Modeling ice/ocean interactions

Hélène SEROUSSI

Jet Propulsion Laboratory - California Institute of Technology
Introduction: ice sheets

Snow accumulation

Evaporation from ocean

Ice Stream within ice sheet

Icebergs

Sea ice

Ice shelf (floating in sea)

Calving Ice Front

Bedrock

Grounding line

Sea Level

Ocean

Ice sheet (resting on bedrock)

Ice sheet

Ice shelf

Sea Level
Jakobshavn Isbrae, West Greenland

Ice front retreat and glacier acceleration

Modeled forced with ice front position

BONDZIO ET AL. 3-D MODELING OF JI'S RETREAT

Bondzio et al., 2017
Jakobshavn Isbrae, West Greenland

Simulated ice front migration

Ensemble for unknow parameters

Bondzio et al., 2018
GRACE Observations of Antarctic Ice Mass Changes

Average Mass Loss:
125 Gigatons/year

Antarctic Ice Loss (meters water equivalent relative to 2002)
Outline

1. Modeling ice sheets and ice shelves
2. Ice shelves around Antarctica
3. Modeling ice shelf melt
4. Coupling ice and ocean models
5. Can we parameterize ice shelf melt?
Mass conservation

Continuity equation: \[
\frac{D\rho}{Dt} + \rho \nabla \cdot \mathbf{v} = 0
\]

**Incompressibility:** a continuum is said to be incompressible if its density remains unchanged during motion

\[
\frac{D\rho}{Dt} = 0
\]

Mass balance of incompressible fluids:

\[
\nabla \cdot \mathbf{v} = \frac{\partial v_x}{\partial x} + \frac{\partial v_y}{\partial y} + \frac{\partial v_z}{\partial z} = 0
\]
Mass conservation

Incompressibility:

\[ \nabla \cdot \mathbf{v} = \frac{\partial v_x}{\partial x} + \frac{\partial v_y}{\partial y} + \frac{\partial v_z}{\partial z} = 0 \]

Surface evolution:

\[
\frac{\partial s}{\partial t} + v_x(s) \frac{\partial s}{\partial x} + v_y(s) \frac{\partial s}{\partial y} - v_z(s) = \dot{M}_s
\]

\[
\frac{\partial b}{\partial t} + v_x(b) \frac{\partial b}{\partial x} + v_y(b) \frac{\partial b}{\partial y} - v_z(b) = \dot{M}_b
\]

- \( s \) glacier surface elevation (m)
- \( b \) glacier base elevation (m)
- \( \dot{M}_s \) surface mass balance (m/s ice equivalent, positive when accumulation)
- \( \dot{M}_b \) basal mass balance (m/s ice equivalent, positive when melting)
Energy balance

Conservation of energy:

\[ \rho \frac{D}{Dt} (cT) = \nabla \cdot k_{th} \nabla T + \Phi \]

- \( T \) ice temperature (K)
- \( c \) ice thermal conductivity (W/m/K)
- \( k_{th} \) ice heat capacity (J/K/kg)
- \( \Phi = \sigma : \varepsilon \) deformational heating (W)

Ice energy balance:

\[ \frac{\partial T}{\partial t} = -\mathbf{v} \cdot \nabla T + \frac{k_{th}}{\rho c} \Delta T + \frac{\Phi}{\rho c} \]

Phase change included by capturing cold/temperate transition or using enthalpy formulations
Momentum balance

Conservation of momentum:

\[
\frac{\partial \mathbf{v}^*}{\partial t^*} + (\mathbf{v}^* \cdot \nabla^*) \mathbf{v}^* = \rho^* \left( \frac{\partial \mathbf{v}}{\partial t} + \nabla \cdot (\mathbf{v} \nabla) \right) + \frac{1}{Fr} \rho^* \nabla \cdot \mathbf{g} + \frac{1}{Ro} \rho^* \mathbf{g} \times 2\rho \Omega \times \mathbf{v}
\]

<table>
<thead>
<tr>
<th>Variable</th>
<th>Glacier</th>
<th>Ice sheet</th>
<th>Ice stream</th>
</tr>
</thead>
<tbody>
<tr>
<td>(V_0)</td>
<td>(10^{-6})</td>
<td>(10^{-5})</td>
<td>(10^{-4})</td>
</tr>
<tr>
<td>(G_0)</td>
<td>10</td>
<td>10</td>
<td>10</td>
</tr>
<tr>
<td>(R_0)</td>
<td>(10^4)</td>
<td>(10^6)</td>
<td>(10^5)</td>
</tr>
<tr>
<td>(\Omega_0)</td>
<td>(10^{-4})</td>
<td>(10^{-4})</td>
<td>(10^{-4})</td>
</tr>
<tr>
<td>(\sigma_0)</td>
<td>(10^5)</td>
<td>(10^5)</td>
<td>(10^5)</td>
</tr>
<tr>
<td>(\rho_0)</td>
<td>(10^3)</td>
<td>(10^3)</td>
<td>(10^3)</td>
</tr>
<tr>
<td>(T_0)</td>
<td>(\frac{R_0}{V_0})</td>
<td>(\frac{R_0}{V_0})</td>
<td>(\frac{R_0}{V_0})</td>
</tr>
</tbody>
</table>

