Sea ice thickening (& thinning) processes: Dynamics & Thermodynamics

Diagram showing various forces and processes involved in sea ice dynamics, including wind forcing, Coriolis force, inertial force, form drag, frictional drag, internal stress/forces, large-scale rheology, long-wave flux, turbulent flux, shortwave flux, conductive heat flux, ocean currents, sea-surface slope, and solar heating of ice and ocean (primary production by phytoplankton).
Surface energy balance

\[(1 - \alpha)F_r - I_0 + F_L - F \uparrow + F_s + F_e + F_c + F_m = 0\]

\[F_c + F_w + \rho_i L \frac{dH}{dt} = 0\]
Short- & longwave radiation balance

- Fraction \((1-i_0)\) of shortwave spectrum (near-IR) absorbed in uppermost 0.1 m of ice, contributing directly to surface heating

\[
F(z) = i_0 (1 - \alpha) F_r \exp[-\kappa_e (z - 0.1)]
\]

- \(i_0\) typically ranges between 0.18 (clear sky) and 0.35 (overcast)

Fig. 2.3.2.1: Downwelling shortwave flux at Barrow, AK for clear (solid curve) and heavily overcast (dashed curve conditions) based on Perovich and Grenfell (1984).
The Surface Heat Budget of the Arctic (SHEBA) experiment

Persson et al., 2002
Shortwave radiation balance & albedo

Persson et al., 2002
Albedo annual cycle

Persson et al., 2002
Longwave radiation balance

- a surface with the physical surface temperature $T_0$ emits a radiative flux $F\uparrow$ proportional to the fourth power of $T_0$:

$$F \uparrow = \varepsilon \sigma T_0^4$$

with $\varepsilon$ the emissivity of the surface and $\sigma$ the Stefan-Boltzmann constant ($= 5.67 \times 10^{-8} \text{ W m}^{-2} \text{ K}^{-4}$)

- Wien's displacement law that relates the physical temperature to the wavelength at which emission is at a maximum

$$\lambda_{\text{max}} T_0 = 2898 \mu m K^{-1}$$

explains why the solar shortwave flux peaks in the visible range of the spectrum at roughly $0.5 \mu m$ whereas the earth’s surface radiates at roughly $12 \mu m$ in the polar regions (Fig. 2.3.2.3)
- Blackbody emission curves for the Earth’s and Sun’s surface (a), absorption spectrum of the Earth’s atmosphere (b, c) and contribution of different (vapor) molecules to absorptivity (d)
Persson et al., 2002

Chen et al., 2003
Longwave radiation

Huwald et al., 2005
Curry et al., 1993
Net radiation balance

Figure 20. Surface energy budget using daily means. The terms are (a) $F_\text{tot}$ (solid), the cumulative mean of $F_\text{tot}$ beginning on November 1, 1997 (dashed), and $Q^*$ (dotted); (b) $Q_s$ (light solid), $Q_I$ (heavy solid), and $\alpha$ (dotted); and (c) $H_s$ (solid), $H_{ib}$ (dotted), and $C$ (dashed) using $k_s = 0.3$ W m$^{-1}$ K$^{-1}$.

Persson et al., 2002
Turbulent heat fluxes

- Sensible and latent heat fluxes are determined by wind speed (vertical component) and temperature or humidity gradient in the boundary layer.

\[ F_s = \rho c_p C_s u_z (T_a - T_0) \]
\[ F_e = \rho L C_e u_z (q_a - q_0) \]

Persson et al., 2002
Turbulent heat fluxes

- Sensible and latent heat fluxes are determined by wind speed (vertical component) and temperature or humidity gradient in the boundary layer

\[ F_s = \rho c_p C_s u_z (T_a - T_0) \]
\[ F_e = \rho L C_e u_z (q_a - q_0) \]

- Boundary-layer theory allows for derivation of velocity profile as a function of height above the surface under certain conditions of atmospheric stability (and for a given surface roughness; details to follow)
Conductive heat flux

- Conductive flux through sea ice also a strong function of snow depth and snow thermal conductivity

\[ F_c(z) = -λ \left( \frac{∂T}{∂z} \right)_z \]

- Solar heating of interior ice and phase changes can be thought of as source terms in heat transfer equation

\[ \left( \rho c \right)_i \left( \frac{∂T}{∂t} \right)_z = k_i \left( \frac{∂^2 T}{∂z^2} \right)_z + I_0 κ_e \exp(-κ_e z) \]

- Latent heat of freezing/melting at surface final term in surface heat budget

\[ F_m = \left[ \rho L \frac{d(H + h)}{dt} \right]_0 \]

Fig. 2.3.4.1: Conductive heat flux through perennial ice in winter and spring as a function of snow depth (Maykut, 1986).
Conductive heat flux

Persson et al., 2002
Balance of fluxes at lower surface of ice cover

\[(1 - \alpha)F_r - I_0 + F_L - F \uparrow + F_s + F_e + F_c + F_m = 0\]

\[F_c + F_w + \rho_i L \frac{dH}{dt} = 0\]
Fig. 2.3.4.2: Oceanic heat flux at the bottom of the ice derived from water temperatures and heat transfer estimates (top). Solar shortwave radiation absorbed in the surface layer of leads (bottom, solid line) and below the 3-m depth level (dashed line) according to Maykut and McPhee (1995).
Ocean heat flux in the Arctic

Fig. 2.3.4.3: Equilibrium thickness of perennial ice in the Arctic as a function of oceanic heat flux (Maykut, 1986).

