Snow-on-Sea-Ice Model for CICE and MPAS-CICE
Requirements and Design

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Contents

Chapter 1

Summary

Once deposited, the character and distribution of snow on sea ice depend on re-transport (wind), melting/wetting, and metamorphism (chiefly producing low conductivity depth hoar or snow-ice). Each of these processes affects the other, and they are crucial for the evolution of the sea ice pack. These processes likely are critical for biological and chemical cycling in the snow pack and the ice below.

For additional background, see Sturm et al (2002), particularly section 5, "Representing the Snow Cover in Ocean-Ice-Atmosphere Models". Although we will not attempt all of these, they suggest the following:

1. Temporal evolution depends critically on a few discrete storm events. Therefore synoptic weather analyses should be used.

2. Wind slab and depth hoar resist densification. "Assuming an initial density for each snow layer, perhaps based on wind speed or precipitation rates, and then keeping the density of that layer constant through the remainder of the winter, may be the most accurate way to represent the snow cover evolution."

3. Depth hoar forms reliably, rapidly and continuously on sea ice, and is an excellent insulator, resulting in low bulk thermal conductivity values. High-density layers like wind slabs resist metamorphosing into depth hoar (Akitaya 1974). Use modeled temperature gradients to evolve depth hoar unless the layer density is high.

4. Wind slab may prevent snow from drifting after deposition.

5. Snow depth and its standard deviation, snow-water equivalent, the number of snow layers and the amount of wind slab and depth hoar

all can be tied to ice class (thin, deformed, undeformed consisting of refrozen melt ponds or hummocky ice).

6. Snow drifts around ridges cover only 6% of the ice surface area and are about 30% deeper than other snow-covered areas, but they prevent seawater filled cracks around the ridges from freezing, with important biological consequences.

7. Snow depth can be treated as a normally distributed random variable with the mean and standard deviation set by ice type. Lateral variability (20-m length scale) is closely associated with melt pond features. Snow over melt ponds is deeper and forms earlier than elsewhere.

Our current sea ice model (CICE) includes a basic snow formulation describing the essential effects of snow on sea ice, such as its albedo, vertical conduction, and growth/melt processes. It also incorporates more detailed processes such as snow-ice formation due to flooding and snow infiltration by melt water, which may form melt ponds. Several potentially important processes are not included in the model, such as compaction and redistribution of snow by wind and their effects on the thermal balance and on effective roughness (form drag). Snow metamorphism due to temperature gradients and liquid water content also are not included in the current model. The model is discretized to run with multiple vertical layers; however, to date it has been run using only 1 layer.

While this work will not result in detailed descriptions of all of these processes within the model, we will test and tune the model to run with multiple snow layers and reanalyzed synoptic-scale precipitation data, and implement parameterizations to represent the following processes:

- Radiative effect of snow redistribution by wind with respect to ice topography

  Snow can be scoured from level ice, blowing into leads and ponds, or piling up on ridges.

- Radiative and conductive effects of snow grain metamorphism (variable grain size)

  The presence of liquid water in snow, such as rain or melt water, changes the surface albedo dramatically. It also alters the conductivity of the snow pack. These effects are associated mainly with the formation of depth hoar (change in grain size).

- Coupling effects of including fresh water and heat associated with snow saturation

> Except for the topo melt pond scheme, melt water and heat in ponds (which may be hidden within a partially saturated snow pack) are “virtual” in the sense that they are provided to the ocean model component immediately upon melting, even though the effects of the liquid water continue to be tracked as if it were retained on the ice. Retaining that water and heat in the sea ice component will alter the timing, location and magnitude of fresh water runoff events into the ocean.

Chapter 2

Requirements

Except for the first, the requirements listed below are outlined in terms of physical processes, most of which can be described within the column package common to both CICE and MPAS-CICE. These processes likely will require the addition of new tracers, however, which must be implemented separately in CICE and MPAS-CICE.

2.1 Requirement: Simulations can be run using more than one snow layer

Date last modified: 2015/02/13
Contributors: Elizabeth Hunke

CICE is already discretized to utilize more than one snow layer. However, newer parameterizations in the model have not been tested, nor has the model sensitivity to snow layer resolution been tested, using more than one snow layer. An optimal number of layers will be chosen and some parameter tuning may be required to roughly match simulations using one layer. Forcing data will include both "normal year" and synoptic precipitation.

2.2 Requirement: Radiative effects of snow redistribution by wind

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Contributors: Elizabeth Hunke

The current model assumes that the snow depth is uniform across each ice thickness category within a grid cell for the vertical thermodynamic calculation. However, there are separate radiation calculations for bare ice, snow-covered ice, and pond-covered ice; snow and ponds interact through snow saturation levels. Redistributing the snow will alter these radiative calculations.

2.3 Requirement: Tracers for ice and liquid water mass in snow

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Contributors: Elizabeth Hunke

Tracers for the mass of ice and liquid water in snow will be implemented, as tracers on snow volume (layers). They will be used for the snow metamorphism parameterizations, and together with snow volume, they can be used to determine effective snow density.

2.4 Requirement: Radiative effects of snow metamorphism

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Contributors: Nicole Jeffery

Snow grain radii is used to compute snow inherent optical properties in the Delta-Eddington radiative scheme. The scheme currently has tables which account for grain radii between 0.005 and 2.5 mm. The requirements of two metamorphic processes, dry and wet metamorphism, are described in the next subsections.

2.4.1 Dry snow metamorphism

The model currently does not account for dry snow kinetic metamorphism (TG metamorphism) which, in the formation of depth hoar, increases the snow grain radius. Dynamic snow effective radius will be included as a snow volume tracer for radiative calculations and will evolve analytically as a function of snow temperature, temperature gradient, and density.

2.4.2 Wet snow metamorphism

Wet metamorphism increases snow grain radius which alters snow radiative properties. The current model has several parameters to account for the relationship between grain size and optical properties. The "nonmelt" snow grain radius can be tuned at run time (within a fixed range is 0.2 mm and a specified maximum; cice.v5 default is 0.875 with R-snw = 1.5). There is a linear temperature dependent transition between non-melt and melt (also specified at run time, rsnw-mlt = 1.5 mm) snow grain radii.

An improved version will use the liquid water mass tracer and effective snow grain radius (already needed for TG metamorphism). Wet metamorphism changes both density (through a volume change) and effective grain size, though we only consider changes in grain radius. Both rain and snowmelt provide sources of snow liquid content. Loss terms are modelled as runoff to meltponds and the ocean.

2.5 Requirement: Physical atmosphere-ice-ocean coupling of fresh water and heat associated with liquid water in snow, including melt ponds

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Code users will be able to choose whether heat and fresh water associated with liquid water in snow (or ponds) is held according to physical processes in the model or immediately fluxed to the ocean. The model will conserve both heat and water in either case.

Chapter 3

Algorithmic Formulations

3.1 Design Solution: Simulations can be run using more than one snow layer

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No algorithmic changes are anticipated, although parameter tuning may be necessary. Performance should be evaluated with multiple layers, particularly as additional tracers are added.

