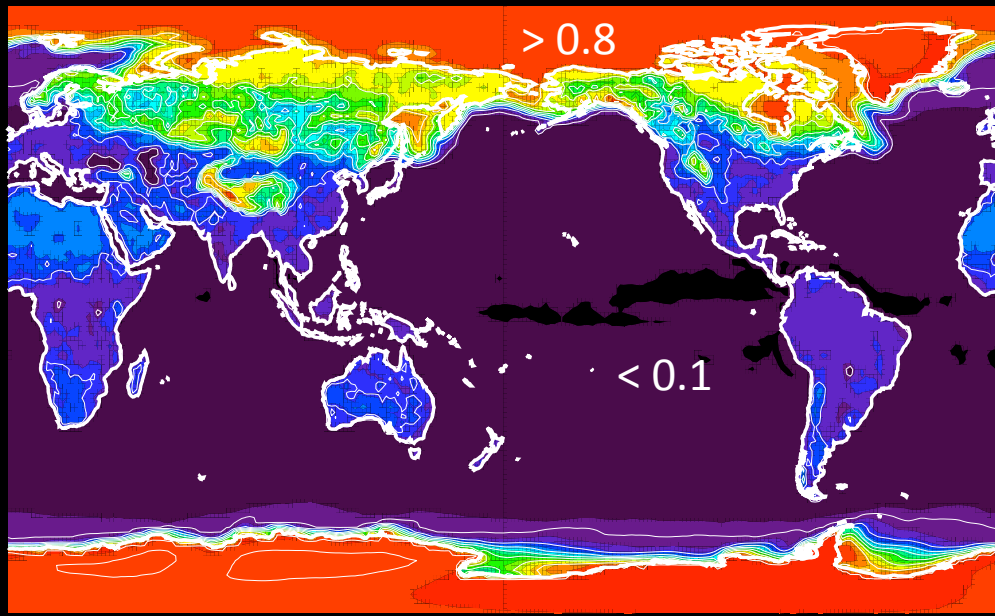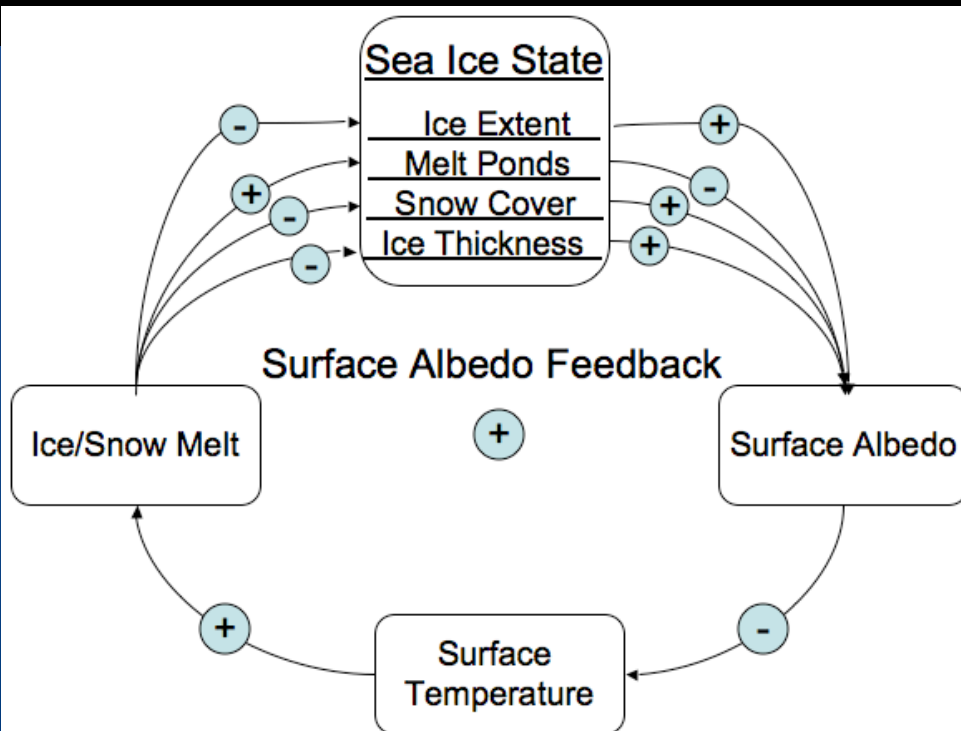# Sea Ice Modeling for Climate Applications

David A Bailey (NCAR)
With contributions from: Marika Holland and
Jennifer Kay (NCAR), Cecilia Bitz (U.
Washington), and Elizabeth Hunke (LANL)

Surface albedo

> 0.8

< 0.1

Why do we care about sea ice? Surface energy (heat) budget

Sea Ice State
Ice Extent
Melt Ponds
Snow Cover
Ice Thickness

Surface Albedo Feedback

Ice/Snow Melt

Surface Albedo

Surface Temperature

- High albedo of sea ice modifies radiative fluxes
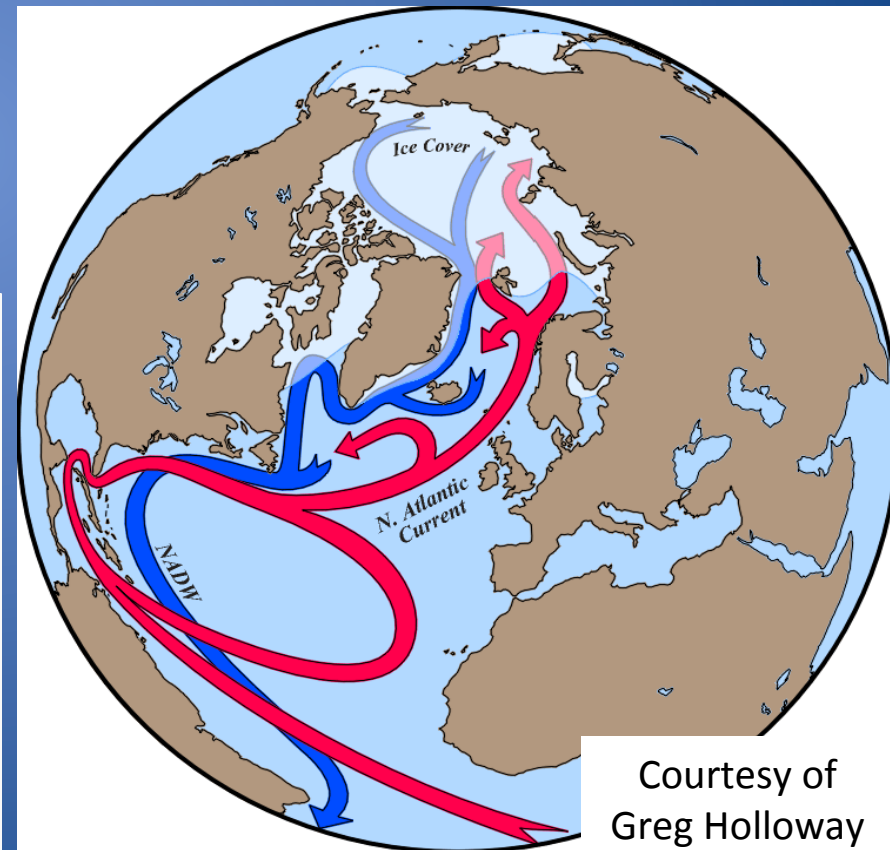- Sea ice insulates ocean from atmosphere influencing turbulent heat & moisture exchange

Fresh Water Flux (cm/day)
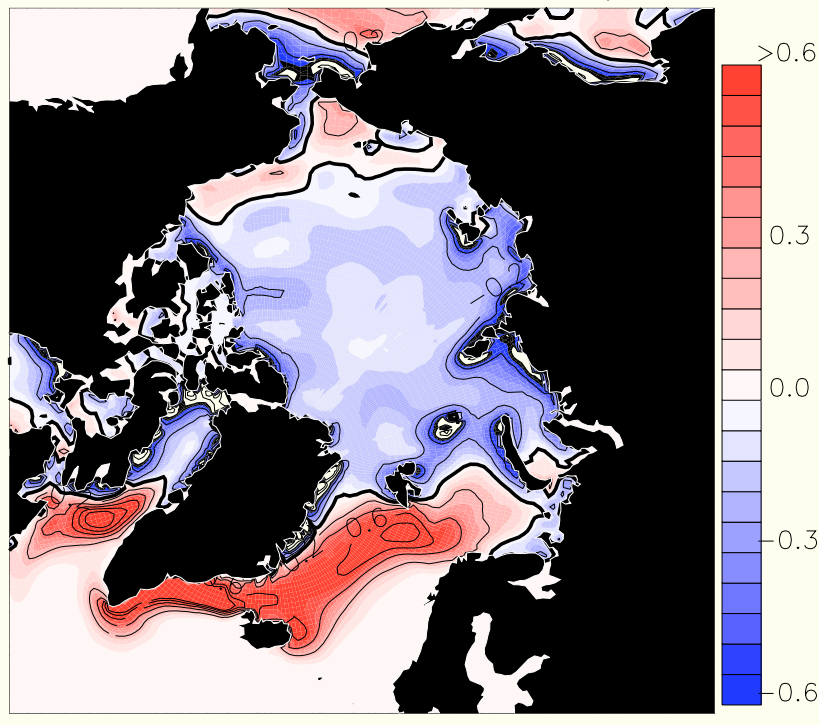
Ice-Ocean Freshwater Exchange

Why do we care about sea ice? Hydrological Cycle

Courtesy of Greg Holloway

- Salt rejection during ice formation leaves sea ice relatively fresh (salt flux to ocean)
- Ice melt releases freshwater back to the ocean
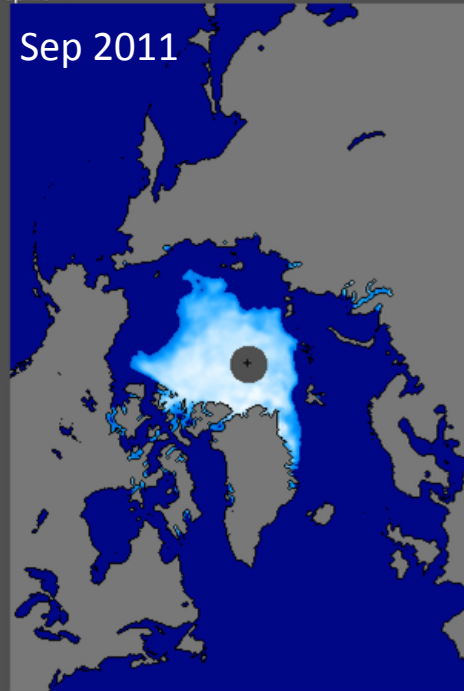- Can modify ocean circulation

Contrasting the Hemispheres

• Arctic Ocean surrounded by land (thicker ice).

• Southern Ocean unbounded (free drift).

• Larger seasonal cycle in south.

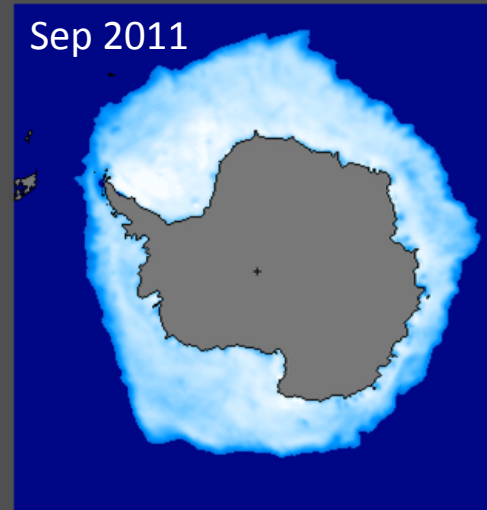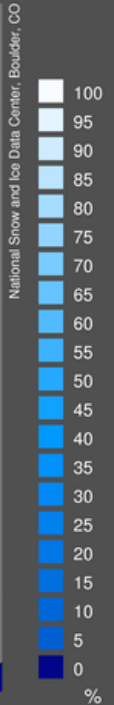• Winter extent set by ocean in south and land/ocean in north.

Sea Ice Concentration
Sep 2011

Sep 2011

Total area = 2.9 million sq km

Sea Ice Concentration
Mar 2011
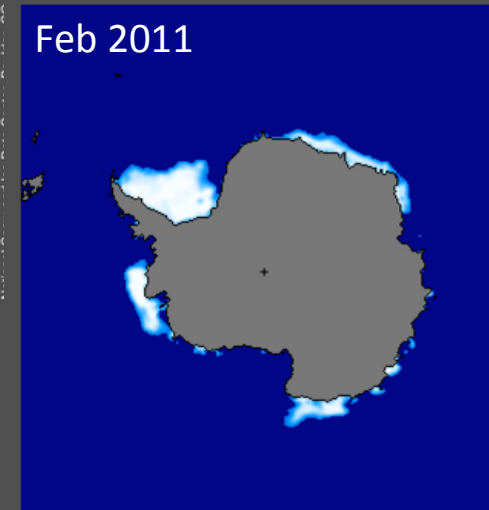
Mar 2011

Total area = 12.4 million sq km

Sep 2011

Total area = 14.4 million sq km

Feb 2011

Total area = 1.8 million sq km

National Snow and Ice Data Center, Boulder, CO

NSIDC
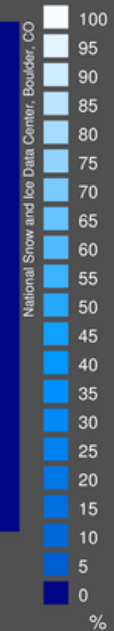
1 km

10 m

From: Feltham, 2008 (photos by Hajo Eicken)

440 m

Sea Ice

- Composed of floes (can freeze to form a continuous cover)
- Typical thickness of meters
- Riddled with cracks (leads) and ridges
- Complex mosiac of ice types within small area

Photo courtesy of Don Perovich

What do we need in a sea ice model for climate applications?