Fig. 2.3.4.5: Pattern of perennial ice growth in the Arctic basin (Maykut, 1986).
Ocean heat flux in the Antarctic

Fig. 2.3.4.3: Equilibrium thickness of perennial ice in the Arctic as a function of oceanic heat flux (Maykut, 1986).

Fig. 2.3.4.4: Thickness evolution of Weddell Sea ice as a function of oceanic heat flux (snow accumulation 0.2 m/a) as simulated with an ice growth model.)
Bottom-melt episodes at SHEBA camp
Heat content of surface waters:
- above 50 m depth sufficient to melt between 0.1 and 0.5 m of ice
Surface energy balance

Figure 20. Surface energy budget using daily means. The terms are (a) $F_{\text{tot}}$ (solid), the cumulative mean of $F_{\text{tot}}$ beginning on November 1, 1997 (dashed), and $Q^*$ (dotted); (b) $Q_s$ (light solid), $Q_l$ (heavy solid), and $\alpha$ (dotted); and (c) $H_s$ (solid), $H_{ib}$ (dotted), and $C$ (dashed) using $k_s = 0.3 \, W \, m^{-1} \, K^{-1}$.

Persson et al., 2002
Figure 21. As for Figure 20, but using monthly means. In panels (a) and (c), values of $F_{\text{tot}}$, $F_{\text{tot-run_mean}}$, and $C$ using the conductivity fluxes with $k_0 = 0.14$ W m$^{-1}$ K$^{-1}$ are also shown with a light line.
<table>
<thead>
<tr>
<th>Parameter</th>
<th>SHEBA</th>
<th>SHEBA (\alpha_{\text{July}} = 0.64)</th>
<th>B66</th>
<th>MU71</th>
<th>M82 (3\text{m})</th>
<th>M82 (0.8-\infty \text{ m})</th>
<th>EC93</th>
<th>L98</th>
</tr>
</thead>
<tbody>
<tr>
<td>(Q_{\text{sl}})</td>
<td>91.88</td>
<td>91.88</td>
<td>98.1</td>
<td>100.04</td>
<td>99.85</td>
<td>99.85</td>
<td>101.3</td>
<td>96.83</td>
</tr>
<tr>
<td>(Q_{\text{s}})</td>
<td>23.36</td>
<td>21.77</td>
<td>23.5</td>
<td>24.16</td>
<td>23.74</td>
<td>23.74</td>
<td>29.49</td>
<td>23.23</td>
</tr>
<tr>
<td>(Q_{\text{li}})</td>
<td>230.80</td>
<td>230.80</td>
<td>211.5</td>
<td>220.22</td>
<td>220.35</td>
<td>220.35</td>
<td>215.34</td>
<td>219.31</td>
</tr>
<tr>
<td>(Q_{\text{t}})</td>
<td>-21.32</td>
<td>-21.32</td>
<td>-24.4</td>
<td>-24.42</td>
<td>-20.66</td>
<td>-22.06</td>
<td>-28.41</td>
<td>-22.74</td>
</tr>
<tr>
<td>(H_{\text{s}})</td>
<td>-2.20</td>
<td>-2.20</td>
<td>-1.3</td>
<td>-3.58</td>
<td>-3.59</td>
<td>-3.59</td>
<td>-1.84</td>
<td>-3.01</td>
</tr>
<tr>
<td>(H_{\text{fb}})</td>
<td>1.06</td>
<td>1.06</td>
<td>0.9</td>
<td>4.25</td>
<td>4.11</td>
<td>3.49</td>
<td>1.55</td>
<td>2.32</td>
</tr>
<tr>
<td>(Q_{\text{s}} + Q_{\text{t}} - H_{\text{s}} - H_{\text{fb}})</td>
<td>3.18</td>
<td>1.59</td>
<td>-0.5</td>
<td>-0.92</td>
<td>2.56</td>
<td>1.79</td>
<td>1.33</td>
<td>1.18</td>
</tr>
<tr>
<td>(C)</td>
<td>[2.45 (5.04)]</td>
<td>[2.45 (5.04)]</td>
<td>N/A</td>
<td>7.96</td>
<td>6.31</td>
<td>5.82</td>
<td>8.12</td>
<td>5.68</td>
</tr>
<tr>
<td>(F_{\text{tot}})</td>
<td>[5.63 (8.22)]</td>
<td>[4.04 (6.63)]</td>
<td>N/A</td>
<td>7.04</td>
<td>8.87</td>
<td>7.60</td>
<td>9.45</td>
<td>6.86</td>
</tr>
</tbody>
</table>