3.2 Design Solution: Radiative effects of snow redistribution by wind

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Contributors: Elizabeth Hunke

Because the thermodynamic schemes in CICE assume a uniform snow depth over each category, ignoring the fractions of level and deformed ice, effects of snow redistribution will be included only via the delta-Eddington radiation scheme. The redistributed snow depth will be used to determine the effective area of bare ice (for very small snow depths) and the effective area and depth of melt ponds over level ice. Once those areas are determined, the redistributed snow volume over them will be known, from which the snow depth for the remaining snow-covered area can be computed and used for its radiation balance calculation.

Level and ridged ice area are tracked using a level-ice area tracer, a_{lvl}, with the ridged ice area diagnosed using $a_{rdg} = 1 - a_{lvl}$. Thus, if the area of ice in a grid cell is a_i, then the the area of level ice in that grid cell is $a_{lvl}a_i$.

Two basic approaches will be tested to determine the sensitivity of the sea ice simulation to snow redistribution by wind. For both, nonlocal redistribution of snow (i.e., between grid cells) is neglected, assuming that the difference between snow mass blowing into a grid cell and that blowing out is negligible. A third approach represents a small extension of the second one, to utilize the level-ice tracer. Finally, the second approach may be extended further to allow more snow to be "caught" by melt ponds.

1) **30% rule.** Sturm et al (2002) noted that on average during the SHEBA experiment, ridged ice carried 30% deeper snow than did undeformed ice. Using this rule of thumb, we can reduce the amount of snow on level ice in the model by reducing the snowfall rate over the sea ice and assuming the removed snow volume passes into the ocean through leads, instantaneously. This approach takes into account the area of open water available, as in the original code, by employing a precipitation flux in units of kg m^{-2} s^{-1}, which accumulates snow only on the ice-covered area of the grid cell.

There are two levels of sophistication at which this approach can be accomplished, to tease out the sensitivities:

1a) The snow on sea ice continues to be evenly distributed over deformed and undeformed ice, so that the only effect is from the reduced precipitation amount (assumed to blow into leads).

Here,

$$h_{lvl} = h_{rdg} = h'_{rdg} = (1+p)h'_{lvl},$$

where $p = 0.3$ and primed quantities represent modified values after snowfall is reduced. The snow volume reduction over the grid cell area A is

$$\Delta V = v_{lvl} - v'_{lvl} = A a_{lvl} a_i h_{lvl} \left(\frac{p}{1+p}\right).$$

Therefore the modified snowfall volume deposited on the ice is

$$A a_i f'_s \Delta t = A a_{lvl} a_i f_s \Delta t \left(\frac{p}{1+p}\right),$$

or

$$f'_s = f_s a_{lvl} \left(\frac{p}{1+p}\right).$$

1b) The effective depth of snow on ridged and level ice differs by 30% for the radiation calculation.

Here,

$$\begin{aligned} h_{rdg} &= h_{lvl} \\ V'_{lvl} &= V_{lvl} - \Delta V \\ V'_{rdg} &= V_{rdg} + \Delta V \\ h'_{rdg} &= (1+p)h'_{lvl}. \end{aligned}$$

Therefore

$$\begin{aligned} V'_{rdg} &= Aa_{rdg}a_i h'_{rdg} \\ &= Aa_{rdg}a_i(1+p)h'_{lvl} \\ &= Aa_{rdg}a_i h_{rdg} + \Delta V \\ &= Aa_{rdg}a_i h_{lvl} + \Delta V \end{aligned}$$

or

$$Aa_{rdg}a_i(1+p)h'_{lvl} = Aa_{rdg}a_i h_{lvl} + \Delta V.$$

Also

$$V'_{lvl} = Aa_{lvl}a_i h'_{lvl} = Aa_{lvl}a_i h_{lvl} - \Delta V.$$

Set $b = Aa_{lvl}a_i$ and $c = Aa_{rdg}a_i$ and solve for the effective depths over level and ridged ice, h'_{lvl} and h'_{rdg}:

$$\begin{aligned} h'_{lvl} &= \frac{b+c}{c(1+p)+b}h_{lvl} = \frac{1}{1+p(1-a_{lvl})}h_{lvl} \\ h'_{rdg} &= \frac{b+c}{c(1+p)+b}(1+p)h_{lvl} = \frac{1+p}{1+p(1-a_{lvl})}h_{lvl}. \end{aligned}$$

1c) These two tests may be run simultaneously, i.e., reducing the snowfall amount and also changing the effective snow depth for the radiation calculation.

2) **LIM.** This formulation was designed and implemented by Olivier Lecomte in the Louvain Ice Model (LIM). It does not incorporate level or ridged ice tracers, instead using the standard deviation of the computed ice thickness distribution.

Following Lecomte (2015), we will parameterize the amount of snow lost into the ocean through leads or redistributed to other thickness categories by defining the redistribution function Φ for snow *mass* as the sum of an erosion rate Φ_E and a redeposition rate Φ_R for each category of thickness h_i:

$$\Phi_E = \frac{\partial m}{\partial t}_{erosion} = -\frac{\gamma}{\sigma_{ITD}}(V - V^*)\frac{\rho_{\max} - \rho_s}{\rho_{\max}}$$

$$\Phi_R = \frac{\partial m}{\partial t}_{redeposition} = m_b g(h_i)\,(1-f)$$

where ρ_s and $\rho_{\max}$ are the (effective or "meta") snow density and the maximum snow density in the model, respectively. Erosion begins when the instantaneous wind speed V exceeds the seasonal wind speed required to compact the snow to a density ρ_s, $V^* = (\rho_s - \beta)/\alpha$. σ_{ITD} is the standard deviation of the ice thicknesses from the thickness distribution g within the grid cell. γ is a tuning coefficient for the eroded mass, which Lecomte (2015) set to 10^{-5} kg m^{-2}. m_b is the total mass of snow in suspension in the grid cell, per unit area. The fraction of this suspended snow lost in leads is

$$f = (1 - a_i)e^{\frac{\sigma_{ITD}}{\sigma_{ref}}},$$

where the scale factor $\sigma_{ref} = 1$ m and a_i is the total ice area fraction within the grid cell,

$$a_i = \int_0^\infty g(h)\ dh.$$

From Lecomte et al (2013),

$$\rho_s = 44.6V^* + 174 \text{ kg m}^{-3}$$

for seasonal mean wind speed V^*, i.e. $\alpha = 174$ kg m^{-3} and $\beta = 44.6$ kg s m^{-4}.

Note: Olivier does not mention whether he weighted the category thicknesses by the category areas when computing the standard deviation σ_{ITD}.

3) **LIM with level and ridged ice.** The LIM approach above can be extended to utilize the level-ice tracer by computing the thicknesses of level and ridged ice within each category, and using those to determine the standard deviation σ_{ITD} instead of using the mean ice thicknesses of the categories.

4) **Snow in melt ponds.** A fraction similar to that lost in leads may be caught within the area of melt ponds:

$$f_p = a_p e^{\frac{\sigma_{ITD}}{\sigma_{ref}}},$$

where a_p is the pond area. If the ponds are frozen, the snow blown into them will accumulate on top of the pond ice; otherwise the new snow will contribute to any existing snow within the pond area.