- Model which simulates a reasonable mean state/variability of sea ice

  – Concentration, thickness, mass budgets

- Realistically simulates ice-ocean-atmosphere exchanges of heat and moisture

- Realistically simulates response to climate perturbations - key climate feedbacks

CICE: the Los Alamos Sea Ice Model
Documentation and Software User's Manual
Version 4.1
LA-CC-06-012

Elizabeth C. Hunke and William H. Lipscomb
T-3 Fluid Dynamics Group, Los Alamos National Laboratory
Los Alamos NM 87545

May 5, 2010

CESM1 uses the CICE Los Alamos Sea Ice Model (Hunke and Lipscomb)

Full documentation available online

Sea Ice Models Used in Climate Simulations

- Two primary components
  - Dynamics
    - Solves force balance to determine sea ice motion
  - Thermodynamics
    - Solves for vertical ice temperature profile
    - Vertical/lateral melt and growth rates

- Some (about 30% of IPCC-AR4, 50% for AR5?) models also include
  - Ice Thickness Distribution
    - Subgridscale parameterization
    - Accounts for high spatial heterogeneity in ice

Dynamics

NOAA

Sea Ice Model - Dynamics

- Force balance between wind stress, water stress, internal ice stress, coriolis and stress associated with sea surface slope

- Ice treated as a continuum with an effective large-scale rheology describing the relationship between stress and deformation

- Ice freely diverges (no tensile strength)

- Ice resists convergence and shear

(e.g. Hibler, 1979)

$$m\frac{D\mathbf{u}}{Dt} = -mf\mathbf{k} \times \mathbf{u} + \boldsymbol{\tau}_a + \boldsymbol{\tau}_w - mg_r \nabla Y + \nabla \cdot \sigma$$

Total derivative Coriolis Air stress Ocean stress Sea Surface Slope Internal Ice Stress

Sea Ice Model - Dynamics

- ## Air Stress

$$\vec{\tau}_a = \frac{\rho_a u^{*2} \vec{U}_a}{|\vec{U}_a|},$$

$$u^* = c_u \left|\vec{U}_a\right|$$

- ## Ocean Stress

$$\vec{\tau}_w = c_w \rho_w \left|\vec{U}_w - \vec{u}\right| \left[\left(\vec{U}_w - \vec{u}\right)\cos\theta + \hat{k} \times \left(\vec{U}_w - \vec{u}\right)\sin\theta\right]$$

(e.g. Hibler, 1979)

$$m\frac{D\mathbf{u}}{Dt} = -mf\mathbf{k} \times \mathbf{u} + \boldsymbol{\tau}_a + \boldsymbol{\tau}_w - mg_r\nabla Y + \nabla \cdot \sigma$$

Total derivative   Coriolis   Air stress   Ocean stress   Sea Surface Slope   Internal Ice Stress

Sea Ice Model – Dynamics

- Ice Interaction Term (Internal Ice Stress)
  - Requires a constitutive law to relate ice stress (σ) to ice strain rate ($\dot{\varepsilon}$)

(e.g. Hibler, 1979)

$$m\frac{D\mathbf{u}}{Dt} = -mf\mathbf{k} \times \mathbf{u} + \boldsymbol{\tau}_a + \boldsymbol{\tau}_w - mg_r\nabla Y + \nabla \cdot \sigma$$

Total derivative

Coriolis

Air stress

Ocean stress

Sea Surface Slope

Internal Ice Stress

Sea Ice Model - Dynamics

- Ice Interaction Term (Internal Ice Stress)
  - Requires a constitutive law to relate ice stress (σ) to ice strain rate ($\dot{\varepsilon}$)

For example – A compressive stress test

At first:

> Ice floe side view

Length L

After applying a compressive force, the ice deforms...

$\sigma \rightarrow$ $\quad$ Volume conserved so ice becomes thicker $\quad \leftarrow \sigma$

σ $\qquad\qquad\qquad\qquad\qquad$ σ

L+δL

Strain: $\varepsilon = \delta L / L$

Strain Rate: $\dot{\varepsilon} = \delta L / L dt$

Sea Ice Model - Dynamics

- ## Ice Interaction Term (Internal Ice Stress)

  - Use variant of Viscous-Plastic Rheology (Hibler, 1979)

  - Treats ice as a continuum - plastic at normal strain rates and viscous at very small strain rates.

  - Ice has no tensile strength (freely diverges) but resists convergence and shear (strength dependent on ice state)
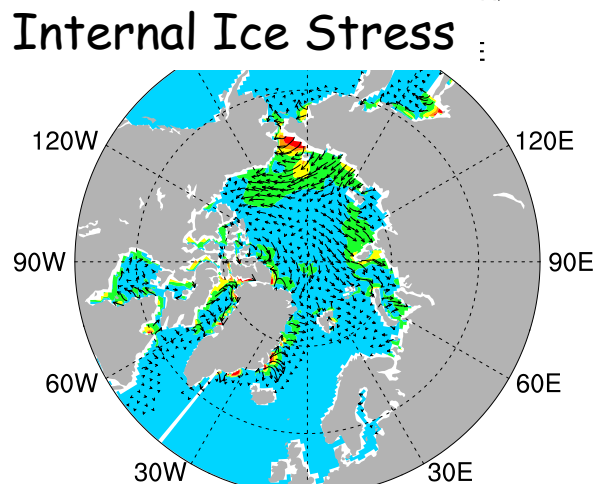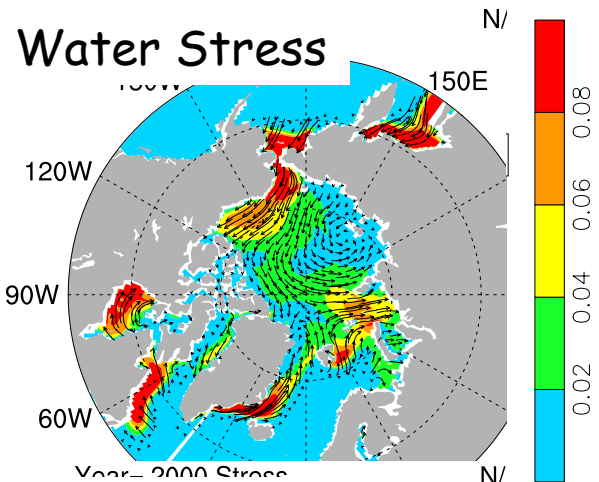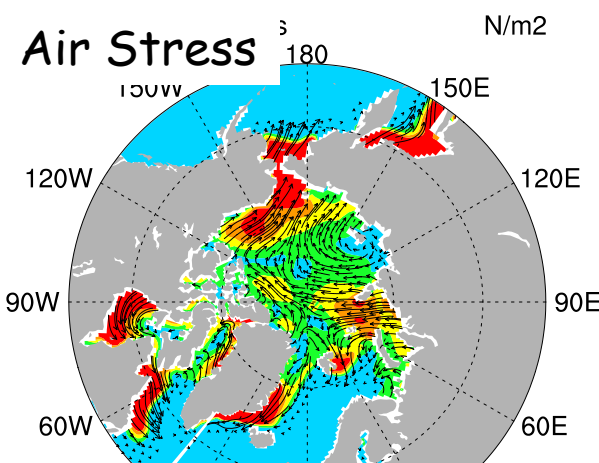
As shown in Feltham, 2008

"Failure Stress"
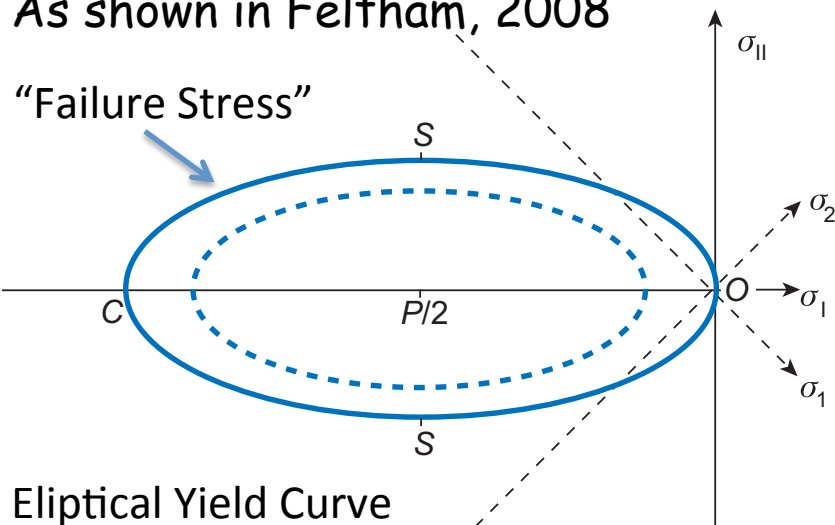
Eliptical Yield Curve

Elastic-Viscous-Plastic Model

EVP model uses explicit time stepping by adding elastic waves to constitutive law (Hunke and Dukowicz, 1997)
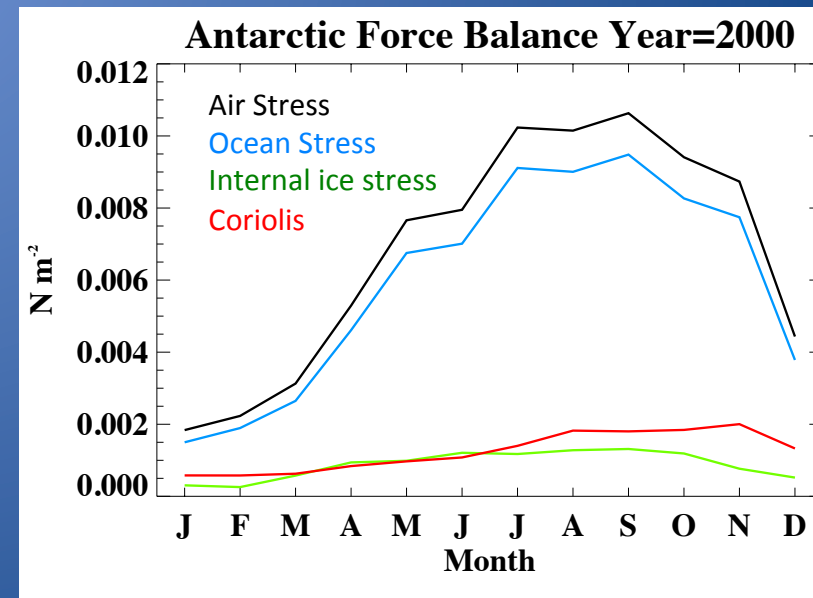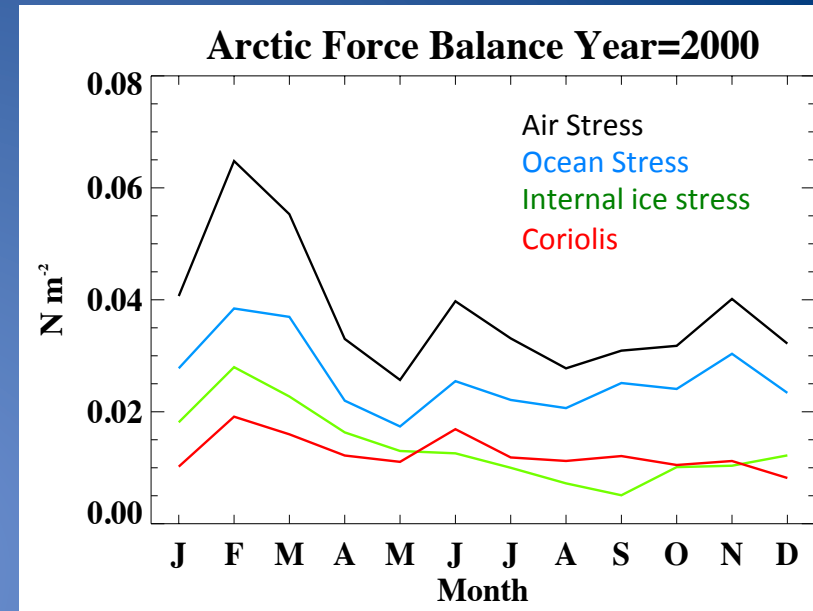
Simulated Force Balance

- Air stress largely balanced by ocean stress.

- Internal ice stress has smaller role

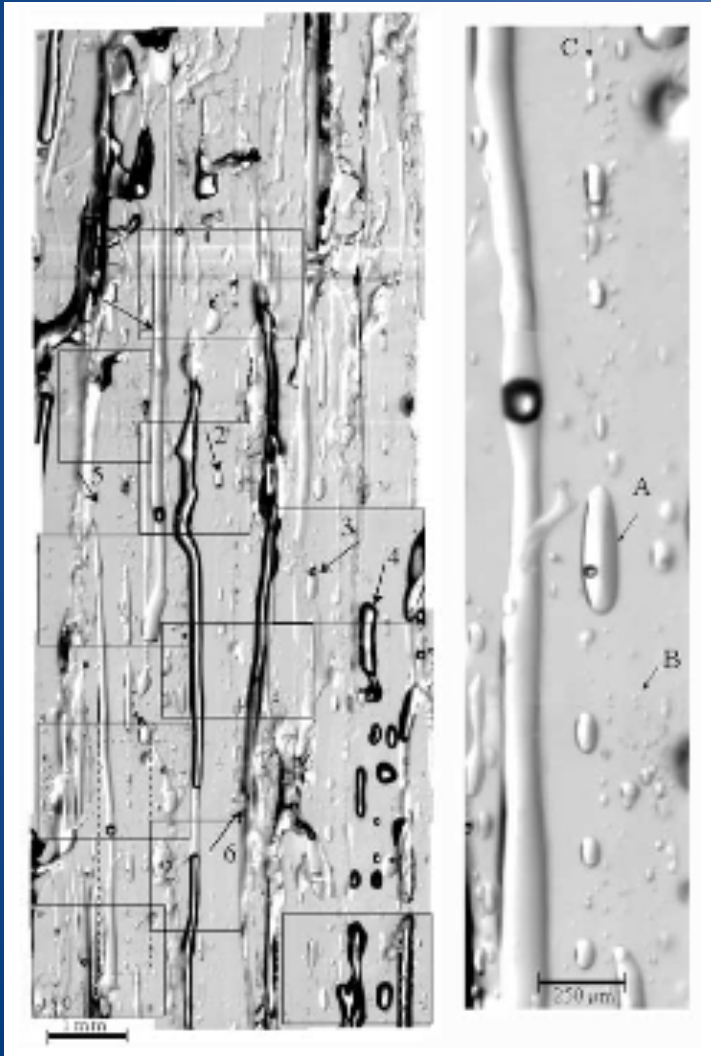- In Antarctic ice in nearly free drift - weak ice interaction term

Thermodynamics

Thermodynamics
Vertical heat transfer

$$\rho c \frac{\partial T}{\partial t} = \frac{\partial}{\partial z} k \frac{\partial T}{\partial z} + Q_{SW}$$

SEA ICE

(from Light, Maykut, Grenfell, 2003)

- Assume brine pockets are in thermal equilibrium with ice

- Heat capacity and conductivity are functions of T/S of ice

- Assume constant salinity profile

- Assume non-varying density

- Assume pockets/channels are brine filled

- Traditionally:

$$Q_{SW} = -\frac{d}{dz} I_{SW} e^{-\kappa z} \quad \text{where}$$
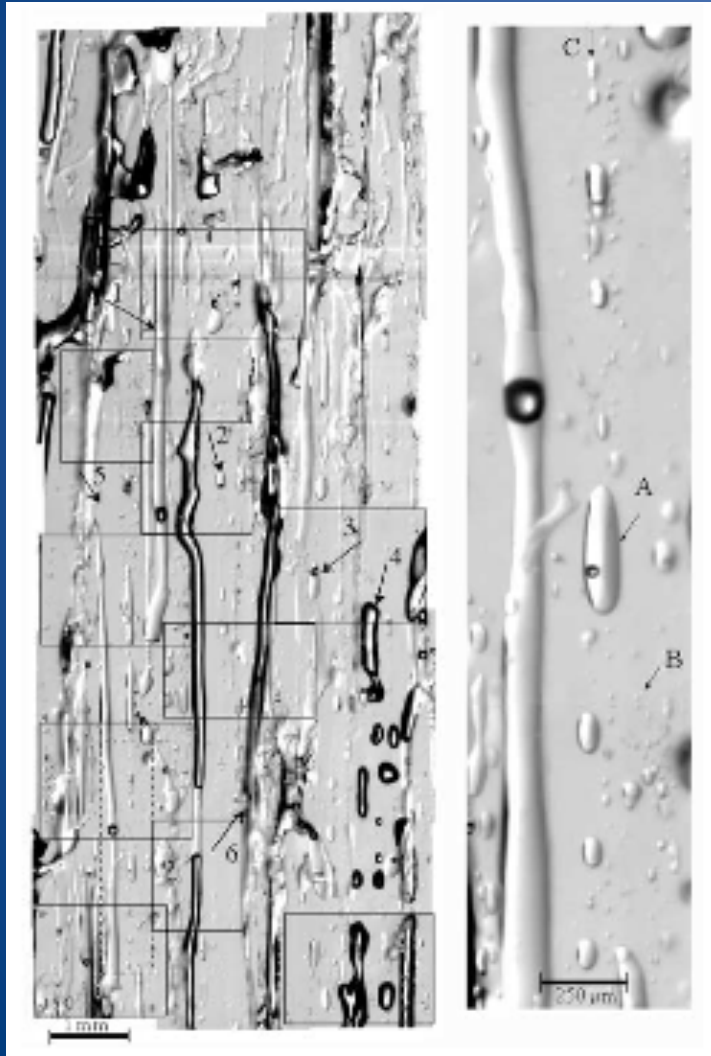
$$I_{SW} = i_0 (1-\alpha) F_{SW}$$