\[
St = \frac{R_0}{T_0 V_0}
\]

\[
Re = \frac{\rho_0 V_0^2}{\sigma_0}
\]

\[
Fr = \frac{V_0^2}{R_0 G_0}
\]

\[
Ro = \frac{V_0}{2\Omega_0 R_0}
\]

Stokes flow: \(\nabla \cdot \mathbf{\sigma} + \rho \mathbf{g} = 0\)
Momentum balance

Incompressible viscous fluid: \( \sigma' = 2\mu \varepsilon \)

Glen’s flow law (1955):
\[
\mu = \frac{B}{2\varepsilon_e^{1-1/n}}
\]

Boundary conditions:
- Ice/air interface: free surface \( \sigma \cdot n \simeq 0 \)
- Ice/ocean interface: water pressure \( \sigma \cdot n = P_w n \)
- Ice/bedrock interface:
  \[
  (\sigma \cdot n + \alpha^2 \mathbf{v})_\parallel = 0 \quad \mathbf{v} \cdot n = -M_b n_z
  \]

Shallow aspect ratio: Shallow ice approximations (shallow ice and shallow shelf) to separate horizontal and vertical motion
Grounding line or grounding zone?

F: landward limit of ice flexure from tidal movement
G: limit of ice floatation (grounding line)
I_b: break-in slope
I_m: local elevation minimum
H: seaward limit of ice flexure from tidal movement

Rignot et al., 2009
Grounding line migration

Contact problem

- Full-Stokes stress balance
- Boundary condition:
  - Ice/bedrock if:
    \[ z_b(x, t) = b(x) \quad \text{and} \quad -\sigma_{nn}|_b > \rho_w(z_b, t), \]
  - Ice/water if:
    \[ z_b(x, t) > b(x), \]
  or
  \[ z_b(x, t) = b(x) \quad \text{and} \quad -\sigma_{nn}|_b \leq \rho_w(z_b, t), \]
- Very high resolution required in the grounding line area

Hydrostatic assumption

- Simplified stress balance
- Hydrostatic condition:
  
  Hydrostatic thickness:
  \[ \frac{\rho_w}{\rho_i} r, \quad r < 0, \]
  \[ H > H_f \quad \text{ice is grounded}, \]
  \[ H = H_f \quad \text{grounding line position}, \]
  \[ H < H_f \quad \text{ice is floating}. \]
Observations

Input parameters

- Surface/bed topography
- Surface temperatures
- Basal friction
- ...

Numerical model

Model output

- Ice temperature
- Mass balance
- Surface velocities
- ...

Physical model

Energy balance

- Heat transfer

\[
\frac{\partial T}{\partial t} = -\nabla \cdot \nabla T + \frac{k_{th}}{\rho c} \Delta T + \frac{\Phi}{\rho c}
\]

Stress balance

- Incompressible Stokes flow

\[
\nabla \cdot \sigma' - \nabla P + \rho g = 0
\]

Mass balance

- Incompressibility

\[
\frac{\partial H}{\partial t} = -\nabla \cdot H \vec{v} + \dot{M}_s - \dot{M}_b
\]
Antarctic ice flow

Basal friction

Morlighem et al., 2013
Rignot et al., 2011
Outline

1. Modeling ice sheets and ice shelves
2. Ice shelves around Antarctica
3. Modeling ice shelf melt
4. Coupling ice and ocean models
5. Can we parameterize ice shelf melt?
Ice shelves around Antarctica

- Limited direct observations
- Melt rate estimates:
  \[ \frac{\partial H}{\partial t} + \nabla \cdot H \mathbf{v} = \dot{M}_s - \dot{M}_b \]
- Equal contribution of calving and melting (∼1300 Gt/yr)
- Similar results in Depoorter et al. (2013)
- Variety of ice shelves (size, melt rate, calving rate, …)

Rignot et al., 2013
Ice shelf buttressing

Fuerst et al., 2016
Larsen B breakup: a natural experiment

Larsen B break-up in 2002

Figure 1. (a) Location map of the Antarctic Peninsula including locations of Larsen C and George VI ice shelves and localities mentioned in the text.