3.3 Design Solution: Tracers for ice and liquid water mass in snow

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Contributors: Elizabeth Hunke, Nicole Jeffery

Implement tracers on snow volume for the mass per unit area of liquid water in snow, m_{liq}, and the mass per unit area of ice in snow, m_{ice}. Then the liquid mass fraction in each layer of snow is

$$\phi_{liq} = \frac{m_{liq}}{m_{liq} + m_{ice}},$$

and the effective density of snow is

$$\rho_s^{eff} = \frac{m_{liq} + m_{ice}}{h_s},$$

where h_s is the snow volume per unit area (i.e., snow depth) in the layer. If needed, the total snow energy can be partitioned based on the liquid and solid fractions, accounting for the latent heat of fusion. Note that the snow water equivalent is $m_{liq} + m_{ice}$ (kg/m^2).

Sources of m_{ice} are snowfall, condensation, and freezing of liquid water within the snowpack; sinks are sublimation and melting. All of the sources and sinks of m_{ice} are already computed in the code except for freezing of liquid water within the snow pack.

Sources of m_{liq} are rain and snow melt; freezing of liquid water within the snowpack and runoff are sinks. Runoff and meltwater entering a snow layer (i.e., runoff from the layer above) are associated with vertical flow through the snow column. Rain and snow melt are already available in the code (although internal snow melt may need to be assigned a variable).

As in CLM, when the liquid water within a snow layer exceeds the layer's holding capacity, the excess water is added to the underlying layer, limited by the effective porosity $(1 - \theta_{ice})$ of the layer. The flow of water is assumed to be zero if the effective porosity of either of the two layers is less than 0.05, the water impermeable volumetric water content. The downward flow between layers is

$$w = \left[\frac{\theta_{liq} - S_r \left(1 - \theta_{ice}\right)}{\Delta t} \right] \rho_{liq} \Delta z \geq 0$$

where the volumetric liquid θ_{liq} and ice θ_{ice} contents are

$$\begin{aligned}\theta_{ice} &= \frac{m_{ice}}{\rho_{ice}\Delta z} \leq 1 \\ \theta_{liq} &= \frac{m_{liq}}{\rho_{liq}\Delta z} \leq 1-\theta_{ice}\end{aligned}$$

and $S_r = 0.033$ is the irreducible water saturation due to capillary retention after drainage has ceased (Anderson 1976). In CLM, $\rho_{ice} = 917$ and $\rho_{liq} = 1000$ kg/m^3. Excess water will be supplied to the melt pond parameterization, which puts a fraction of it into the pond volume and allows the rest to run off into the ocean.

Liquid water freezing, which converts m_{liq} to m_{ice}, will need to be computed based on thermal constraints in the snow. (HOW?)

First compute $m_{ice}^{t+\Delta t}$ for each snow layer i (snow-ice formation, mechanics, and advection are computed elsewhere and not included here)

$$m_{ice}^{t+\Delta t}|_i = m_{ice}^{t}|_i + \delta_{i1}(P_{recip} - S_{ub})\Delta t + (\Delta m_{ice})_p \tag{3.1}$$

where P_{recip} $(kg\, m^{-2}\, s^{-1})$ is precipitation, S_{ub} is sublimation, and $(\Delta m_{ice})_p$ represents a phase change.

Then compute the liquid mass change in two steps: 1) compute the change due to rainfall, R_{ain} $(kg\, m^{-2}\, s^{-1})$, and melt $(\Delta m_{liq})_p$

$$m'_{liq}|_i = m_{liq}^{t}|_i + \delta_{i1} R_{ain}\Delta t + (\Delta m_{liq})_p \tag{3.2}$$

Then, calculate w_1 using m'_{liq} and $m_{ice}^{t+\Delta t}$ to find the liquid above irreducible water saturation that flows to the next level. For $w_1 > 0$

$$m_{liq}^{t+\Delta t}|_1 = m'_{liq}|_1 - w_1\Delta t \tag{3.3}$$

For the interior snow layers, add the contribution of liquid water from above:

$$m''_{liq}|_i = m'_{liq}|_i + w_{i-1}\Delta t \tag{3.4}$$

Then find the loss w_i using $m''_{liq}|_i$ and $m_{ice}^{t+\Delta t}$.

$$m_{liq}^{t+\Delta t}|_i = m''_{liq}|_i - w_i\Delta t \tag{3.5}$$

The loss $w_{N_s}\Delta t$ from the bottom layer flows to the meltponds or oceans.

Save the snow mass fractions of precipitation, refrozen ice and old ice for the snow grain radius metamorphism.

3.4 Design Solution: Radiative effects of snow metamorphism

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Contributors: Nicole Jeffery, Elizabeth Hunke

The tracers m_{liq} and m_{ice} characterize the snow in a vertical snow layer, for each ice category and horizontal grid cell. Meltpond liquid covers a fraction of the grid cell and represents liquid in excess of m_{liq}. The radiative effects of snow grain radius in the fraction of ice with meltponds is only applied when the meltpond liquid has not yet saturated the snow pack. Otherwise, Delta-Eddington transfer uses meltpond properties. Therefore, modelled changes in snow grain radii from metamorphism are designed specifically for the fraction without meltponds.

At each time step, determine the fractions of snow mass that is old (f_{old}), new (f_{new}, freshly fallen) or refrozen (f_{rfrz}, m_{liq} to m_{ice} phase change). Here, $f_{old} + f_{new} + f_{rfrz} = 1$. Then, following CLM, the new snow grain radius at time t is computed as a weighted function of snow grain radii:

$$r(t) = [r(t-1) + \Delta r_{wet} + \Delta r_{dry}]\, f_{old} + r_{new} f_{new} + r_{rfrz} f_{rfrz},$$

where

$$\begin{aligned} r_{new} &= 5.45 \times 10^{-5} \text{ m} \\ r_{rfrz} &= 10^{-3} \text{ m} \\ \Delta r_{wet} &= \frac{4.22 \times 10^{-5}}{4\pi r^2} f_{liq}^3 \Delta t \\ \Delta r_{dry} &= \nu \left(\frac{\eta}{r - r_{new} + \eta} \right)^{\frac{1}{\kappa}} \Delta t \end{aligned}$$

where f_{liq} is the liquid mass fraction and the parameters ν, η, and κ are obtained from a look-up table that depends on snow temperature, temperature gradient and (effective) density.

The only piece remaining to calculate is f_{rfrz}.

3.5 Design Solution: Physical atmosphere-ice-ocean coupling of fresh water and heat associated with liquid water in snow, including melt ponds

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Contributors: Elizabeth Hunke, Nicole Jeffery

The volume of water and amount of heat associated with liquid water in snow, including melt ponds, will be tracked separately from the current fresh water and heat variables, to allow their flux to the ocean model component to be delayed based on the model simulation. Additionally, the fresh water flux can be used for "physical" coupling, as opposed to "virtual" coupling in which the ocean volume remains constant. There needs to be some care in determining whether a coupled configuration assumes that the snow has any energy content, and an appropriate option available in the code.