(Maykut and Untersteiner, 1971; Bitz and Lipscomb, 1999; others)

Thermodynamics
Vertical heat transfer

$$\rho c \frac{\partial T}{\partial t} = \frac{\partial}{\partial z} k \frac{\partial T}{\partial z} + Q_{SW}$$

SEA ICE

$$c(T, S) = c_o + \frac{\gamma S}{T^2}$$

where T is in Celsius,

$$\gamma = L_o \mu \quad \text{and} \quad T_m = -\mu S$$
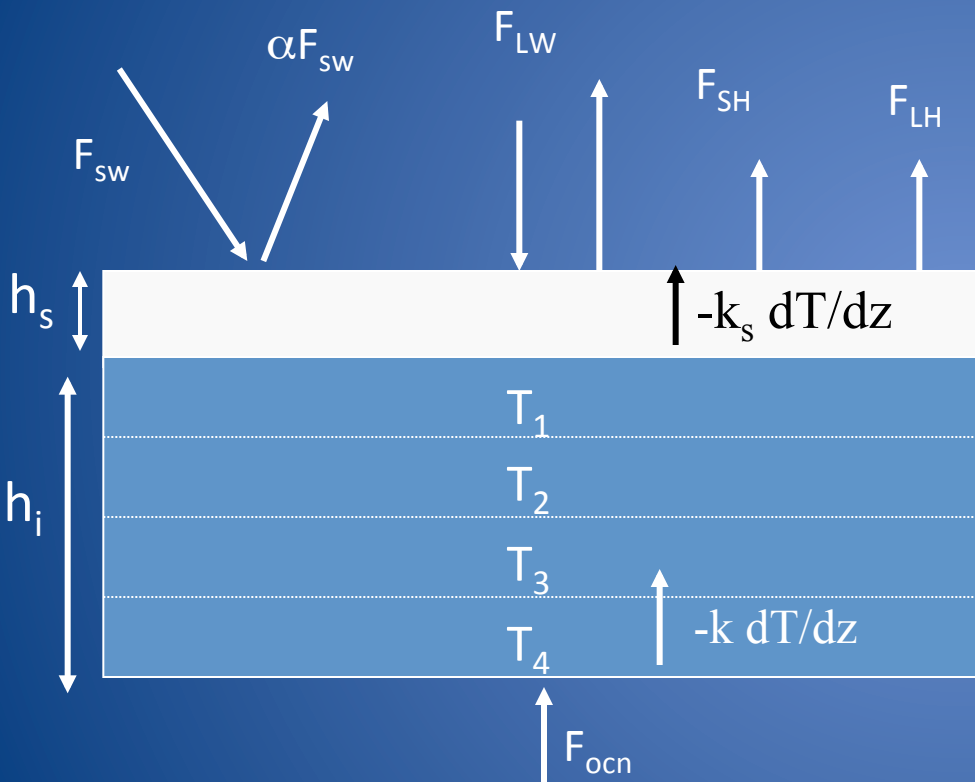
Untersteiner, 1961

Enthalpy:
Heat required to melt a unit of ice

$$q(S, T) = \rho c_o(-\mu S - T) + \rho L_o \left(1 + \frac{\mu S}{T}\right)$$

(from Light, Maykut, Grenfell, 2003)

(Maykut and Untersteiner, 1971; Bitz and Lipscomb, 1999; others)

Sea ice thermodynamics

Balance of fluxes at surface

$$(1-\alpha)F_{SW} + F_{LW} - \sigma T^4 + F_{SH} + F_{LH} + k\frac{\partial T}{\partial z} = -q\frac{dh}{dt}$$

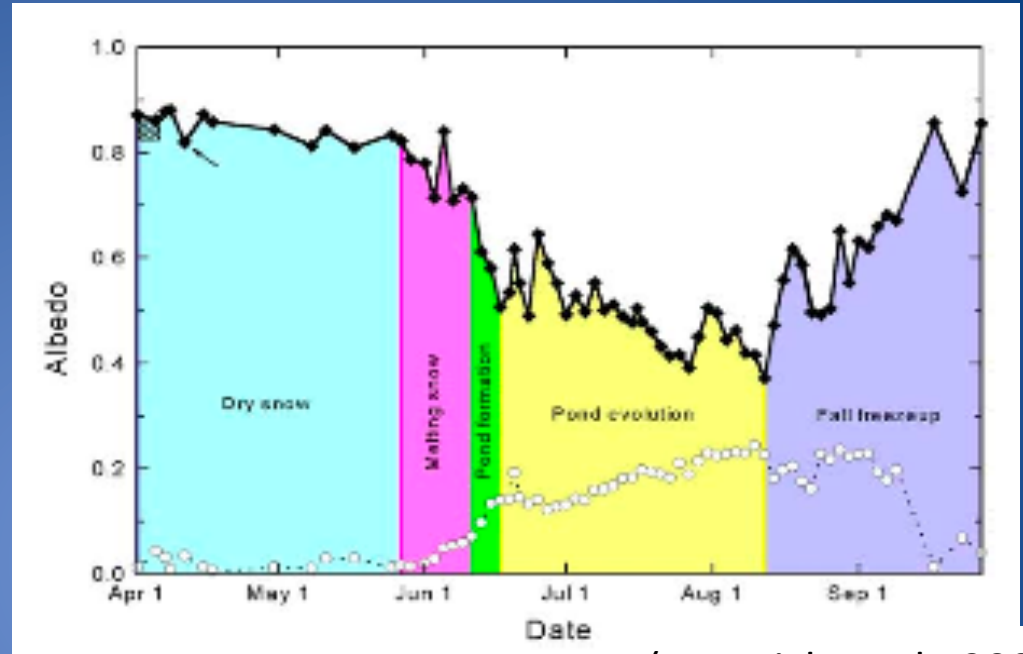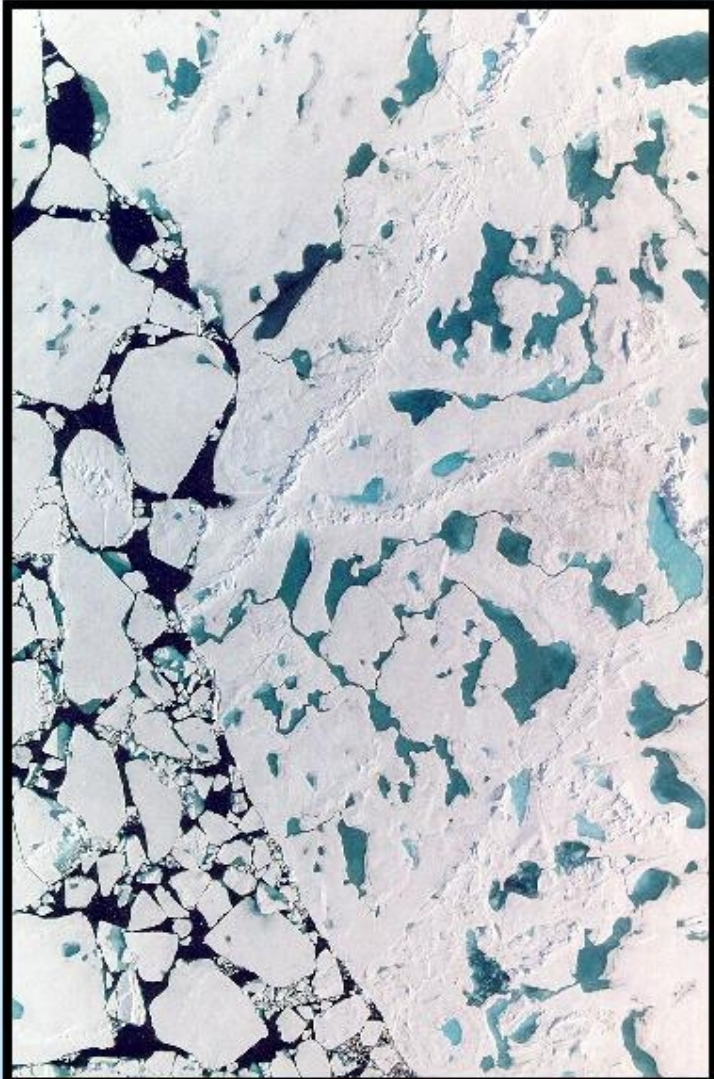Vertical heat transfer (conduction, SW absorption)

$$\rho c \frac{\partial T}{\partial t} = \frac{\partial}{\partial z} k \frac{\partial T}{\partial z} + Q_{SW}$$

Balance of fluxes at ice base

$$F_{ocn} - k\frac{\partial T}{\partial z} = -q\frac{dh}{dt}$$

αF_{sw} F_{LW} F_{SH} F_{LH}

F_{sw}

h_s

h_i

$-k_s \, dT/dz$

T_1

T_2

T_3

T_4

$-k \, dT/dz$

F_{ocn}

Allows us to compute surface melt (snow or ice), ice basal melt and ice growth

Albedo

(Perovich et al., 2002)

Often the parameterized sea ice albedo depends on characteristics of surface state (snow, temp, ponding, h_i).

Surface albedo accounts for fraction of gridcell covered by ice vs open ocean

A Delta-Eddington Multiple Scattering Parameterization for Solar Radiation in the Sea Ice Component of the Community Climate System Model

B. P. Briegleb and B. Light
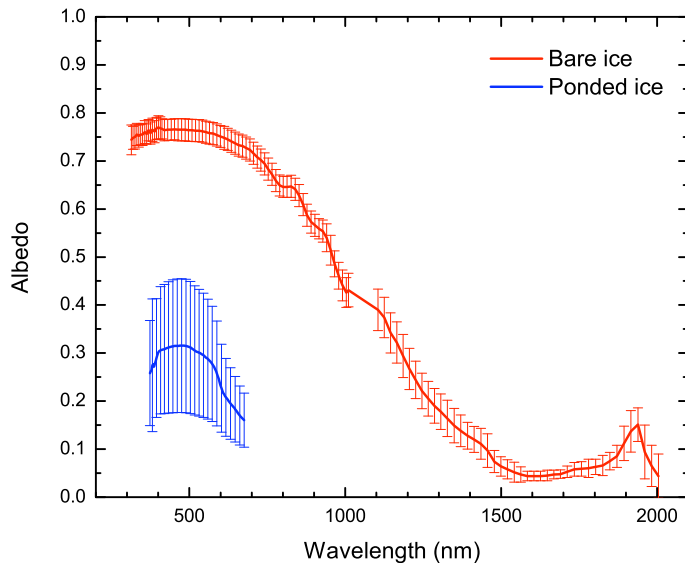
New Solar Radiation parameterization
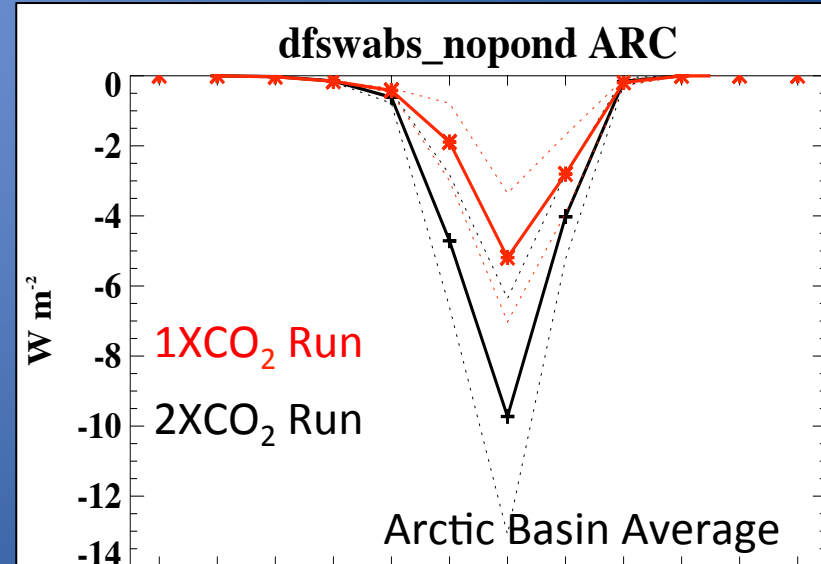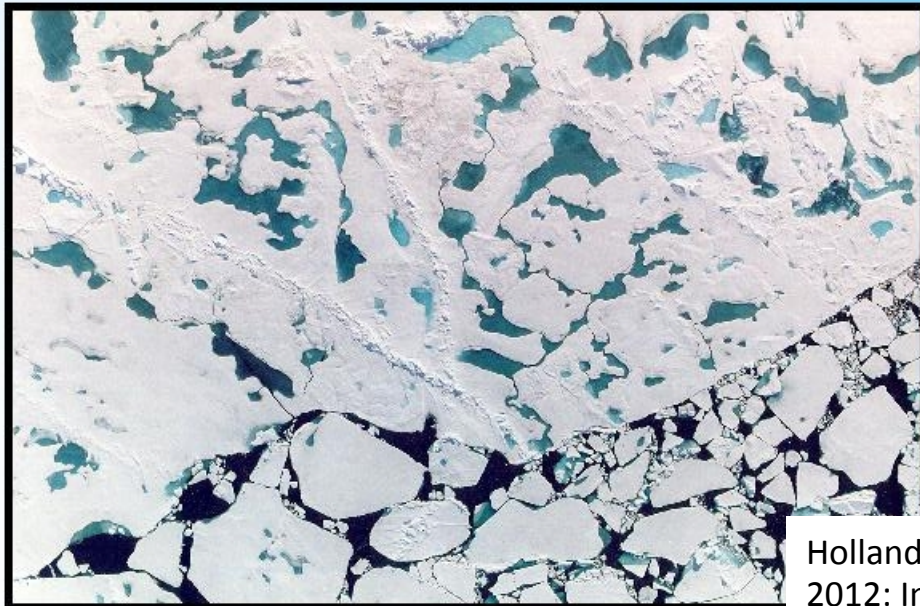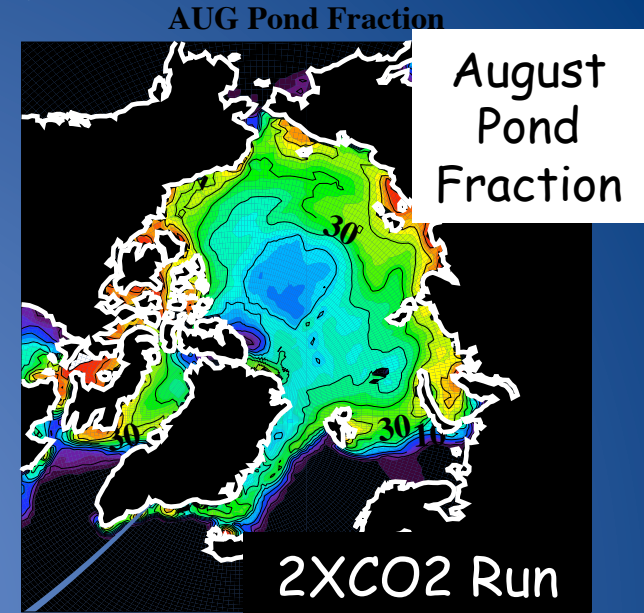
Better physics:

- makes use of inherent optical properties to define scattering and absorption of snow, sea ice and included absorbers

More flexible

- Explicitly allows for included absorbers in sea ice

Melt Pond Parameterization

- New radiative transfer allows (requires) a pond parameterization

- Only influences radiation

- Pond volume depends on surface meltwater, assuming a runoff fraction

AUG Pond Fraction

August Pond Fraction

2XCO2 Run

dfswabs_nopond ARC

1XCO$_2$ Run

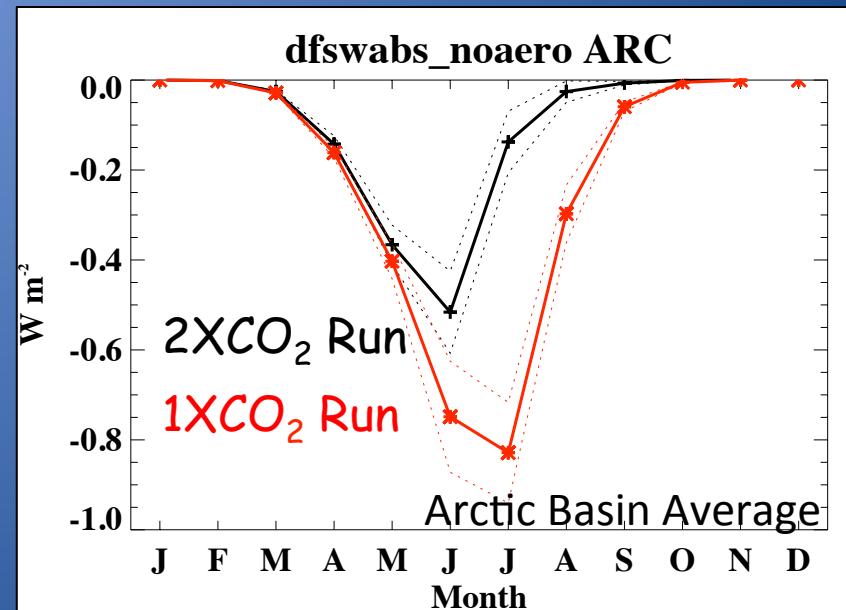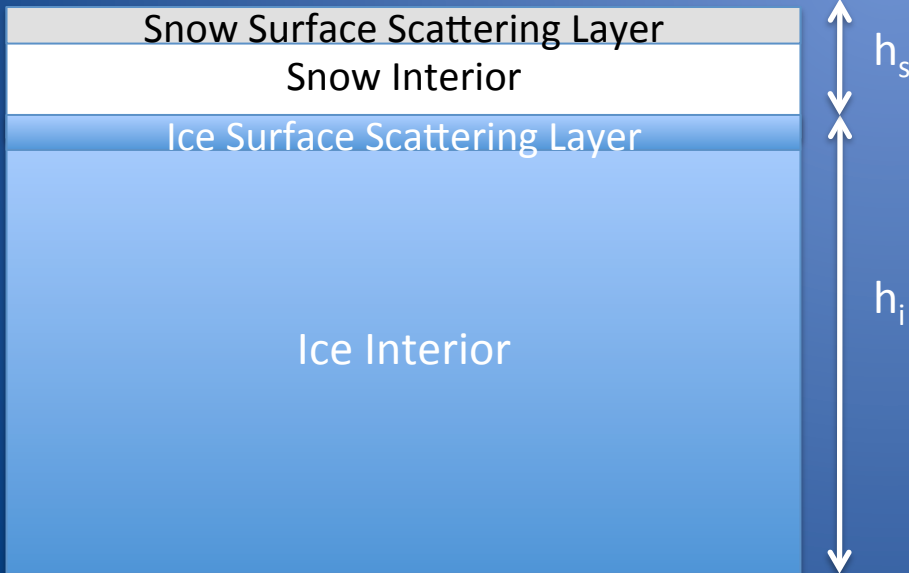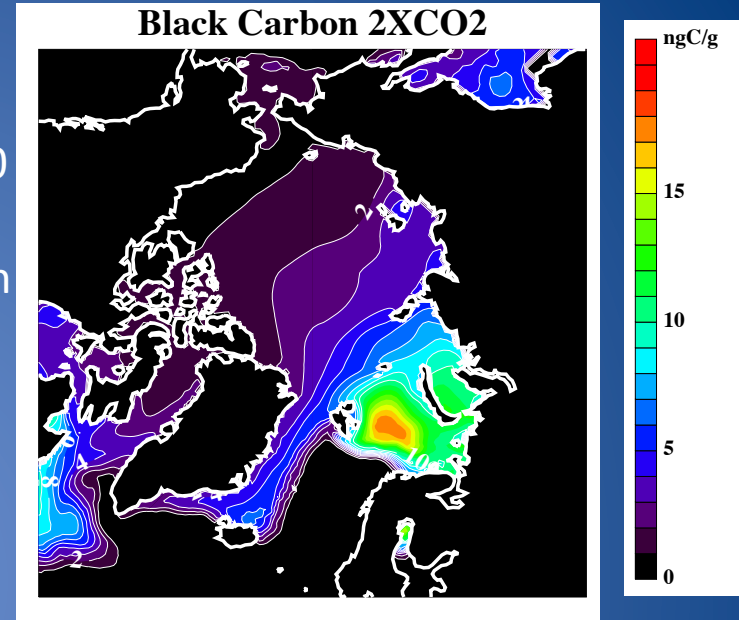2XCO$_2$ Run

Arctic Basin Average

W m^{-2}

Holland, M. M., D. A. Bailey, B. P. Briegleb, B. Light, and E. C. Hunke, 2012: Improved sea ice shortwave radiation physics in CCSM4: The impact of melt ponds and black carbon. J. Climate, 25, 1413-1430.

Aerosol deposition and cycling

- Aerosol deposition and cycling now included.
- Account for black carbon and dust aerosols
- These are deposited from the atmosphere and modified by melt and transport

With 1850 Aerosol Deposition

Black Carbon 2XCO2

Snow Surface Scattering Layer
Snow Interior
Ice Surface Scattering Layer
Ice Interior
h_s
h_i

dfswabs_noaero ARC

2XCO$_2$ Run
1XCO$_2$ Run
Arctic Basin Average

Holland, Bailey, Briegleb, Light, and Hunke, 2012

Ice Thickness Distribution

To represent high spatial heterogeneity of sea ice
Schematic of model representation with five ice "categories"
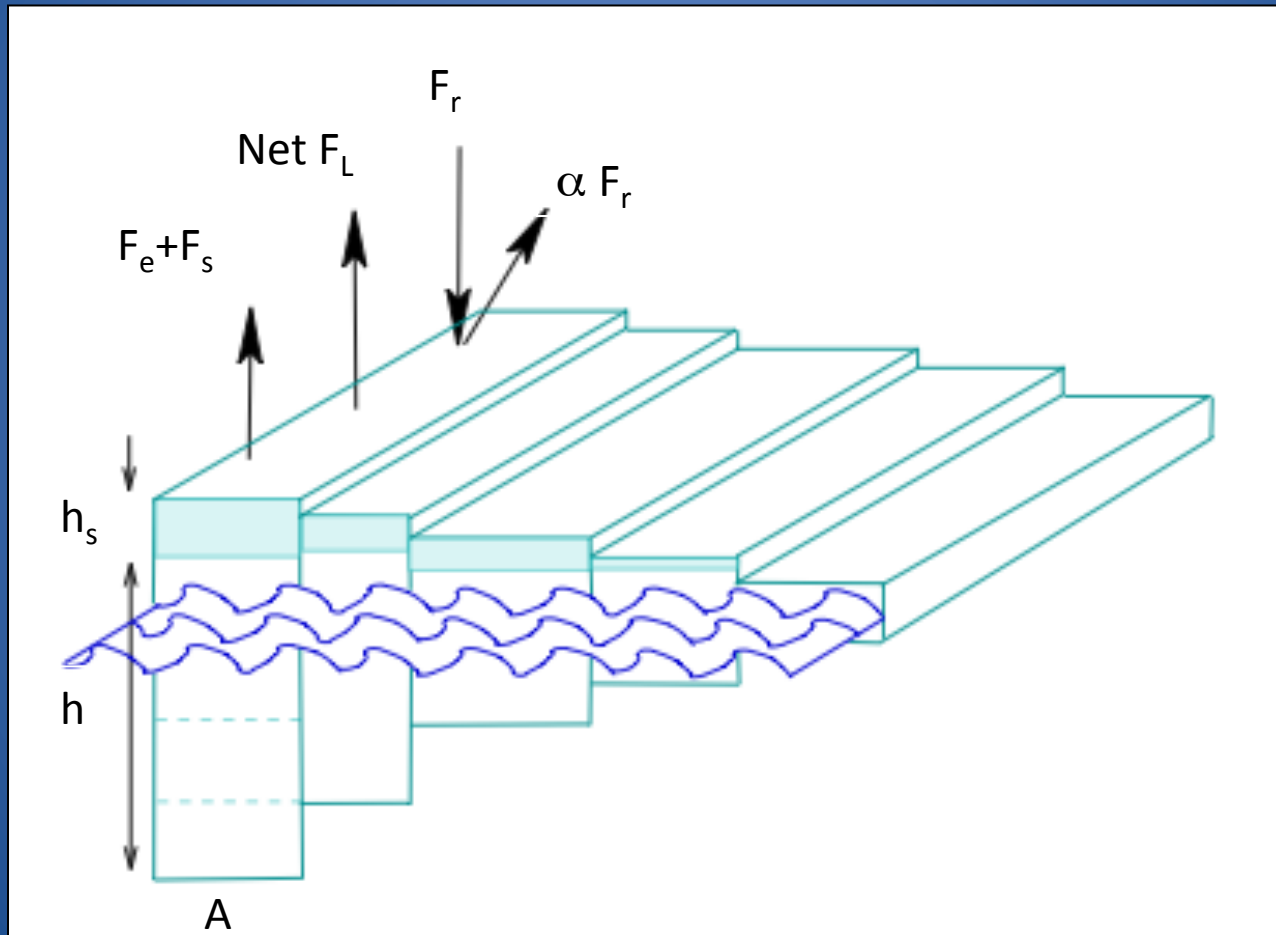
A=fractional coverage of a category

Ice Thickness Distribution

Ice thickness distribution g(x,y,h,t) evolution equation
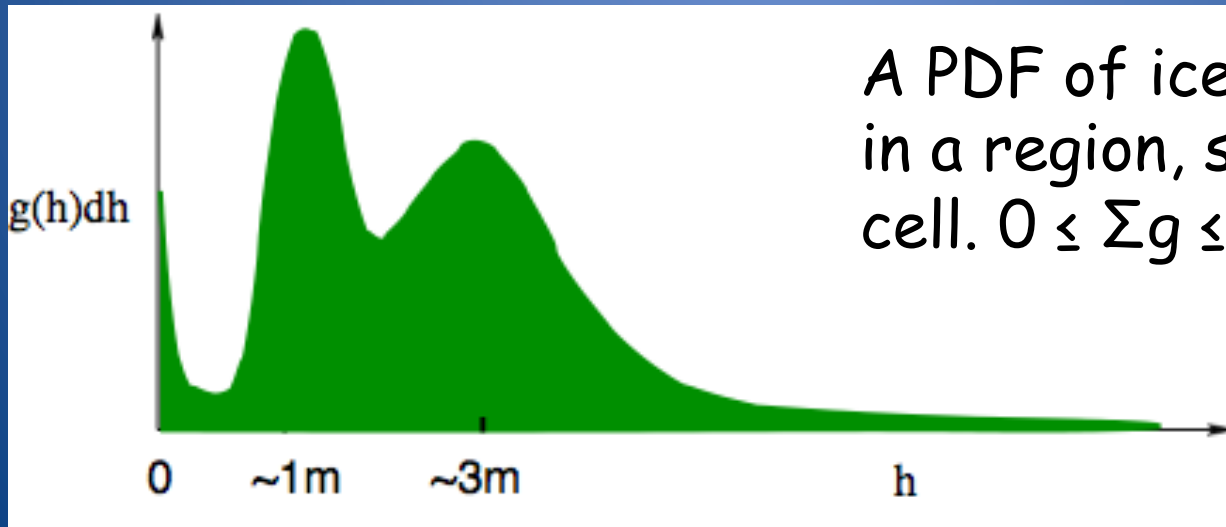   from Thorndike et al. (1975)

$$\frac{\partial g}{\partial t} = -\frac{\partial}{\partial h}(fg) + L(g) - \nabla \cdot (\vec{v}g) + \Psi(h, g, \vec{v})$$

Ice Growth   Lateral Melt   Convergence   Mechanical Redistribution

g(h)dh

0   ~1m   ~3m   h
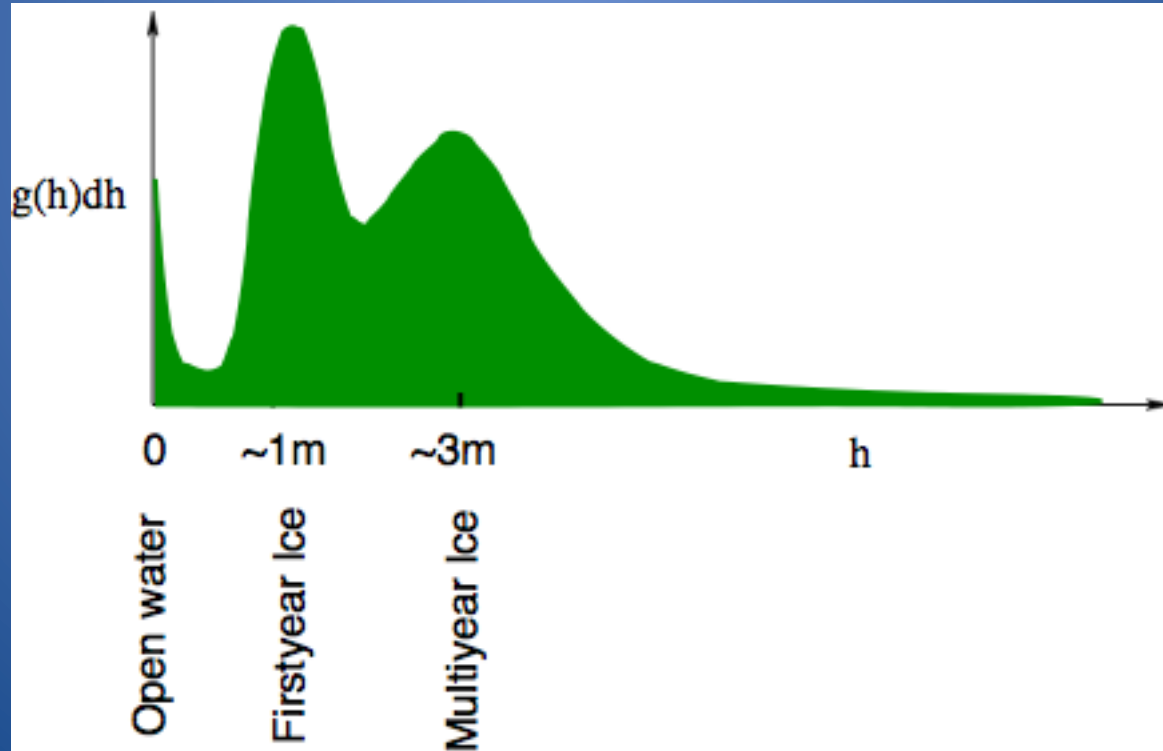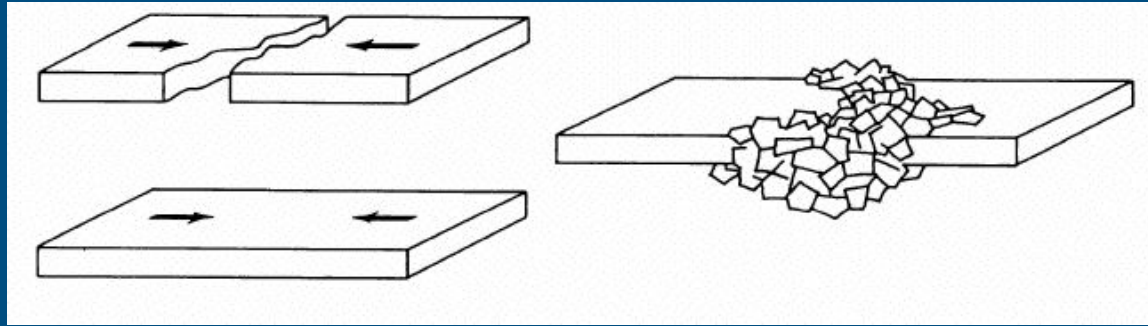
A PDF of ice thickness h in a region, such as a grid cell. $0 \leq \Sigma g \leq 1$

