energy (Q^{ice}, W/m^2), momentum (τ_a^{ice}, N/m^2) and mass (emp^{ice}, kg/m^2/s) that are applied at the top of the snow-sea ice system, as functions of forced atmospheric fields (from reanalyses or climatologies).
Two bulk formulations are available in the NEMO code: CLIO [7] and CORE [25]. Our configuration uses the CORE formulation, which is described for both ocean and ice in the [25] report. Here, we only provide details for the ice-atmosphere coupling as they are implemented in NEMO.
The CORE formulation requires the a priori knowledge of the following atmospheric fields: wind vector ($\mathbf{U_a}$, m/s), air temperature (T_a, K), and specific humidity of dry air (q_a, kg/kg) at specific reference levels; solar and longwave (Q^{sr} and Q^{lw}, W/m^2) radiation fluxes; as well as total and solid precipitation (P^t and P^{solid}, kg/m^2/s).

The atmosphere-ice momentum flux τ_a^{ice}, or stress, is:

$$\tau_a^{ice} = \rho_a C_D \Delta\mathbf{U} |\Delta U| \qquad (19)$$

where $\rho_a = 1.22$ kg/m^3 is the air density, C_D is the ice-atmosphere drag coefficient and $\Delta\mathbf{U}$ is the difference between $\mathbf{U_a}$ and the ice drift velocity.

The atmosphere-ice heat flux Q^{ice} is decomposed into solar (Q_{sr}^{ice}) and non-solar (Q_{ns}^{ice}) components:

$$Q^{ice} = Q_{sr}^{ice} + Q_{ns}^{ice}(T_{su}^{ice}) \qquad (20)$$

The use of an iterative Newton procedure to compute the sea ice surface temperature T_{su}^{ice}, requiring the derivative $\partial Q^{ice}/\partial T_{su}^{ice}$, explains such a decomposition.

The net solar flux is:

$$Q_{sr}^{ice} = Q_{sr}(1 - \alpha^{ice}) \qquad (21)$$

where α^{ice} is computed with the parameterization of [1], i.e., as a function of snow depth, ice thickness and surface temperature. In principle, the surface albedo also depends on cloud fraction. This is because the albedo depends on wavelength, and that cloud fraction changes the spectral distribution of the incoming light.

Consequently, the overcast albedo is about 0.06 higher than the clear-sky value. In the present version of the NEMO surface module, the clear-sky albedo value is used because CORE atmospheric fields do not include cloud fraction. This would imply a systematic low albedo bias since in polar regions, skies are overcast during a large part of the year. This should be improved in the next NEMO release.

The net non-solar flux and its derivative with respect to the sea ice surface temperature are the sum of the longwave, sensible and latent heat components:

$$Q_{ns}^{ice} = Q_{lw}^{ice} - Q_{sh}^{ice} - Q_{lh}^{ice} \tag{22}$$

The net long-wave flux and its temperature derivative are

$$Q_{lw}^{ice} = \epsilon(Q_{lw} - \sigma {T_{su}^{ice}}^4) \tag{23}$$
$$\frac{\partial Q_{lw}^{ice}}{\partial T_{su}^{ice}} = -4\epsilon\sigma {T_{su}^{ice}}^3 \tag{24}$$

where σ is the Stefan-Boltzmann constant and $\epsilon = 0.95$ is the surface emissivity. The sensible heat flux and its derivative are

$$Q_{sh}^{ice} = \rho_a c_a C_H |\mathbf{U}|(T_{su}^{ice} - T_a) \tag{25}$$
$$\frac{\partial Q_{sh}^{ice}}{\partial T_{su}^{ice}} = \rho_a c_a C_H |\mathbf{U}| \tag{26}$$

where $c_a = 1000.5$ J/kg/K is the specific heat of the air and C_H is the heat transfer coefficient. The latent heat flux (and its derivative), associated with condensation and sublimation, is

$$Q_{lh}^{ice} = \rho_a L_s C_E |\mathbf{U}|[q_{sat}(T_{su}^{ice}) - q_a] \tag{27}$$
$$\frac{\partial Q_{lh}^{ice}}{\partial T_{su}^{ice}} = \rho_a L_s C_E |\mathbf{U}| \frac{\partial q_{sat}}{T_{su}^{ice}} \tag{28}$$

where $L_s = 2.839 \times 10^6$ J/kg is the latent heat of sublimation, C_E is the water vapour transfer coefficient. The ice surface is assumed to be at water vapour saturation, which explains the use of $q_{sat}(T_{su}^{ice})$ is used. By default, the sensible heat flux is capped to zero, allowing only negative latent heat fluxes and sublimation (e.g., condensation is precluded). The CORE formulation assumes a simple expression for the saturating specific humidity with respect to ice (kg/kg), warranting an easy differentiation:

$$q_{sat} = \rho_a^{-1} q_1 e^{(q_2/T_{su}^{ice})} \tag{29}$$
$$\frac{\partial q_{sat}}{T_{su}^{ice}} = -\frac{q_1 q_2 e^{(q_2/T_{su}^{ice})}}{\rho_a {T_{su}^{ice}}^2} \tag{30}$$

with $q_1 = 11.6378 \times 10^6$ kg/m^3, $q_2 = -5897.8$ K. By default, the heat, water vapour and drag coefficients are assumed equal: $C_H = C_D = C_E = 1.63 \times 10^{-3}$.

The water vapour flux is

$$emp^{ice} = E - P^{solid} \tag{31}$$

Evaporation or more precisely sublimation E corresponds to Q_{lh}/L_s. Only solid precipitation contributes to snow accumulation. Liquid precipitation over sea ice is the difference between total and solid, and is directly sent to the ocean.

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