Chapter 4

Design and Implementation

4.1 Implementation: Simulations can be run using more than one snow layer

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Contributors: Elizabeth Hunke

Based on the "column_pkg" branch, create a "snow" branch in the CICE repository which will be used for snow model development. Provide precipitation data and forcing subroutines to read synoptically varying data.

Set `nslyr` to an integer value greater than 1.

4.2 Implementation: Radiative effects of snow redistribution by wind

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Contributors: Elizabeth Hunke

1) Define a parameter $p = 0.3$ (`snwlvlfac`). This parameter could be put in namelist but I do not expect this to become a standard parameterization in the model.

1a) In the **ice_step_mod.F90** module, just before the call to *thermo_vertical*, reduce the precipitation flux f_s as

$$f_s' = f_s a_{lvl} \left(\frac{p}{1+p} \right),$$

where a_{lvl} is the level-ice area tracer value (so $a_{lvl}a_i$ is the level ice area fraction of the grid cell). The factor of a_i is included implicitly, as the snowfall flux is applied only over the ice area.

1b) In the shortwave module for level-ice ponds, create a new variable (`hsnlvl`) for snow depth over the level ice.

$$h'_{lvl} = \frac{b+c}{c(1+p)+b} h_{lvl} = \frac{1}{1+p(1-a_{lvl})} h_{lvl}.$$

Replace `hsn` with `hsnlvl` for the snow infiltration calculation and for the calculation of snow depth over refrozen melt ponds. This test does not require implementation of a snow-on-level-ice tracer.

2) Add a new module for snow-specific calculations (**ice_snow.F90**). It can also be used later for a snow-on-level-ice tracer. In a new snow redistribution subroutine (*snow_redistr*), define the critical seasonal wind speed (`Vseas`) as $V^* = (\rho_s - 44.6)/174$ m/s, compute the weighted standard deviation of the ice thickness distribution (`ITDsd`) over $N =$ `ncat` categories,

$$\sigma_{ITD} = \sqrt{\sum_{n=1}^{N} a_{in} \left(h_{in} - \sum_{k=1}^{N} a_{ik} h_{ik} \right)^2},$$

and assume $\sigma_{ref} = 1$.

Set $\gamma = 10^{-5}$ kg m^{-2}, $\rho_{\max} = 450$ kg m^{-3} (based on Figure 3.2 in Olivier's thesis), and set the fraction f of snow mass lost to leads,

$$f = (1 - a_i) e^{-\frac{\sigma_{ITD}}{\sigma_{ref}}}.$$

Note that this fraction will need to be added to the fresh water flux for strict conservation.

Compute the total mass of snow in suspension (per unit area, `snw_susp`),

$$m_b = \sum_{n=1}^{N} \Delta m_n^{erosion} = \min \left(\frac{\gamma \Delta t}{\sigma_{ITD}} \max(V - V^*, 0) \frac{\rho_{\max} - \rho_s}{\rho_{\max}}, \sum_{n=1}^{N} a_{in} h_{sn} \rho_s \right)$$

The mass erosion rate is equal for all categories unless the maximum value (total snow mass available) is reached. Combining the mass erosion and redeposition rates and converting them to thickness changes, we have

$$\Delta h_{sn} = \frac{1}{a_{in}\rho_s} \left[-\min \left(\frac{\gamma \Delta t}{\sigma_{ITD}} \max(V - V^*, 0) \frac{\rho_{\max} - \rho_s}{\rho_{\max}}, h_{sn}\rho_s \right) + m_b a_{in} (1-f) \Delta t \right]$$

Update the snow state variable `vsnon` using $v_{sn} = v_{sn} + a_{in}\Delta h_{sn}$ (or $h_{sn} = h_{sn} + \Delta h_{sn}$ if thickness is the local state variable).

3) In (2), compute the standard deviation using the level and ridged ice thicknesses, separately:

$$\sigma_{ITD} = \sqrt{\sum_{n=1}^{N} a_{in}a_{lvln}\left(h_{ilvln} - \sum_{k=1}^{N} a_{ik}h_{ik}\right)^2 + a_{in}a_{rdgn}\left(h_{irdgn} - \sum_{k=1}^{N} a_{ik}h_{ik}\right)^2}.$$

4) In this experiment, some of the snow that would have blown into leads is caught in melt ponds instead. In (2), replace f with

$$(1 - a_i - a_{pnd}a_{lvl}a_i)e^{-\frac{\sigma_{ITD}}{\sigma_{ref}}}.$$

The extra snow fraction should be deposited only over the level ice, and for purposes of the radiation calculation, wholly over ponds. To do this properly, the mean snow depth over level ice should be carried as a tracer on level ice, from which the snow depth over ponds can be backed out for the next time step, using the pond area. The state variable `vsnon` will continue to represent the snow volume per unit area over both deformed and undeformed ice. That is, if h_{slvl} is tracked as a snow thickness tracer on level-ice area, then (suppressing category subscripts) all quantities in

$$h_s = \frac{v_s}{a_i} = h_{slvl}(1 - a_{pnd})a_{lvl} + h_{spnd}a_{pnd}a_{lvl} + h_s a_{rdg}$$

are known except h_{spnd}. Once computed, h_{spnd} and the tracer h_{slvl} can be used in the radiation calculation to determine the effective area fractions of snow-covered, ponded and bare ice. (Note: The shortwave calculation for level-ice ponds already tracks the difference in snow depth between unponded ice and refrozen pond ice, due to temporal delays in snow accumulation over ponds. This difference can also be applied to h_{spnd}.)

The snow-on-level-ice tracer will be incremented or decremented following changes to the snow volume `vsnon`, and additionally incremented by the increase in snow depth over all level ice (ponded and unponded) associated with pond-capturing in each category:

$$\Delta h_{sn} = \frac{(m_b a_{in})(a_{pndn}a_{lvln}a_{in})\Delta t}{\rho_s} e^{-\frac{\sigma_{ITD}}{\sigma_{ref}}}.$$

Here, $m_b a_{in}$ is the fraction of the total mass in suspension that is available to each category n, and $a_{pndn}a_{lvln}a_{in}$ is the fraction of the grid cell covered by ponds that "captures" that portion of the snow which would have blown into leads.

4.3 Implementation: Tracers for ice and liquid water mass in snow

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Contributors: Elizabeth Hunke

Tracers m_{ice} (`smice`) and m_{liq} (`smliq`) will be implemented following standard procedures in CICE for adding tracers, to include necessary initialization and restart capability. They will be incremented or decremented during the thermodynamic calculations for each process, to match changes in snow depth. These changes to the mass tracers may all be done at once except for changes associated with snowfall, for which the density may be different. Vertical flow of liquid water through snow will be handled in a separate subroutine in **ice_snow.F90**.