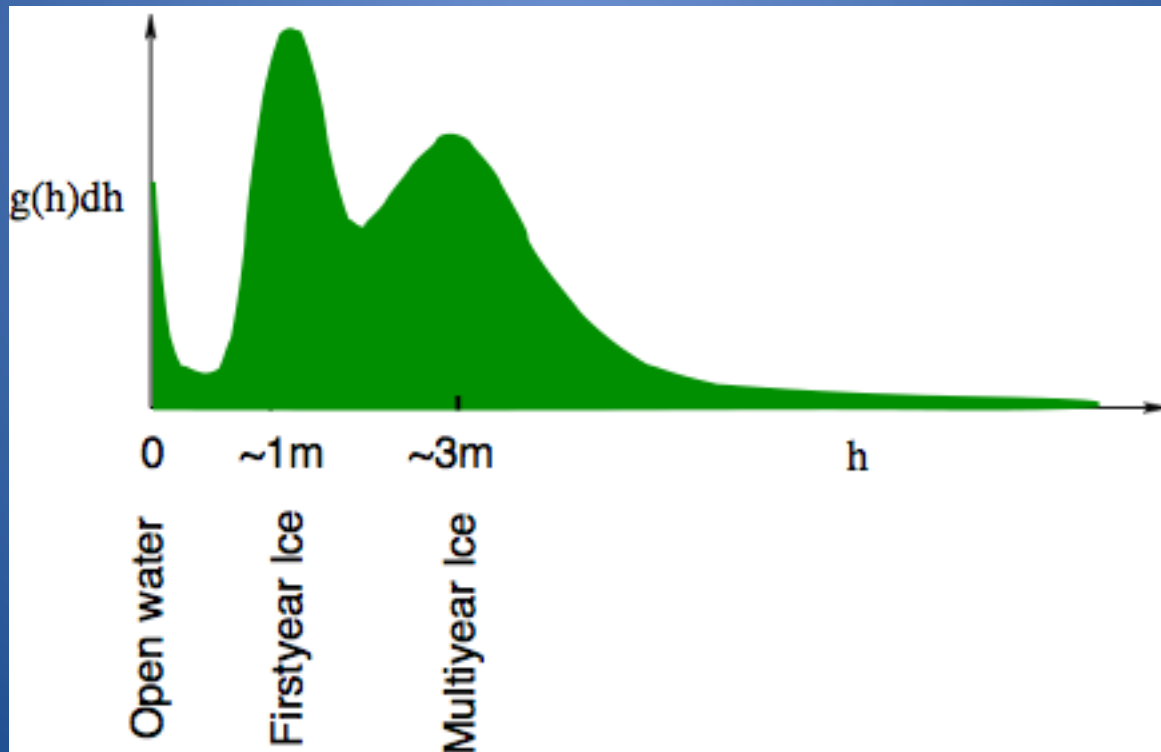g(h)dh is the fractional area covered by ice of thickness h to h+dh

Ψ = Mechanical redistribution

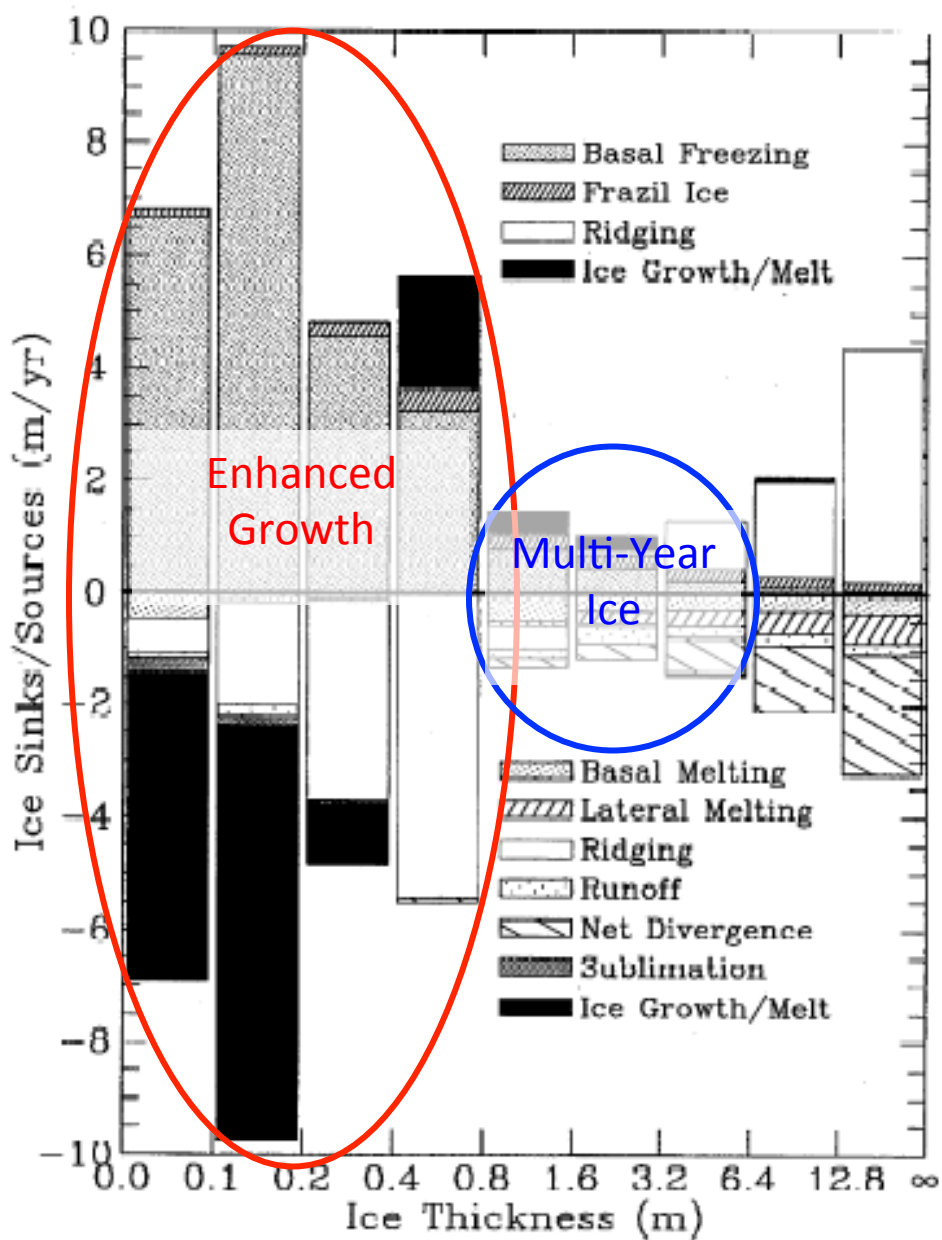Transfers ice from thin part of distribution to thicker categories

Converging hypothetical floes

Ice growth:
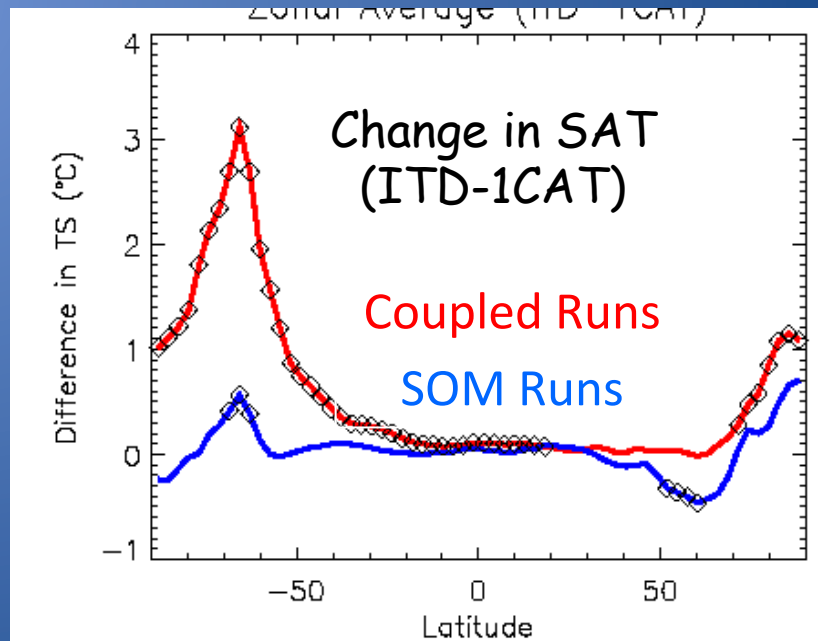
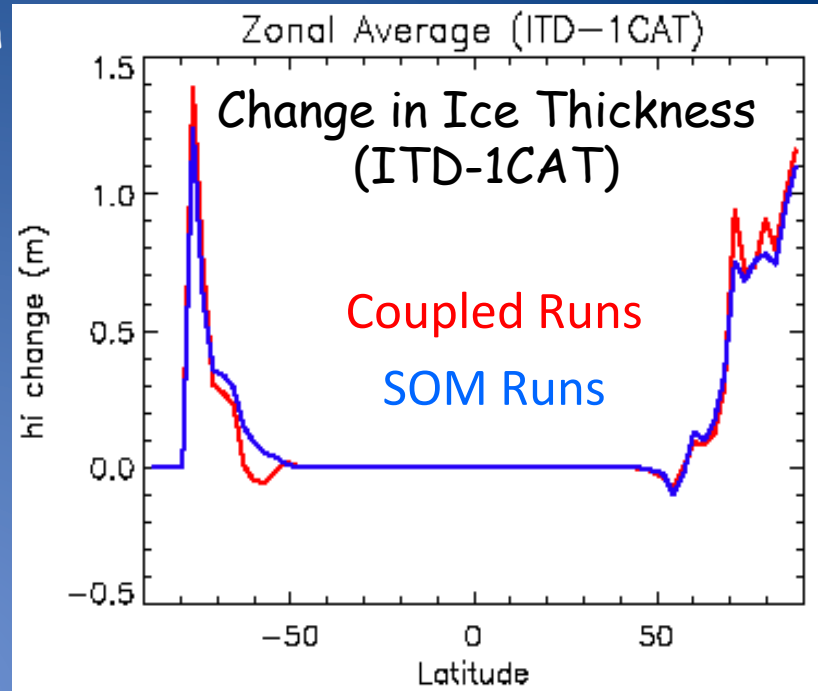# Influence of including an Ice Thickness Distribution

Enhanced Growth

Multi-Year Ice

Change in Ice Thickness (ITD-1CAT)

Coupled Runs

SOM Runs

Change in SAT (ITD-1CAT)

Coupled Runs

SOM Runs

Schramm et al., 1997

(Holland et al., 2006)

State variables for each category:

A, V_i, V_s, $E_i(z)$, $E_s(z)$, T_{surf}, melt pond state, aerosol contents (z), etc.

A = category area per unit gridcell area (or fractional coverage)
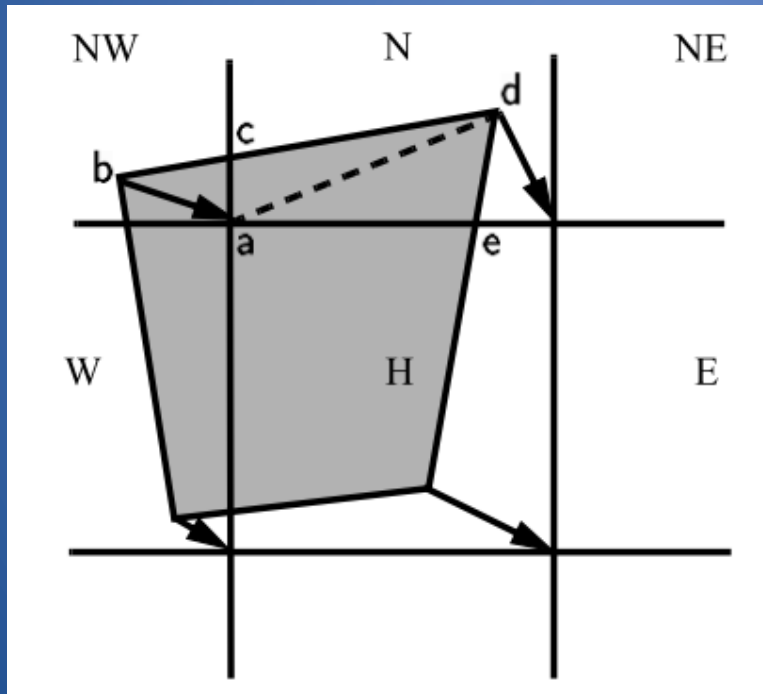$V = hA$ is the category volume per unit gridcell area
$E = Vq$ is the category enthalpy per unit gridcell area

V and E are preferred as state variables because they are conserved quantities (rather than T).