(b) Bedrock elevations below sea level in metres for the Antarctic Peninsula from BEDMAP2 (Fretwell et al., 2013). The colour bar is truncated at 0 m. Red inset rectangles delineate the locations of zoomed-in areas in Figs. 2, 5, 6, and 8. Black polygons denote ice-sheet model domains.
Larsen B breakup: a natural experiment

Scambos et al., 2004
Impact of ice shelf melt

Change in grounding line flux for a 1 m thinning over 20 x 20 km$^2$

Reese et al., 2016
Observed changes in the Amundsen Sea

Elevation change

2003 – 2009

m/yr

0.6
0.3
0.0
-0.3
-0.6
-0.9
-1.2
-1.5
-1.8
-2.1
-2.4
-2.7
-3.0

Sutterley et al., 2014

Acceleration

(b)

Grounding line retreat

Mouginot et al., 2014

Rignot et al., 2014

ECCO Summer School 2019 - Friday Harbor
Outline

1. Modeling ice sheets and ice shelves
2. Ice shelves around Antarctica
3. Modeling ice shelf melt
4. Coupling ice and ocean models
5. Can we parameterize ice shelf melt?
Ice/ocean interactions

Greenland tidewater glacier

- Near vertical face
- Large amount of subglacial runoff with strong seasonal signal
- Small systems (1 kms)

Antarctic Ice shelf

- Near horizontal face
- Limited amount of subglacial runoff with no seasonal signal
- Large systems (100 kms)
Southern Ocean

Rignot et al., 2013

Schmidtko et al., 2014

Downloaded from www.sciencemag.org on June 13, 2013
Cold ice shelves

- Dense Shelf Water dominates in sub-ice cavity
- Shelf Water has temperature close to the surface freezing point
- Brine rejection during sea ice growth
- Pressure dependence of the freezing point so melt at depth
- Refreezing occurs as water produced by melting becomes supercooled as it rises
- Ross/Weddell Sea

Jenkins et al., 2016
Warm ice shelves

- No Shelf Water
- Circumpolar Deep Water densest water on the shelf
- Circumpolar Deep Water around 3°C above the surface freezing
- Rapid melting
- No refreezing
- Amundsen/Bellingshausen Seas

Jenkins et al., 2016
Varying ocean conditions

**Dutrieux et al., 2014**
Ice shelf melt from an ocean model

Three equations model (Jenkins et al., 2010)

- Heat balance at the phase change interface

\[ \rho_i m L_i = \rho_i c_i \kappa_i \frac{\partial T_i}{\partial z} \bigg|_b - \rho_w c_w \gamma_T (T_f - T_w) \]

- Freezing point of sea water

\[ T_f = aS_b + b + cz_b \]

- Salt balance at the phase change interface

\[ \rho_i m (S_b - S_i) = -\rho_w \gamma_S (S_b - S_w) \]

- Velocity dependent heat and salt exchange coefficients

\[ \gamma_T = \Gamma_T \sqrt{C_d (u^2_b + u^2_{tide})} \quad \gamma_S = \Gamma_S \sqrt{C_d (u^2_b + u^2_{tide})} \]
Ocean circulation

Schodlok et al., 2012
Varying ocean conditions

Melt spatially and temporally variable: example of Pine Island ice shelf

Schodlok et al., 2012
Impact of unknown coefficients

Figure 9. (a) Melt rate (m/yr) simulated using the velocity-dependent model in the realistic PIIS setup and $C_{d_5}/C_{d_{0}} = 3.10^{2}$. The maximum and area-averaged melt rates are indicated in the top right corner of the figure. For this value of drag coefficient, the area-averaged melt rate is comparable to the ice flux divergence based estimate of Payne et al. [2007] (20.7 m/yr). (b) Difference between the velocity-dependent melt rate simulated using $C_{d_5}/C_{d_{0}}$ and $C_{d_{0}}$. Positive differences indicate a higher melt rate for the larger drag coefficient experiment. The maximum and minimum differences are indicated in the top right corner of the figure. Black contours indicate the depth of the ice shelf base (m) on both figures.

Figure 10. Melt rate (m/yr) simulated using (a–c) the velocity-independent and (d–f) the velocity-dependent model in the realistic PIIS setup with $C_{d_{0}}$ and a (a and d) 10 m, (b and e) 20 m, and (c and f) 50 m thick mixed layer for averaging of $T_M$, $S_M$, and $U_M$. Dashed contours show the distribution of water column thickness (m). The maximum and area-averaged melt rates are indicated in the top right corner of each plot.