The order of operations here would be
1) Save the mass of new snowfall.
2) before adjusting the tracers to 2) Update m_{ice} and m_{liq} based on existing thermodynamic processes.
3) Compute freezing of liquid water in snow and save its mass.
4) Compute θ_{ice} throughout the snow and θ_{liq} in the top snow layer only.
5) Compute the liquid water excess in the top layer $w\Delta t \geq 0$, add to $m_{liq}|_2$, and subtract from $m_{liq}|_1$.
6) Use the updated $m_{liq}|_2$ to compute θ_{liq} subject to the constraint $\theta_{liq} \leq 1 - \theta_{ice}$ and repeat the procedure for the remaining snow layers.
7) Provide excess water left over after this process to the melt pond parameterization (either to increase pond volume or run off into the ocean). Note: the snow saturation parameterization in the shortwave module will need to be reviewed and possibly changed.
8) Verify that $m_{liq} \leq S_r m_{ice} \rho_{liq}/\rho_{ice}$ in each layer.
9) Compute and save the fractions f_{old} and f_{new} (later save f_{rfrz}) in the updated snow column.

4.4 Implementation: Radiative effects of snow metamorphism

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Contributors: Nicole Jeffery, Elizabeth Hunke

Implement tracer r (`rsnw`) following standard procedures in CICE for adding tracers.

After the thermodynamic calculations and before the radiation calculation at the end of the time step: Given updated m_{ice} and m_{liq}, compute ϕ_{liq} or the mass liquid fraction (see Appendix). Using snow temperature and temperature gradient, obtain ν, η and κ from the look-up table. (In the future, compute and use the effective density ρ_s^{eff}, also using m_{ice} and m_{liq}.). Compute the wet and dry changes to snow radius, and using the updated fractions f_{old}, f_{new} and f_{rfrz}, compute the new snow grain radius for use in the Delta-Eddington radiation parameterization on the Delta-Eddington grid. Use a flag option `tr_rsnw` to define `rsnwn` for category n used in *shortwave_dEdd*. If true, set `rsnwn` equal to the dynamic $r_j(n)$, otherwise use the value determined in *shortwave_dEdd_set_snow*.

See below for specific details in defining Δr_{dry} and Δr_{wet}

4.4.1 Dry snow metamorphism

ECH note: The column code is organized a bit differently, so some of the modules referred to below are incorrect.

1) Define namelist `snow_nml` with parameter `rsnwfall` to choose an initial grain radius for falling snow (0.02 mm to 0.09 mm, Domine et al,2008). Also need a `r_tgmax` parameter for the maximum snow grain radius due to TG metamorphism. Other new snow parameters can go here.
2) In **ice_colpkg_tracers.F90**, add integer `nt_rsnw` for volume-weighted effective snow grain radius. Define a flag `tr_rsnw` to indicate if using metamorphic snow grain radius. Add to namelist `tracer_nml` along with flag `restart_rsnw`.
3) Modify **ice_init.F90** in *input_data* and *init_state* to initialize `nt_rsnw` and define tracer dependence as snow volume weighted (`trcr_depend` of 2). Initialize grain size in *set_state_var* as `rsnwfall`.
4) Define *restart_snow* in (**ice_snow.F90**) for snow grain tracer.

5) In **ice_step_mod.F90** module, after the call to *thermo_vertical* and before *update_aerosol* call a new subroutine *snow_dry_metamorph* in **ice_snow.F90** for grid cells with cold snow.
6) Define parameter table for dry metamorphism.
7) Compute the snow temperature gradient. Because of low vertical resolution, we try an approach similar to Flanner et al. (2006). For `nslyr`$= 1$

$$\frac{\Delta Tsn}{\Delta z} = \left| \frac{(Tsn_1 hsn + Tin_1 \Delta hin_1)/(hsn + \Delta hin_1) - Tsf}{hsn} \right| \quad (4.1)$$

where Tin_1 is the upper grid level ice temperature, Δhin_1 and is the upper ice grid spacing. Similarly for `nslyr`> 1, the top layer gradient will use Tsf and the bottom layer will average Tin_1 and Tsn_{nslyr}.
8) Compute Δr_{dry} using T, $\Delta T/\Delta z$ and table.

4.4.2 Wet snow metamorphism

Once the liquid and ice mass tracers are updated for each snow layer, the change in grain radius from wet metamorphism (Δr_{wet}) is straight forward from equation 3.6.

4.5 Implementation: Physical atmosphere-ice-ocean coupling of fresh water and heat associated with liquid water in snow, including melt ponds

Date last modified: 2015/02/24
Contributors: Elizabeth Hunke, Nicole Jeffery

The volume of fresh water held in meltponds is already tracked for the topo pond parameterization, and optionally combined with the freshwater coupling variable `fresh`. Its functionality will be reused for the level-ice pond scheme and for other liquid water held in snow.

Snow energy content is already tracked in the model. Heat fluxes associated with changes in snow energy will also be provided as a separate variable that can be combined with the ocean heat flux variable `fhocn` prior to coupling. Currently, there is no atmosphere-ice heat flux coupling variable associated with precipitation in CESM, and therefore for coupling purposes, the snow heat content is assumed to be zero. For now, all heat fluxes associated with precipitation will be assumed to move between sea ice and ocean, even if they are fundamentally atmospheric fluxes. (We already do this elsewhere in the model.)

Chapter 5

Testing

Runs reaching the satellite era will be validated by comparing standard diagnostics (ice extent, area, thickness, age) with observational data. Further diagnostics will be compared as appropriate: albedo, ice motion/deformation, ice-ocean fluxes. Tests within the projects 5.2–5.5 will be compared with the multilayer control produced in 5.1.

5.1 Testing and Validation: Simulations can be run using more than one snow layer

Date last modified: 2019/06/06
Contributors: Elizabeth Hunke and Nicole Jeffery

1. Standard control: 1958–2009, 1 snow layer, snow and rain precip distinguished (JRA-55) ocean-ice configuration - complete

2. Synoptic precip: 1958–1962, 1 snow layer, synoptic precip data, otherwise standard configuration.

3. Multilayer testing: 1958–1998, various snow layers (3, 5, 7), determine convergence. synoptic configuration + tuning. Determine convergence properties and choose optimal number of layers based on convergence and computational efficiency. Tune parameters as needed.

4. Multilayer control: 1958–2009, optimal number of layers, synoptic configuration.

5.2 Testing and Validation: Radiative effects of snow redistribution by wind

Date last modified: 2015/02/13
Contributors: Elizabeth Hunke

Each of these tests will branch from the multilayer control, 1980–2009.

1. 30% rule, reduced precipitation
2. 30% rule, radiative effect of reduced snow over level ice
3. 30% rule, reduced precipitation and radiative effect of reduced snow over level ice
4. basic LIM approach
5. compute standard deviation using level and ridged ice thicknesses
6. snow blown into meltponds

5.3 Testing and Validation: Tracers for ice and liquid water mass in snow

Date last modified: 2015/02/26
Contributors: Elizabeth Hunke

Initially, implement the tracers such that the mass of liquid water in snow is zero and the mass of ice in snow is such that the effective snow density equals the constant ("bulk") density parameter currently in use. Ensure mass conservation as the tracers evolve.

Gradually begin using the effective snow density calculated from the mass tracers in various parameterizations that use snow density. Initially this should not change the answers (significantly), since the effective snow density should equal the constant density parameter value.