Advection

Would make so many state variables prohibitive, if it weren't for remapping by Lipscomb and Hunke 2004.
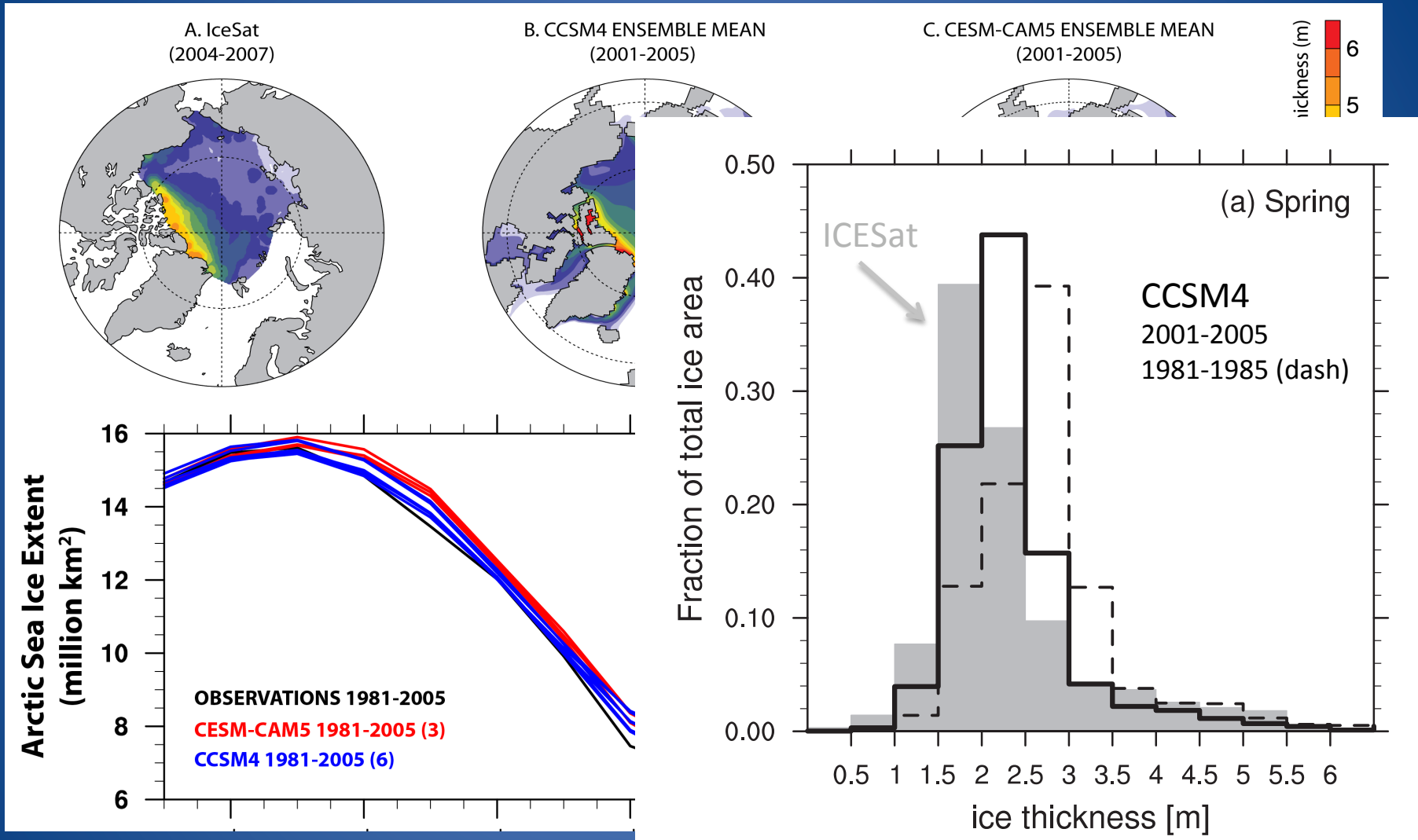

Conserved quantities are remapped from the shaded "departure region", which is computed from backward trajectories of the ice motion field.

Science Highlights

- How well does the model actually simulate the sea ice cover?

- What does the model say about the future of sea ice?

- Northern versus Southern Hemisphere?

CCSM4/CESM1 Simulation of Arctic sea ice cover

A. IceSat
(2004-2007)

B. CCSM4 ENSEMBLE MEAN
(2001-2005)

C. CESM-CAM5 ENSEMBLE MEAN
(2001-2005)

thickness (m)

6

5

Arctic Sea Ice Extent (million km²)

OBSERVATIONS 1981-2005
CESM-CAM5 1981-2005 (3)
CCSM4 1981-2005 (6)

(a) Spring

ICESat

CCSM4
2001-2005
1981-1985 (dash)

Fraction of total ice area

ice thickness [m]
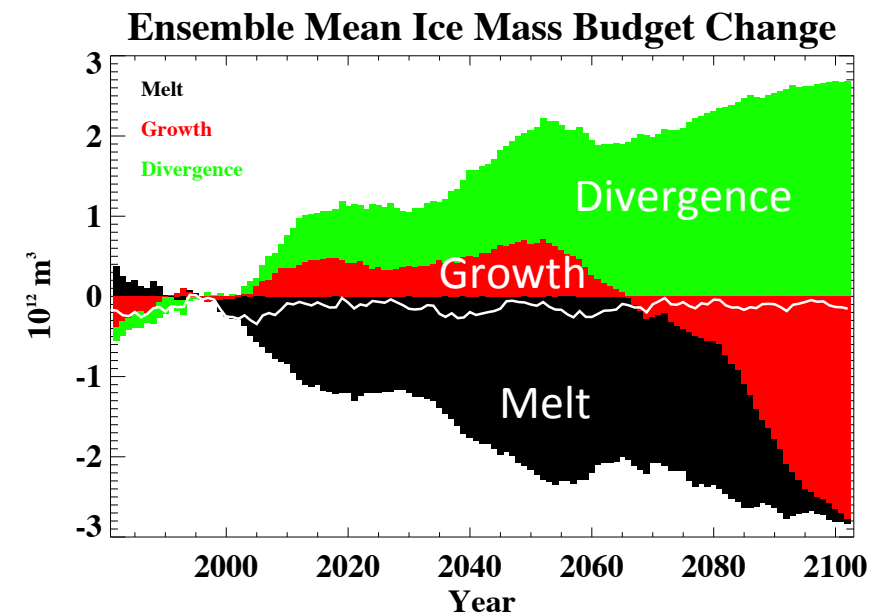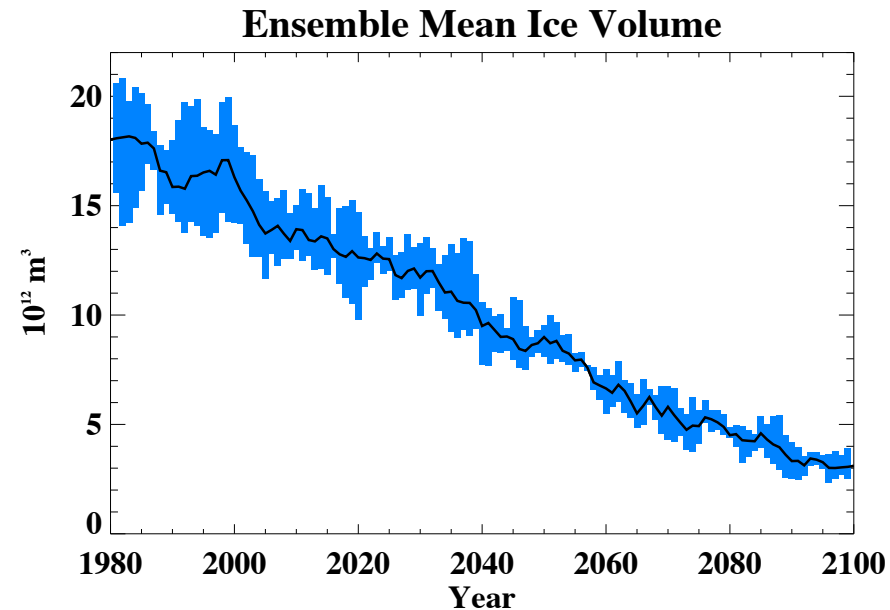
Neale et al. in preparation

Jahn, A and Coauthors, 2012: Late-Twentieth-Century Simulation of Arctic Sea Ice and Ocean Properties in the CCSM4. J. Climate, 25, 1431–1452.
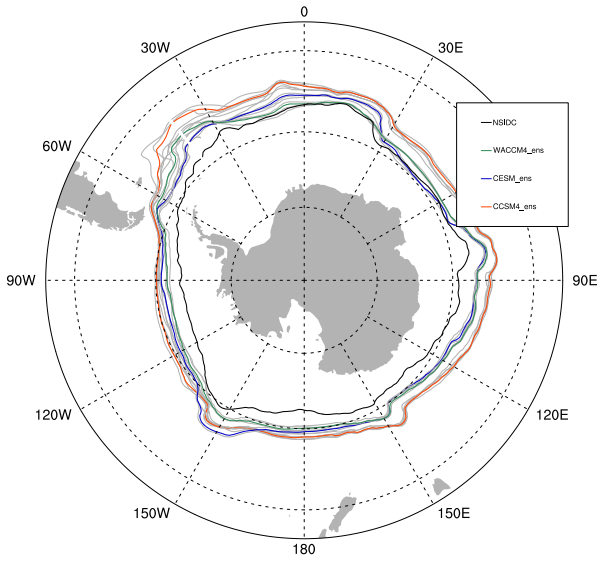
CCSM4 21ˢᵗ Arctic Ice Loss
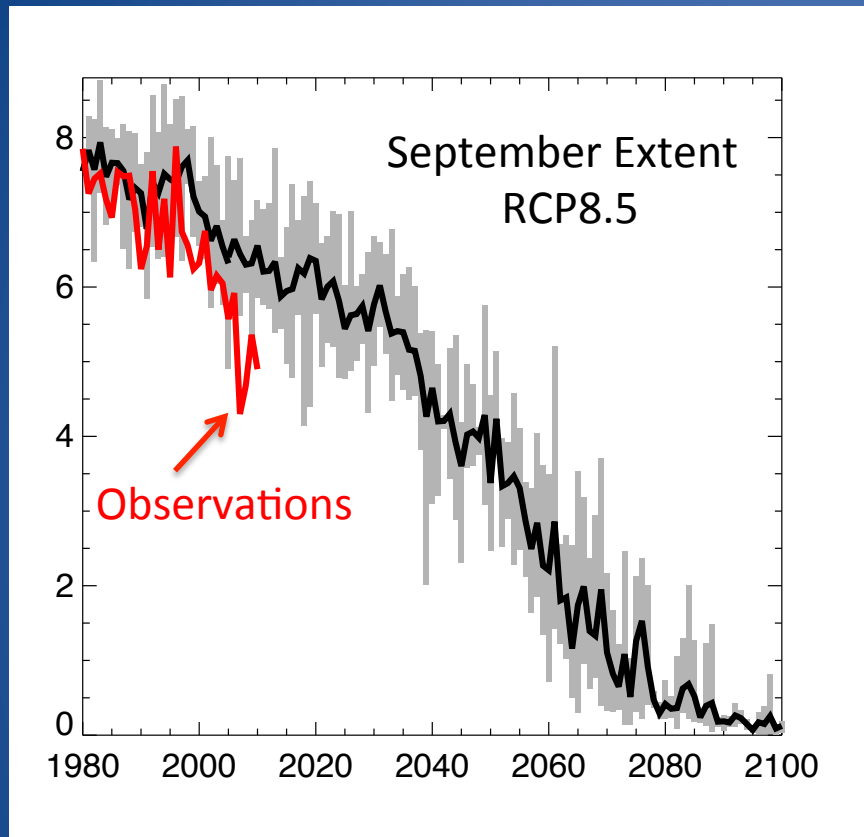
Vavrus, SJ, MM Holland, A Jahn, DA Bailey, BA Blazey, 2012: Twenty-First-Century Arctic Climate Change in CCSM4. J. Climate, 25, 2696–2710.
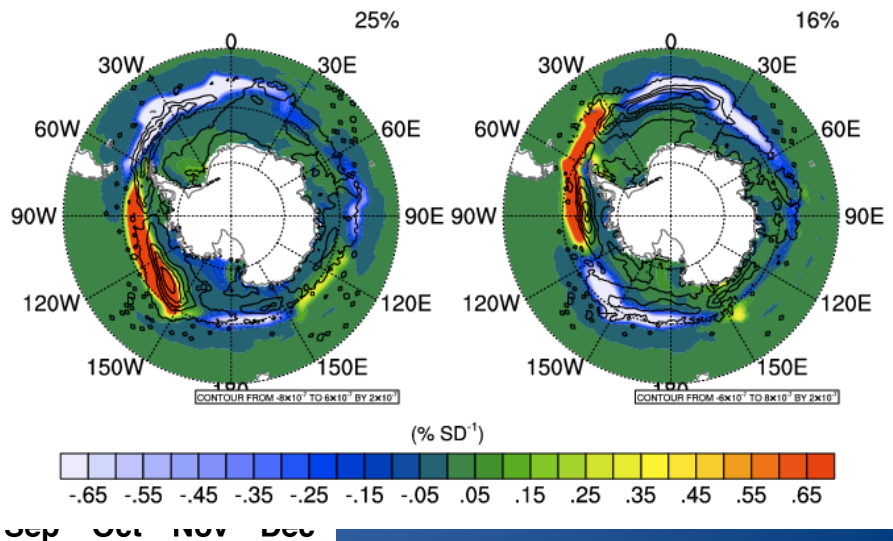
CCSM4/CESM1 Simulation of Antarctic sea ice
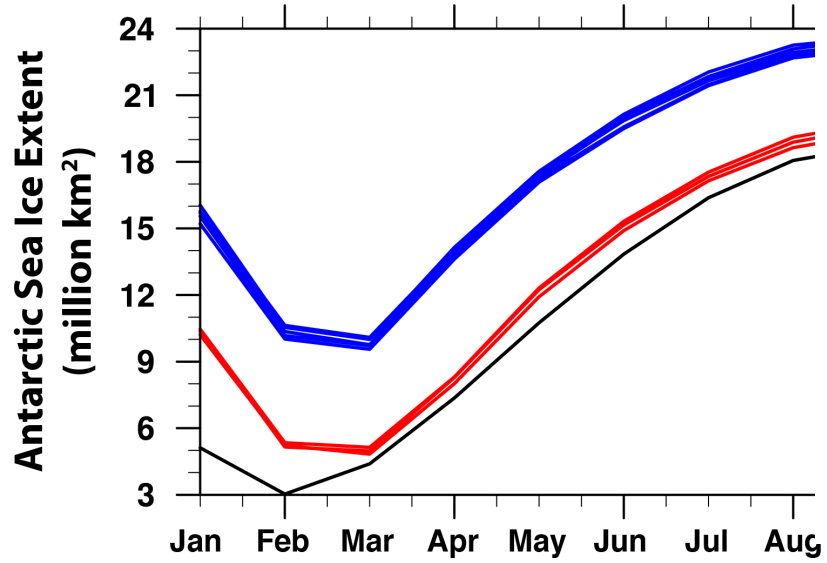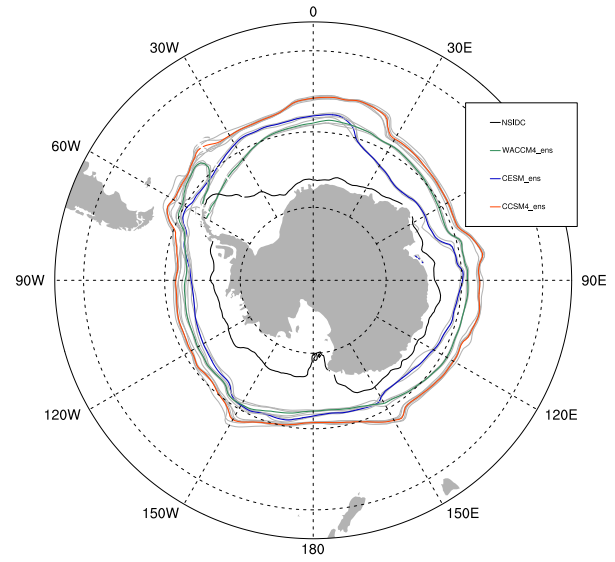
Summary

- CESM1 uses the Los Alamos CICE model
- This includes:
  - EVP dynamics,
  - thermodynamics that account for brine inclusions,
  - and a subgridscale ice thickness distribution.
- CCSM4 and CESM1 simulate very good Arctic sea ice overall.
- CCSM4 Antarctic sea ice is too extensive but variability in ice concentration looks realistic. CESM1 Antarctic sea ice is greatly improved.