Journal of Geophysical Research: Oceans
10.1002/2013JC008846
Dansereau et al., 2014
Outline

1. Modeling ice sheets and ice shelves
2. Ice shelves around Antarctica
3. Modeling ice shelf melt
4. Coupling ice and ocean models
5. Can we parameterize ice shelf melt?
Ice dynamics sensitive to ocean melting
(Joughin et al., 2012, 2014; Favier et al., 2014; Seroussi et al., 2014)

Basal melting sensitive to cavity shape
(Goldberg et al., 2012; Schodlok et al., 2012)

Favier et al., 2014
Schodlok et al., 2012
Coupled ice/ocean simulations

- Interpolation between grids
- Timescales
- Evolution of modeled domain
Ice domain:
- ALE
- Horizontal layers follow topography

Ocean domain:
- Fixed grid
- Remeshing
- Add/remove cells

Goldberg et al., 2018
Idealized case of Pine Island Glacier

Complex grounding line retreat from a seabed ridge

De Rydt and Gudmundsson, 2016
Simulation of Thwaites Glacier

ISSM-MITgcm simulations:
• 50 year simulations
• 1 month coupling
• 500 m resolution at GL (sub-element parameterization)
• 2 km ocean model
• 5 year spin-up of ocean with fixed cavity shape

Experiments:
• 1992 forcing (ECCO)
• 1992 + 0.5°C forcing
• Uncoupled

Seroussi et al., 2017
Simulated melt rates:
- Pine Island: 88 Gt/yr
- Thwaites: 81 Gt/yr
- Cosgrove: 37 Gt/yr
- Dotson/Crosson: 24 Gt/yr
Evolution of Thwaites Glacier
Comparison with observations

Ice shelf melt

97.5 ± 7 Gt/yr (Rignot et al., 2013)

69 ± 18 Gt/yr (Depoorter et al., 2013)

Ocean induced melting (Gt/yr)

Volume change (mm s.l.e.)

Time (yr)

Grounding line

Bedrock elevation (m)

Time (yr)
Comparison with parameterized melt

Initial melt

Melt parameterization

- Basal melting rate (m/yr)
- Ice shelf base depth (m)
- Water column thickness (m)

Model data
Linear fit

$R^2 = 0.87$
Uncoupled simulation (1992UC)
Grounding line evolution

1992F

1992F+

1992UC

Time (yr)

Bedrock elevation (m)

20 km
Outline

1. Modeling ice sheets and ice shelves
2. Ice shelves around Antarctica
3. Modeling ice shelf melt
4. Coupling ice and ocean models
5. Can we parameterize ice shelf melt?
Parameterizations of ocean conditions

- Depth parameterization

- Quadratic local dependence on thermal forcing
  \[
  \dot{m} = \gamma_T \left( \frac{\rho_w c_{po}}{\rho_i L_i} \right)^2 (T_o - T_f)^2
  \]
  Holland et al. 2008

- Quadratic local/non local dependence on thermal forcing
  \[
  \dot{m} = \gamma_T \left( \frac{\rho_w c_{po}}{\rho_i L_i} \right)^2 \langle T_o - T_f \rangle (T_o - T_f)
  \]
  Favier et al. GMDD 2019

- PICO (Potsdam Ice-shelf Cavity mOdel): Box model

- PICOp (PICO + plume model)

Reese et al. 2017

Pelle et al. 2019
Reese et al., 2018

\[ q (T_{k-1} - T_k) - A_k m_k \frac{\rho_i}{\rho_w} \frac{L}{c_p} = 0 \]

\[ q (S_{k-1} - S_k) - A_k m_k S_k = 0 \]

- \( T_k \) Temperature of \( B_k \)
- \( A_k \) box surface area
- \( m_k \) melt rate of \( B_k \)
- \( q = C (\rho_0 - \rho_1) \) strength of the overturning circulation
Comparison of ice shelf melt rates

Depth-Dependent (Modeled)
- Pine Island: 18.24 m/yr
- Thwaites: 17.62 m/yr
- Dotson: 17.98 m/yr

Quadratic (Modeled)
- Pine Island: 18.96 m/yr
- Thwaites: 17.94 m/yr
- Dotson: 15.72 m/yr

PICO (Modeled)
- Pine Island: 9.29 m/yr
- Thwaites: 7.29 m/yr
- Dotson: 13.32 m/yr

PICOP (Modeled)
- Pine Island: 10.25 m/yr
- Thwaites: 11.60 m/yr
- Dotson: 4.53 m/yr

Observations
- Pine Island: 16.02 ± 1 m/yr
- Thwaites: 17.73 ± 1 m/yr
- Dotson: 7.80 ± 0.6 m/yr
Summary

• Ice is a laminar viscous incompressible material
• Ice/ocean interactions are driving most of the dynamic changes observed in the Amundsen Sea (and elsewhere)
• Coupled ice-ocean model:
  • produce more realistic estimates of glacier retreat rates than ice model driven by parameterized melt
  • limited observations to constrain and validate models
• Ice sheets starting to be included in Earth System models mostly for ice/atmosphere coupling, not ocean (need ocean cavities)
Questions ?