5.4 Testing and Validation: Radiative effects of snow metamorphism

Date last modified: 2015/02/13
Contributors: Nicole Jeffery

A grain radii output variable will be compared with the current range of values to assess consistency. Delta-Eddington currently only has optical properties for snow radii within 0.005-2.5 mm, however larger grain radii are given the same properties as a 2.5 mm radius. Each metamorphism process will be tested separately. For dry metamorphism the implementation can be compared with the analytic expression in Appendix 9. Wet metamorphism radii will be compared with the interpolation used currently in CICE.

5.5 Testing and Validation: Physical atmosphere-ice-ocean coupling of fresh water and heat associated with liquid water in snow, including melt ponds

Date last modified: 2015/02/13
Contributors: Elizabeth Hunke, Nicole Jeffery

Ensure conservation.
Each of these tests will branch from the multilayer control, 1980–2009.

1. Fresh water: (output should be bit-for-bit the same as the control, except for the history variable `fresh`
2. Heat: output should change due to coupling in the ocean mixed layer
3. Fresh water and heat

Chapter 6

Future Work

Several processes have been reserved for future implementation.

6.1 The effects of liquid water refreezing

Refreezing of liquid water will alter the liquid water content in snow and change the rate of wet metamorphism. The term f_{rfrz} is the mass fraction of liquid water lost to refreezing, and it appears in the liquid water transport and grain metamorphism sections. To account for freezing, snow temperatures are first evaluated without phase change in the `ice_thermo_vertical` routine. Then if

$$T_i^{t+\Delta t} \quad < \quad T_f \quad \text{and } m_{liq} > 0 \tag{6.1}$$

freezing occurs. The energy deficit (H_i in W/m^2) is first determined, Then the mass increase in ice is computed from $\min(m_{liq}, |H_i \Delta t / L_f|)$ where L_f is the latent heat of fusion. The following is from CLM. The energy deficit H_i at the surface is

$$H_1 \quad = \quad Q_{in} + \frac{\partial Q}{\partial T}(T_f - T_i) + K_s \frac{(T_i - T_{i+1})}{z_i - z_{i+1}} - \frac{c_i \rho_s}{\Delta t}(T_f - T_1) \tag{6.2}$$

where c_i is the heat capacity ($J\,kg^{-1}\,K^{-1}$), Q_o is the surface heat flux (which includes the sensible, longwave, solar, and latent heat fluxes) For interior layers

$$H_j \quad = \quad I_j + K_s \frac{(T_i - T_{i+1})}{z_i - z_{i+1}} - K_s \frac{(T_{i-1} - T_i)}{z_{i-1} - z_i} - \frac{c_i \rho_s}{\Delta t}(T_f - T_1) \tag{6.3}$$

and I_j is the solar absorbed flux. If part of the energy deficit H_i is not released in freezing because there is not enough liquid, then and enthalpy adjustment is calculated for the layer Δq_i

$$\frac{\Delta q_i}{\Delta t} = H_i - \frac{L_f(m^t_{ice} - m^{t+\Delta t}_{ice})}{\Delta t} \tag{6.4}$$

ie., a decrease in temperature.

6.2 The effects of snow on form drag

6.3 The effects of variable density

All of the thermodynamic equations involving snow will need to be reviewed to determine what terms may have been neglected under the assumption of constant density. Snow density appears in many modeled processes (see Appendix); these will need to be evaluated for consistency if density is changed for some but not all processes.

6.3.1 Radiative effects

The Delta-Eddington radiation scheme is already coded for varying snow density in layers. Radiative effects of density variations can be associated with

1. snow redistribution
2. variable grain size

6.3.2 Thermal conduction effects of snow compaction

Thermal effects of density variations can be due to

1. wind

   High wind speeds compact the upper portion of a snow pack into "wind slab", a dense and more conductive medium that resists further drifting. An effective snow density will be computed based on wind speed, through which conductivity can be varied.

   Sturm et al (2002) note that once snow is deposited, its density changes very little. During deposition, the density primarily falls into one of two types, wind slab for wind velocities great than about 10 m/s, and loose snow for lighter winds. Their table 3 indicates densities for a variety of snow types. "Hard slab," deposited at $v = 13$ m/s, has a density of $\rho_s = 403$ kg/m^{-3} and "soft slab" is $\rho_s = 321$ kg/m^{-3}, deposited at $v = 10$ m/s. Linearly interpolating between these values, we have $\rho_s = 27.3v + 47.7$. Simeral (2005) measured snow density deposited at lighter wind speeds. Interpolating between two data points from that work, $\rho_s = (50, 125)$ kg/m^{-3} at $v = (4, 10)$ m/s, we have $\rho_s = 12.5v$. Lecomte (2014) estimated $\rho_s = 44.6V^* + 174$ for the seasonal wind speed V^*. These three lines are vastly different.

CLM assigns new snow density based on air temperature (C),

$$50 \leq \rho_s^{new} = 50 + 1.7(T_a + 15)^{1.5} \leq 50 + 1.7(17)^{1.5} \sim 119.15 \text{ kg/m}^3,$$

assuming the freezing temperature of snow is 0°C. We will use CLM's temperature-dependent density as the base density for new-fallen snow, and add to it the gradient associated with wind speed from Sturm et al. (2002):

$$\rho_s^{new} = 50 + \max\left[1.7(T_a + 15)^{1.5}, 1.7(17)^{1.5}\right] + 27.3v.$$

This value can be used for the initial computation of the change in dry snow radius (density dependence in the look-up table). Following the Sturm et al (2002) suggestion, density need not evolve further, other than by transport. However, we may want to include compaction effects as in CLM.

The snowfall rate determines the initial mass of newly fallen snow. Later, we can use that mass and ρ_s^{new} to determine the initial volume of snow. Snow volume will continue to be the primary state variable, on which mass and energy are carried, and from which effective density can be computed when needed.

From Jordan (1991), the thermal conductivity of snow is

$$k = k_{air} + \left(7.75 \times 10^{-5} \rho_s^{eff} + 1.105 \times 10^{-6} \rho_s^{eff\,2}\right)(k_{ice} - k_{air}).$$

In CLM, $k_{air} = 0.023$, $k_{ice} = 2.29$ and $k_{liq} = 0.6$ W/m/K.

2. overburden

   See CLM section 7.2.5

3. dry metamorphism

   See CLM section 7.2.5

4. wet metamorphism

$$C_3 = -\frac{1}{\Delta t} \max \left(0, \frac{\Delta f_{ice}}{f_{ice}} \right)$$

where Δf_{ice} is the change in snow ice fraction after melting as occurred and

$$f_{ice} = \frac{m_{ice}}{m_{ice} + m_{liq}}.$$

Density changes due to overburden, dry and wet metamorphism are applied using $\Delta z^{new} = \Delta z \left[1 + (C_1 + C_2 + C_3)\Delta t\right]$.