Where are we heading?

- Prognostic salinity
- Biogeochemistry (Iron, Isotopes, Algae)
- More sophisticated melt pond modeling
- Snow model improvements
- Improved ice-ocean coupling

Much of this work is being done by collaborators at DOE Labs (primarily LANL) and Universities.

Simulated Ice Thickness Climatology

1980-1999
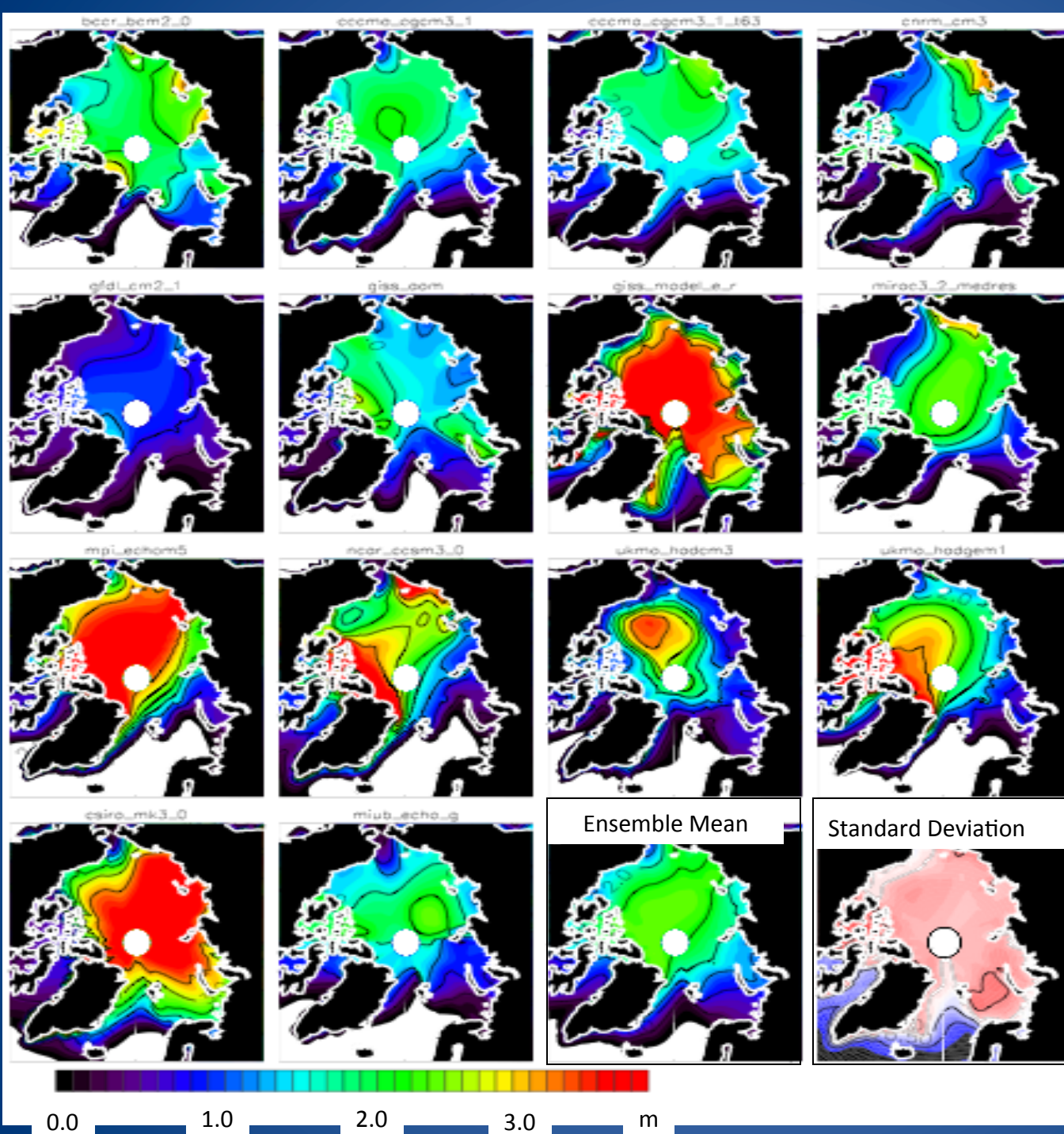
Thickness varies considerably across models

Differences in mean and distribution

Largest inter-model scatter is in the Barents Sea region

Assessing Sea Ice Mass Budgets

- Equilibrium Ice Thickness Reached when
  - Ice growth is balanced by ice melt + ice divergence
  - Illustrative to consider how different models achieve this balance and how mass budgets change over time

$$\frac{d\bar{h}}{dt} = \Gamma_h - \nabla \cdot (\vec{u}h)$$

Ice volume change   Thermodynamic source   Divergence
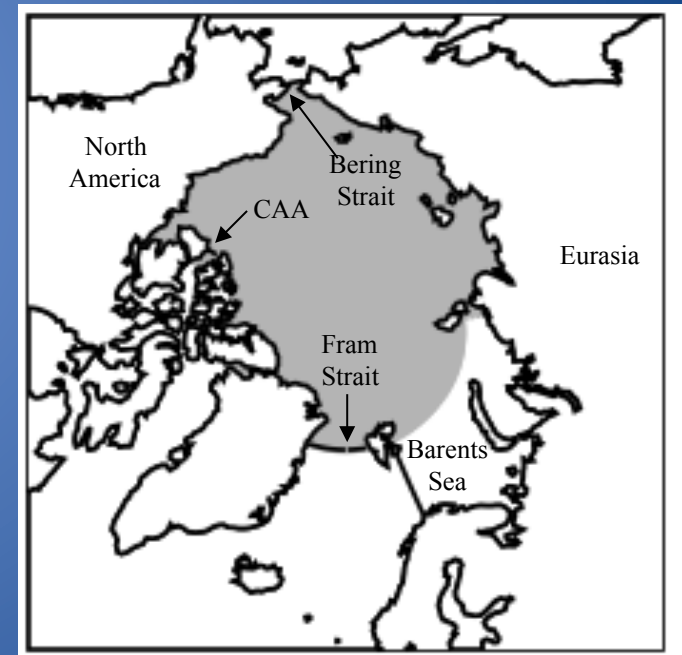
Climate model archive of monthly averaged ice thickness and velocity

Assess Arctic ice volume, transport through Arctic straits, and solve for ice growth/melt as residual

North America

Bering Strait

CAA

Eurasia

Fram Strait

Barents Sea

Holland et al., 2010

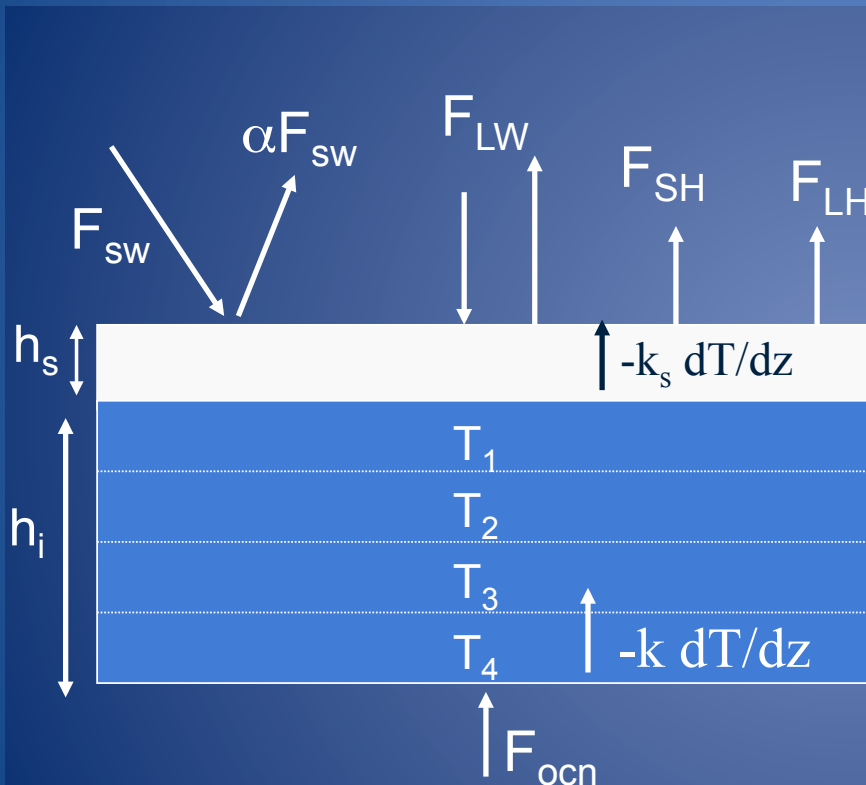# Sea ice loss is modified by climate feedbacks

- Fundamental sea ice thermodynamics gives rise to a number of important feedbacks

Balance of fluxes at surface

$$(1-\alpha)F_{SW} + F_{LW} - \sigma T^4 + F_{SH} + F_{LH}$$

$$+k\frac{\partial T}{\partial z} = -q\frac{dh}{dt}$$

αF_{sw} F_{LW} F_{SH} F_{LH}

F_{sw}

h_s

$-k_s \, dT/dz$

T_1

h_i

T_2

Vertical heat transfer (conduction, SW absorption)

T_3

T_4 $-k \, dT/dz$

F_{ocn}

Surface albedo changes modify SW absorption in ice and ocean heat flux
Ice loss lowers albedo – positive feedback

Balance of fluxes at ice base

$$F_{ocn} - k\frac{\partial T}{\partial z} = -q\frac{dh}{dt}$$

Ice mass budgets affected by climate feedbacks

- Fundamental sea ice thermodynamics gives rise to a number of important feedbacks

αF_{sw} F_{sw} F_{LW} F_{SH} F_{LH}

h_s

$-k_s \, dT/dz$

T_1
T_2

h_i

T_3
T_4 $-k \, dT/dz$

Balance of fluxes at surface

$$(1-\alpha)F_{SW} + F_{LW} - \sigma T^4 + F_{SH} + F_{LH}$$

$$+k\frac{\partial T}{\partial z} = -q\frac{dh}{dt}$$
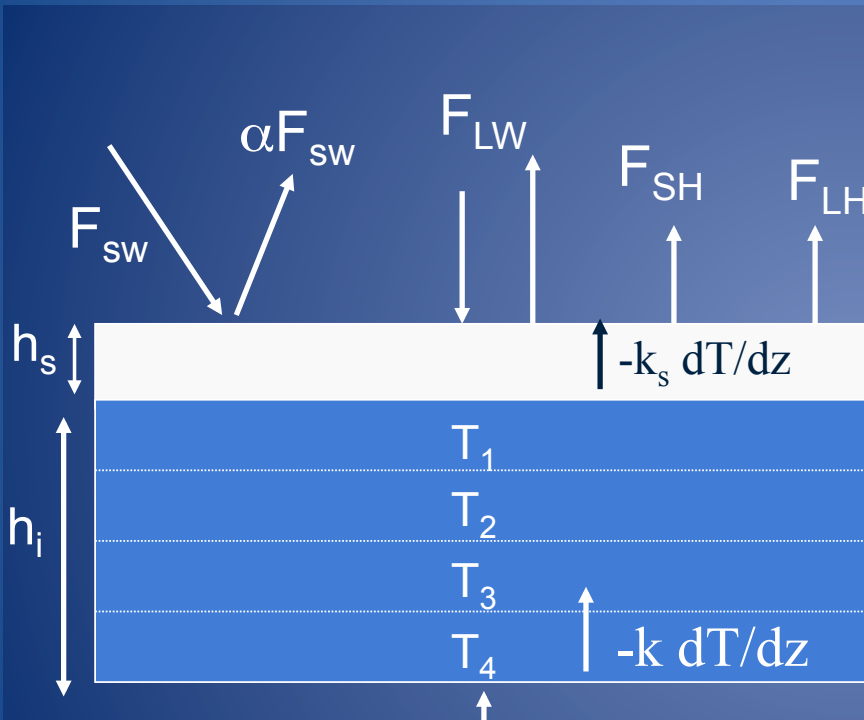
### Vertical heat transfer (conduction, SW absorption)

Balance of fluxes at ice base

$$F_{ocn} - k\frac{\partial T}{\partial z} = -q\frac{dh}{dt}$$

Heat conduction related to vertical temperature gradient
Causes ice growth to vary as 1/h
Has a stabilizing effect on ice thickness since thin ice grows more rapidly

# Evidence that model parameterizations influence feedback strength
## Enhanced albedo feedback in ITD run

NORTHERN HEMISPHERE
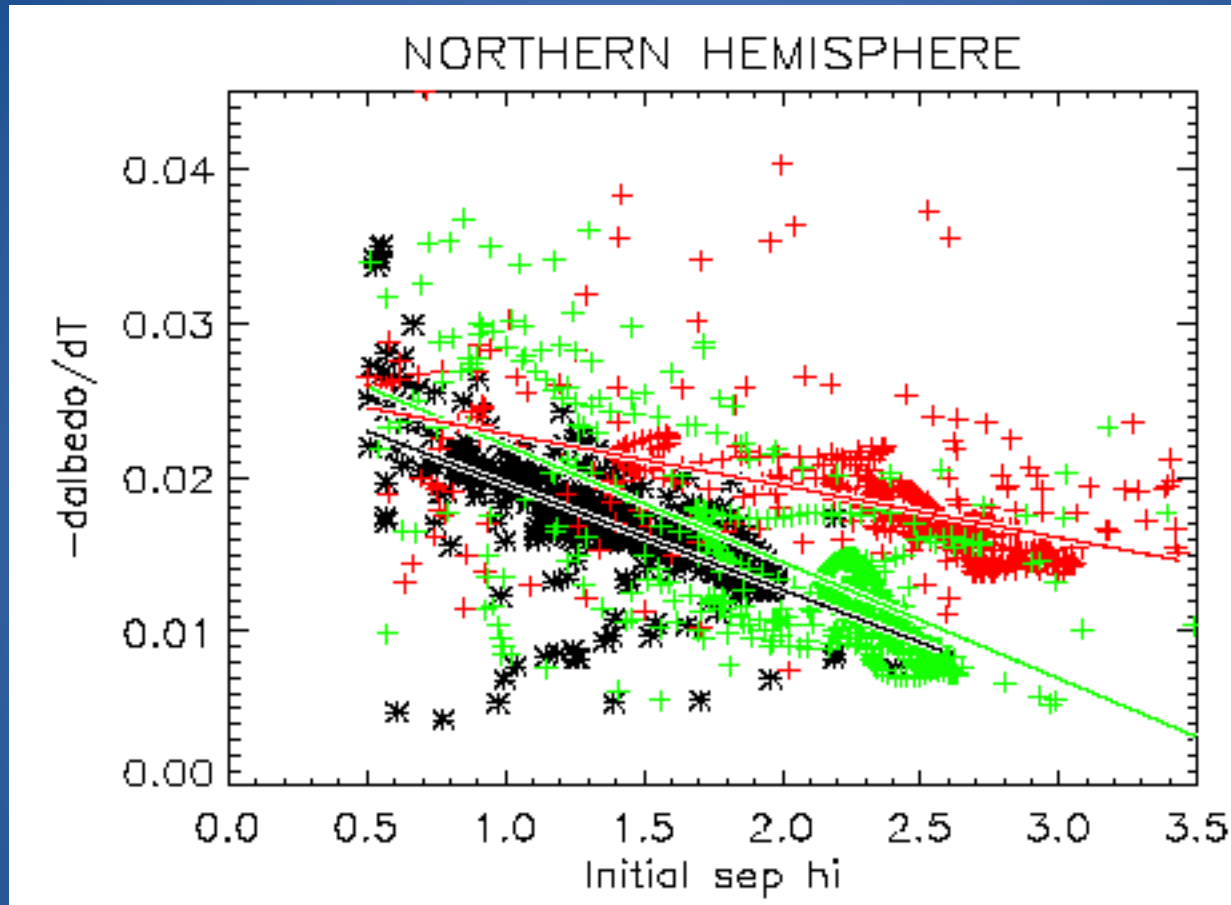
ITD (5 cat)
1 cat.
1cat tuned