Chapter 7

Appendix: Snow density

Snow density appears frequently in the code, factoring into the following processes:

1. isostatic balance
   - ice_atmo.F90 (form drag)
   - ice_brine.F90 (brine tracer)
   - ice_meltpond_lvl.F90, ice_meltpond_topo.F90 (meltponds)
   - ice_therm_vertical.F90 (for snow-ice)
2. snow-ice formation
   - ice_aerosol.F90
   - ice_therm_mushy.F90
   - ice_therm_vertical.F90
3. thickness changes
   - ice_therm_vertical.F90
4. mass/freshwater
   - ice_dyn_shared.F90
   - ice_diagnostics.F90
   - ice_itd.F90
   - ice_mechred.F90
   - ice_therm_itd.F90 (lateral melting)

- ice_therm_vertical.F90

5. energy/enthalpy
   - ice_diagnostics.F90
   - ice_init.F90
   - ice_restoring.F90
   - ice_therm_bl99.F90
   - ice_therm_mushy.F90
6. temperature
   - ice_history.F90
   - ice_therm_mushy.F90
   - ice_therm_vertical.F90
7. conduction
   - ice_therm_bl99.F90
   - ice_therm_mushy.F90
8. porosity/infiltration
   - ice_meltpond_topo.F90
   - ice_shortwave.F90 (level-ice ponds)
   - ice_zbgc.F90
9. pond volume
   - ice_meltpond_cesm.F90
   - ice_meltpond_lvl.F90
   - ice_step_mod.F90 (topo ponds)
10. radiation
    - ice_shortwave.F90
11. dry snow metamorphism (to be implemented)
    - ice_snow.F90
12. negative definite transport

- ice_therm_itd.F90
- ice_transport_driver.F90 (upwind)

13. conservation check

    - ice_therm_itd.F90
    - ice_itd.F90
    - ice_mechred.F90

14. aerosol content

    - ice_history_bgc.F90

Chapter 8

Appendix: Some details about Snow Liquid Content

In general, the capacity of snow to hold liquid is quite small, and we do not have the resolution to model advection. So, we need to make some assumptions about the transport and storage of this water. The CLM approach assumes that water in excess of the snow irreducible liquid water saturation moves instantaeneously to the next level. It is also possible to derive a velocity based on the snow permeability that describes gravity flow to the next level and would allow for snow liquid contents greater the the irreducible level, thus permitting higher snow radii growth rates. Below are the details of this approach.

Some background first. Water in snow is found in three forms: irreducible (or hygroscopic) water, capillary water, and gravitational water. Irreducible water is held by grain surfaces (adsorption) against the force of gravity and does not contribute to the melt runoff unless the whole crystal is melted away. Capillary water is held by surface tension in capillary spaces around snow particles and is free to move under the influence of capillary forces. However it is not available to runoff until the snow melts or the spacing between crystals changes. The free water content includes only the water permanently held within the snowpack by adsorption and capillary action. Snowmelt and rain percolation runoff is gravitational water. Liquid water is mobile after the irreducible water content is satisfied. We will track total liquid water vomass and compute irreducible water content (see below).

Some more details that may be useful later: From observations, snow has an irreducible saturation of around 5% of pore volume or 9% mass with typical snow densities of about 0.35 g/cm^3. There are three regimes for snow: 1) wet snow in the Pendular regime (air exists in continuous paths) has saturation volumes of 3-8%; 2) Very wet snow in the Funicular regime (liquid exists in coninuous paths but air is still present) has values of 8-15%; and 3) slush or flooded snow with very little air has values $> 15\%$ (Singh and Singh, 2001).

The permeability of snow during two-phase (unsaturated) flow is related to the effective saturation (L^*)

$$L^* = \frac{(L_w - L_{wi})}{1 - L_{wi}} \tag{8.1}$$

where L_w is the liquid saturation (volume of liquid water per volume of pore space) and L_{wi} is the irreducible liquid saturation (max volume held against gravity per volume of pore spacing). Note that L_{wi} can be expressed as an empirical function of snow density (ref?):

$$L_{wi} = 0.0745\frac{\rho_s}{\rho_w} - 0.000267\left(\frac{\rho_s}{\rho_w}\right)^2 \tag{8.2}$$

When $L_w = L_{wi}$, the snowpack is "ripe" and the effective saturation is 0. When $L_w < L_{wi}$ then all liquid water is held by the snow and not free to flow. For $L^* \geq 0$, the snow permeability is

$$k_s = k_{so}(L^*)^n \tag{8.3}$$

Typical values of n range from 2 to 4. In ice, we use 3. Shimiszu (1970) relates k (m^2) to the grain size

$$k_{so} = 0.0192r^2 \exp[-6.7(1-\phi)\rho_i/\rho_w] \tag{8.4}$$

Liquid porosity is the volume of liquid water per unit volume of snowpack:

$$\phi_w = L_w\phi \tag{8.5}$$

where ϕ is the porosity of the snowpack (volume of pore space per volume of snow). With no liquid water $\phi = (\rho_i - \rho_s)/\rho_i$ (where $\rho_{air} \ll \rho_i$), then $\phi = (\rho_i - \rho_s)/(\rho_i - L_w\rho_w)$

Using Darcy's law to describe movement of liquid through snow

$$u = \frac{\rho_w k_s g}{\mu_w} \tag{8.6}$$

where μ_w is the dynamic viscosity of water ($\sim 1.19 \times 10^{-2}\, kg\, m^{-1}\, s^{-1}$) Typical values: for $r = 1.5mm$, $L_{wi} = 0.04$ and $L_w = 0.15$, $k_w = 1.02 \times 10^{-11}\, m^2$ and $u = 5.6 \times 10^{-6}\, m\, s^1$ or $2\, cm\, h^{-1}$ (DeWalle and Rango,2008). Nicolaus et al., 2009 measure velocities of $1.2\, cm\, h^{-1}$ in the melting snowpack from ISPOL (2004). If we define an effective volume $h_l a_{ice} \phi$ which represents the total space within the snow pore volume that contains gravitational liquid. Then

$$L_w - L_{wi} \quad = \quad \frac{h_l a_{ice} \phi}{h_s a_{ice} \phi} = \frac{h_l}{h_s} \tag{8.7}$$

Letting $u \approx (dh_l/dt)$ due to gravity flow,

$$\begin{aligned} \frac{d(L_w - L_{wi})}{dt} &\approx \frac{\rho_w k_{so} g}{\mu_w h_s} \frac{(L_w - L_{wi})^3}{(1 - L_{wi})^3} \\ \frac{d(L_w - L_{wi})}{(L_w - L_{wi})^3} &\approx \frac{\rho_w k_{so} g}{\mu_w h_s} \frac{dt}{(1 - L_{wi})^3} \end{aligned} \tag{8.8}$$

Then, for $L_w > L_{wi}$ the change in L_w at Δt from gravity flow alone is

$$L_w(t + \Delta t) \quad \approx \quad L_{wi}(t) + \left(\frac{1}{(L_w - L_{wi})^2} - \frac{2 \rho_w k_{so} g \Delta t}{\mu_w h_s (1 - L_{wi})^3} \right)^{-1/2} \tag{8.9}$$

or in terms of water available for meltponds and flushing/meteoric ice formation, $\Delta h_l \equiv h_l(t) - h_l(t + \Delta t)$,

$$\Delta h_l \quad \approx \quad h_l(t) - h_s \left(\frac{h_s^2}{h_l^2} - \frac{2 \rho_w k_{so} g \Delta t}{\mu_w h_s (1 - L_{wi})^3} \right)^{-1/2} \tag{8.10}$$

For multiple snow layers, we could model this using a "bucket" approach. Then thickness h_s and h_l are replaced with layer thicknesses in (8.10). First compute Δh_l for an upper layer. Upper snow layers send liquid to the next lower layer and so on until the bottom layer contributes to the ice and meltponds.

Rain also contributes to the snow LWC. Two cases apply: In case 1, rain of temperature T_r falling at a rate of r_{ain} ($m\, s^{-1}$) on a snowpack that is at the freezing point (T_m) provides a heat input H_r:

$$H_r \quad = \quad \rho_w c_w r_{ain} (T_r - T_m) \tag{8.11}$$

where c_w is the heat capacity of water ($4.19 \times 10^{-3} MJ\, kg^{-1}\, K^{-1}$). Rain cools to the freezing point, giving up sensible heat but that heat is used in melting. All the liquid contributes to LWC which is exported according to (8.10).

In case 2, rain falling on a snowpack below the freezing point:

$$H_r = [\rho_w c_w (T_r - T_m) + \rho_w \lambda_f] r_{ain} \tag{8.12}$$

The rain cools to the freezing point, giving up sensible heat, and then the water freezes, giving up latent heat (second term whereλ_f is the latent heat of fusion.) In this case, all the liquid freezes, contributing to meteoric snow-ice formation.

The fate of the exported water is modeled as follows: a fraction f_{sp} contributes to meltpond volume, a fraction f_{si} contributes to meteoric snow-ice formation, and a fraction $1 - f_{sp} - f_{si}$ runs off directly into the ocean. The direct runoff may be paramterized as a function of a_{ice} as is currently done. Weve opted here to model the fraction f_{si} as meteoric snow-ice rather than a source to the brine height tracer primarily because the mushy-layer thermodynamics is not computed on the same vertical domain as defined by the brine height which makes computing the source term challenging. Also, the fraction of liquid (either exported LWC or rain on cold snow) that is exported to the ice is fresher (or at least as fresh as) and warmer than the sea ice upper layers and will, in general, be stored as an upper ice layer of meteoric snow-ice.

Meteoric snow-ice formation is not currently modelled in CICE, but it similar to flooding snow-ice formation except that latent heat release should be included in the enthalpy conversion. We assume a minimum salinity for the snow melt (S_{min} which is already in the cice thermodynamics).

Chapter 9

Appendix: Details on Dry Metamorphism

Although the snow density does not change appreciably as temperature gradients in the snow evolve, the grain sizes do change; the snow metamorphoses into depth hoar. Along with wind packed layers, depth hoar is the most common dry snow type found on sea ice. High temperature gradients ($> 20\,^oC\,m^{-1}$ Domine et al. 2008) generate strong water vapor fluxes which induce rapidly growing crystals. This is called TG metamorphism. Isothermal growth operates on fresh snow and can dominate crystal growth. However, it is a short timescale (1 day) phenomena generally associated with weak temperature gradients. For simplicity, we will not include this process.

The CLM uses a fitted expression from Flanner et al. (2006) which is similar in form to their physically based model of isothermal metamorphism. The idea is that as temperature gradients increase, the TG model should smoothly approach the isothermal model. Taillander et al. (2007), present a more recent approach for TG metamorphism based on experiment that may be better. There is still no consensus.

Another expression that could be used to help verify our CICE code is from Fukuzawa and Akitaya (1993). They provide experimental results for snow grain growth rates (y in $m\,s^{-1}$) with respect to temperature gradient (x in $K\,m^{-1}$) for snow of temperature -16°C; the linear relationship estimated from their figure 8 is

$$y = 0.8 \times 10^{-11} x + 0.2 \times 10^{-9}.$$

If the snow layer density is high enough (how high?), it will not metamorphose.Presumably there is also a maximum snow grain size for TG metamorphism. Domine et al., 2008 gives effective radius values ranging from 0.4 to 2.8 mm.

Flanner et al. (2006) apply a model of isothermal SSA evolution to snow on sea ice (which we are not using) and compute the temperature gradient for a single snow layer in the following way. The temperature difference is taken between surface air and the mean of sea-ice and snow temperatures. We may need to do this as well. The rationale is that a single snow layer cannot resolve the strong temperature gradient that often exists in near-surface snow and likely underestimates aging.

Chapter 10

Appendix: Details on Wet Metamorphism

Below are additional details of snow wet metamorphism. Wet snow metamorphism involves dynamic growth from liquid water redistribution among the grains. According to Colbeck (1973), the larger snow grains grow during melt while the smaller grains become liquid. This is because of different melting-point tempeartures being related to different radii of curvature at the ice interface. There is an empirical relationship derived from laboratory experiments (Brun 1989, Marshall 1989) that represents grain size growth as a function of liquid water content (f_{ice}: the total amount of liquid available in a snowpack, also called "free water content" and includes irreducible and gravity f_{ice}. f_{ice} is measured in % mass).

The growth of "saturated" and "low-f_{ice}" snow increases approximately linearly in volume over time for fixed f_{ice}. For low-f_{ice} snow, radius (r)

$$\begin{aligned}\frac{dV}{dt} &= \frac{4\pi}{3}(a + b \cdot f_{ice}^3) \\ \frac{dV}{dr}\frac{dr}{dt} &= \frac{4\pi}{3}(a + b \cdot f_{ice}^3) \\ \Delta r &= \frac{(a + b \cdot f_{ice}^3)}{3r^2}\Delta t \qquad (10.1)\end{aligned}$$

with $a = 3.0558 \times 10^{-9}\, mm^3 s^{-1}$, $b = 1.01 \times 10^{-10} mm^3 s^{-1}$, r in mm, time t in seconds and f_{ice} is in % mass with a maximum of around 10 (Brun 1989). The maximum volume growth rate for an f_{ice} of 10 is $4.3 \times 10^{-7}\, mm^3\, s^{-1}$.

Conversion from % mass (f_{ice}) to % volume (ϕ_{liq}) is

$$f_{ice} = \frac{\phi_{liq}\rho_l}{\rho_s} \qquad (10.2)$$

In liquid saturated snow, mean grain volume (v_g) increases at the constant rate (Raymond and Tusima, 1979) of $(\dot{v}_g)_{max} = (1.3{-}1.6) \times 10^{-6}\, mm^3\, s^{-1}$. This is about 3-4 times larger than the maximum value of (**??**). According to Brun (1989), this is because (**??**) refers to the Pendular regime where the liquid water is not continuous and thus limits growth.

Presumably this relationship can be used up to some maximum radius for wet snow (Domine et al., 2008 give a range of 0.5 to 2 mm). However much larger values may be possible. Meinander et al, 2013 measure grain sizes for wet snow on land of 3 mm, but argue that the effective grain size may be much higher because of the surrounding water.