Larger albedo change per temperature change for thinner initial ice
With ITD have larger a change for ice with same initial thickness
Suggests surface albedo feedback enhanced in ITD run

Holland et al., 2006

Model parameterizations modify ice growth rate feedback

NH Scatter Plot

5 category
1 category
1cat tuned
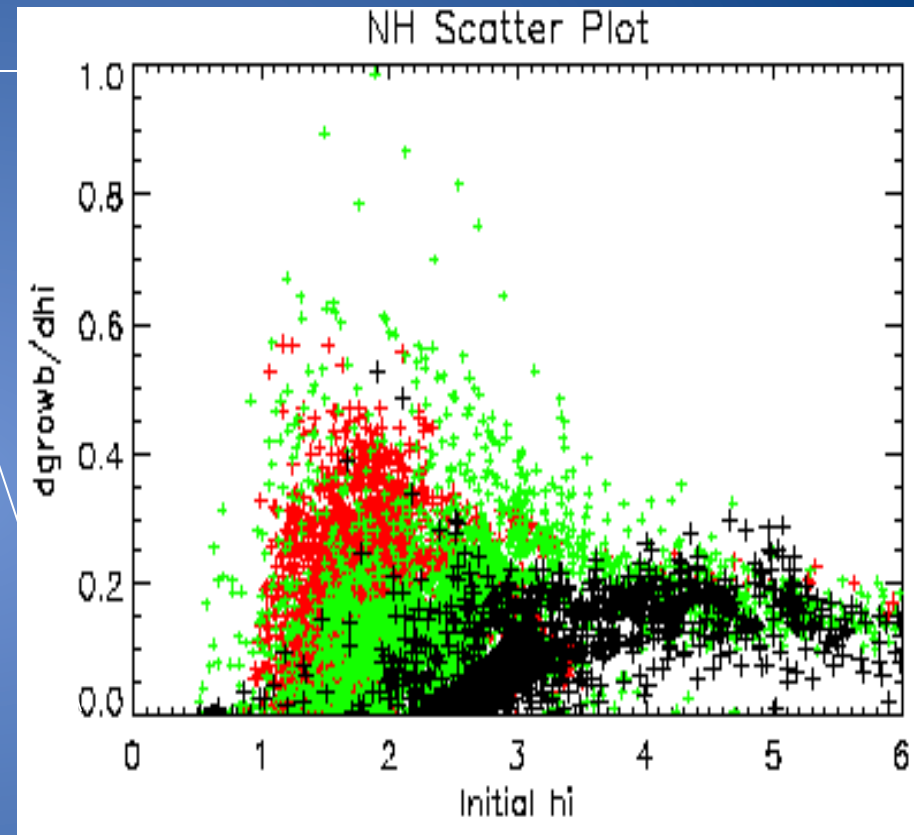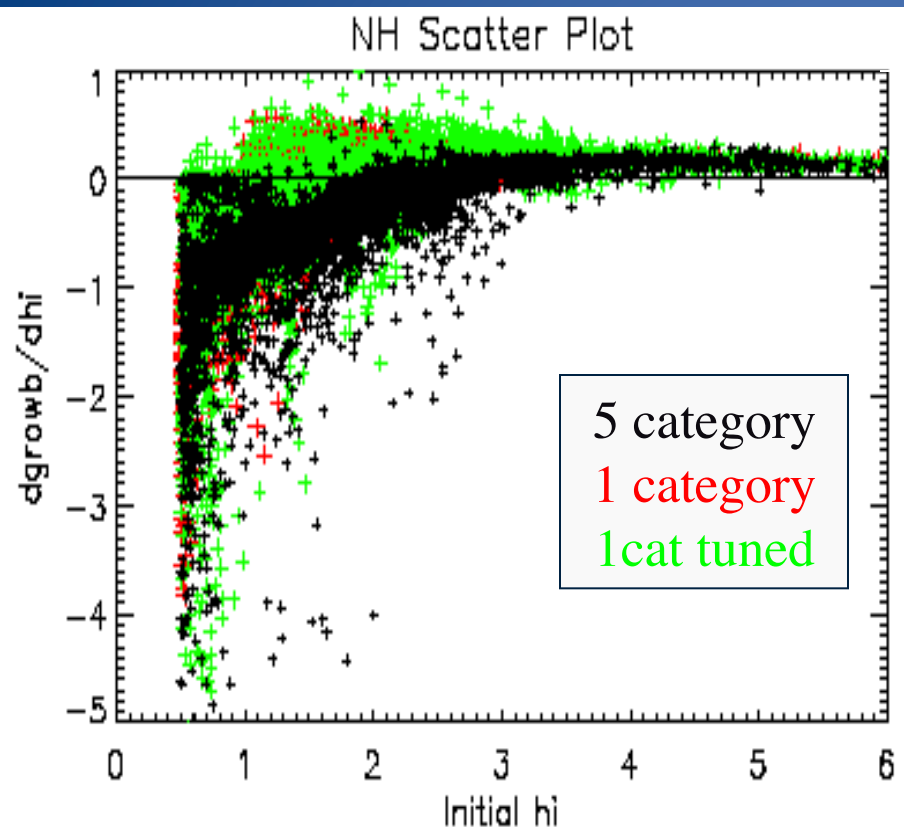
For ice of the same mean thickness,

• The ITD has fewer locations with increased ice growth.

• This suggests a reduced negative feedback on ice thickness

Challenges in Modeling Sea Ice in a Changing Environment

- So, is it all hopeless?

- Recent studies providing insight on what is needed if we are to accurately simulate sea ice change:
  - present day ice conditions, including extent and the spatial distribution of ice thickness;
  - the evolving surface energy budget

- To achieve this involves numerous and interacting factors across the coupled ice-ocean-atmosphere system

- Models are continuously improving and have provided considerable insight into the functioning of sea ice and its role in the climate system

Sea Ice Dynamics in climate models
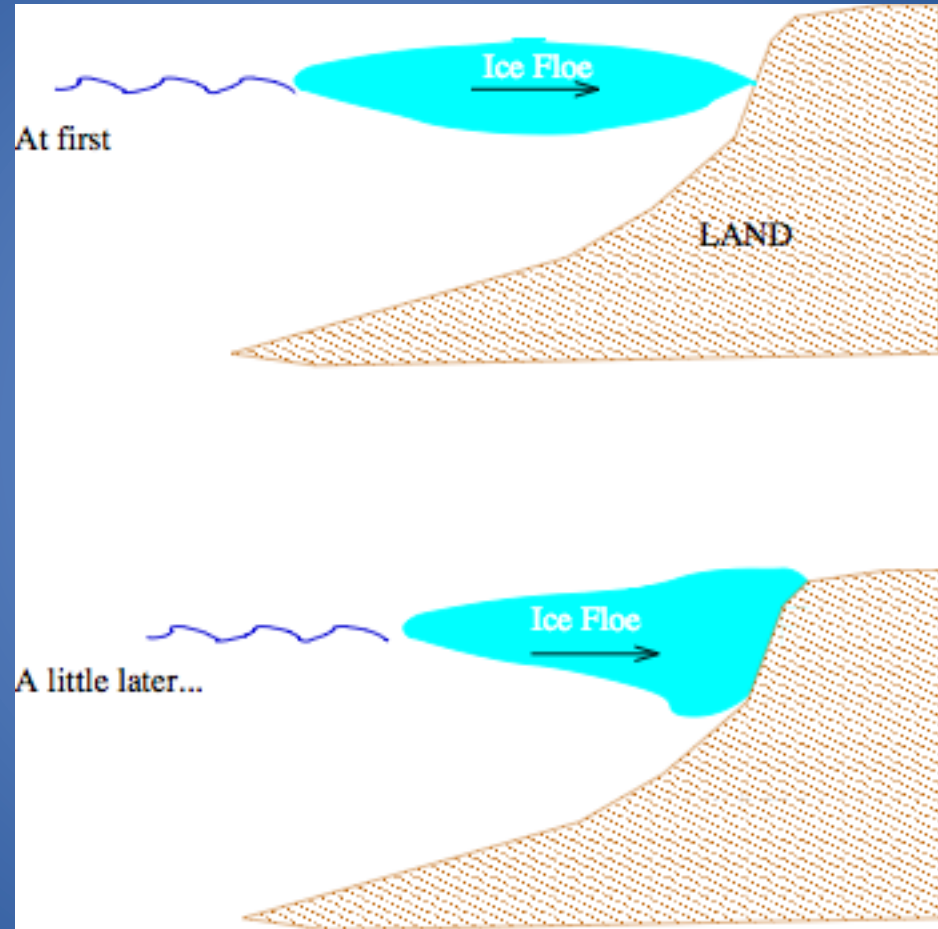
Past ad hoc method was to stop ice from moving at a critical thickness, sometimes called stoppage.

$$\frac{Dg}{Dt} = -g\nabla \cdot \mathbf{u} + \Psi - \frac{\partial}{\partial h}(fg) + \mathcal{L}$$

  1   2   3   4   5

1. Lagrangian time derivative of g following "parcel"

2. Convergence of parcel

3. Ψ = Mechanical redistribution

4. Ice growth/melt results in "advection of g in thickness space"

5. $\mathcal{L}$ = Reduction of g from lateral melt

> h = ice thickness
> $\mathbf{u}$ = ice velocity
> f = growth rate

Heat Equation used to find temperature T

$$\rho c \frac{\partial \mathsf{T}}{\partial \mathsf{t}} = \frac{\partial}{\partial \mathsf{z}} \mathsf{k} \frac{\partial \mathsf{T}}{\partial \mathsf{z}} + \kappa \mathsf{I}_0 \mathsf{e}^{-\kappa \mathsf{z}},$$

Untersteiner (1961) suggested the heat capacity of sea ice is

$$c(\mathsf{T}, \mathsf{S}) = \mathsf{c_o} + \frac{\gamma \mathsf{S}}{\mathsf{T}^2}$$

where T is in Celsius,

$$\gamma = \mathsf{L_o}\mu \quad \text{and} \quad \mathsf{T_m} = -\mu \mathsf{S}$$
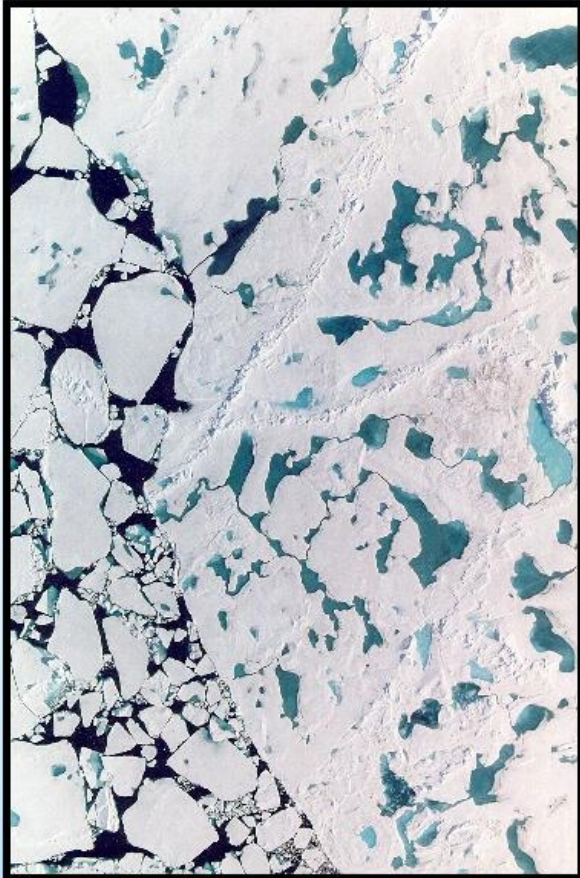
Ice Thickness Distribution
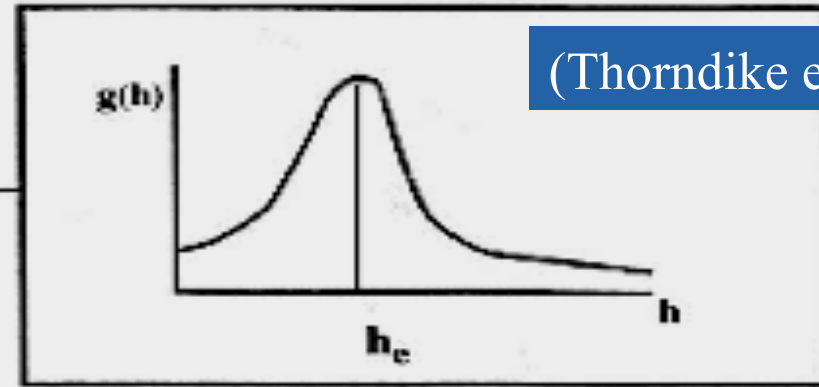
(Thorndike et al., 1975)
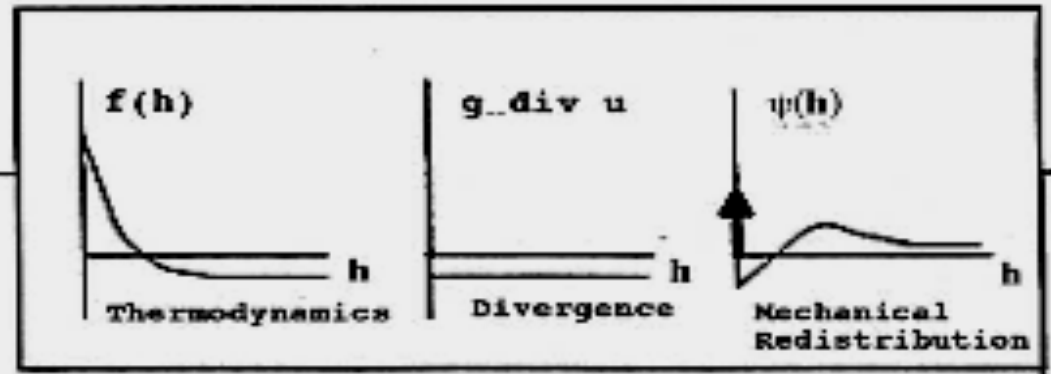
$$\frac{Dg}{Dt} = -\frac{\partial}{\partial h}(fg) + L(g) - \nabla \bullet (\vec{v}g) + \Psi(h,g,\vec{v})$$

Evolution depends on: Ice growth, lateral melt, ice divergence, and mechanical redistribution (riding/rafting)