\(^a\)Values are given as annual average fluxes (W m\(^{-2}\)). They were computed from the monthly mean components, with SHEBA values for the month of October interpolated from the September and November values. The third column is identical to the second, except that the \(Q_{\text{sl}}\) for July was determined assuming an albedo of 0.64 [Perovich et al., 2002]. \(C\) and \(F_{\text{tot}}\) in columns 2 and 3 use \(k_{\text{s}} = 0.14\) W m\(^{-1}\) K\(^{-1}\) (0.3 W m\(^{-1}\) K\(^{-1}\)). The square brackets show annually averaged hourly atmospheric flux and \(F_{\text{tot}}\) values (requiring all components to be measured that hour), rather than the summation of monthly mean budget components. Two ice categories from M82 are shown. \(C\) and \(F_{\text{tot}}\) are not available for B66.
Figure 22. Comparisons of selected SHEBA monthly mean surface energy budget components with previous studies. Shown are (a) net radiation ($Q^*$, heavy lines) and albedo (thin lines), (b) incoming shortwave radiation ($Q_S$), (c) incoming longwave radiation ($Q_L$), (d) sensible heat flux ($H_s$), (e) latent heat flux ($H_l$), (f) conductive flux ($C$), and (g) the residual or net surface flux ($F_{res}$). The previous studies used in the comparisons are Badgley [1966] (B66), Maykut [1982] (M82) and Lindsay [1998] (L98). In (f) and (g), SHEBA curves are shown using $C$ determined from $k_c = 0.14$ W m$^{-2}$ K$^{-1}$ (squares) and $k_c = 0.3$ W m$^{-2}$ K$^{-1}$ (dots). The error bars show ± one standard deviation of the monthly means from L98. The 3-m ice category from M82 is used in (a), (d), (e), (f), and (g).
Figure 22. (continued)
Francis et al., 2005

Figure 3. Percentages of variance (y-axis) in anomalies of sea ice maximum retreat explained by anomalies in zonal wind (U, black), meridional wind (V, blue), downwelling longwave flux (DLF, green), and the convergence of advected sensible heat (ADV, red) in each peripheral sea of the Arctic Ocean. The bars represent explained variance at lags of 0, 10, 25, 50, and 80 days, where the ice edge anomaly lags the forcing anomaly. Black dots mark locations in Table 1.
Sea-ice change around Alaska: Observations, predictions, impacts

Shimada et al. (2006):
- Ice-ocean interaction amplifies ice retreat in Chukchi & Beaufort Seas
Solar heating of surface waters in pack ice

Perovich et al., GRL, 2007

Mean ann. heat input, MJ m–2.

Solar heat input at 75˚N 165˚W

Oceans north of Alaska have received at least twice as much heat from the sun in recent years compared to the 1980s.
Lateral vs. bottom melt

Figure 7. Time series of solar partitioning during SHEBA: (a) reflected to the atmosphere, (b) absorbed in the snow and ice, and (c) transmitted to the ocean. Contributions from leads, ponds, and bare and snow-covered ice are highlighted.

Perovich et al., 2005
ALBEDO FEEDBACK: A SPACE / TIME LAG

1. GENERAL SETTING
   (BERING/CHUKCHI)
   PACIFIC
   (WINTER)

2. SUMMER
   (SUMMER)
   ICE
   PML
   PSW
   PWW
   LHW
   AW
   CANADA BASIN

3. FALL
   LATE FREEZE-UP
   advected with PSW

4. WINTER
   INCREASED OCEAN HEAT TRANSFER
   EARLY RETREAT

5. NEXT SPRING
   EARLIER RETREAT

Temperature 0°C on Salinity 31.3 psu

Shimada, Carmack et al.
Large-scale heat budget of polar regions

- Serreze et al. (2007): 1979-2001, ERA-40 reanalysis (w/ limited input of satellite data) [shown in red]
- 10 W m\(^{-2}\) over 1 yr surface flux melts c. 1 m ice at \(T_m\)
- Ice export 3 W m\(^{-2}\)
• Meridional profiles of albedo, absorbed solar energy, emitted longwave energy and the net radiation balance at the top of the atmosphere from Peixoto & Oort (1992)
Interannual variability in poleward atmospheric heat flux

- Overland & Turet (1994): Variability in atmospheric heat flux into the North Polar Cap (>70°N) for winter (top) and summer (bottom), with upper solid line indicating total flux
- How is interannual variability in poleward heat flux impacting Arctic ice cover and vice-versa?
Large-scale heat budget of polar regions

Table 2.4.2.1: Longwave heat flux perturbation at which perennial sea ice disappears or snow does not melt away in summer anymore (Curry et al., 1995)

<table>
<thead>
<tr>
<th>Case</th>
<th>Yearround snow (W m⁻²)</th>
<th>Ice melts in summer (W m⁻²)</th>
</tr>
</thead>
<tbody>
<tr>
<td>EC</td>
<td></td>
<td></td>
</tr>
<tr>
<td>n = 1</td>
<td>-17</td>
<td>3</td>
</tr>
<tr>
<td>n = 10</td>
<td>-17</td>
<td>8</td>
</tr>
<tr>
<td>MU</td>
<td>-17</td>
<td>4</td>
</tr>
<tr>
<td>Semtner</td>
<td></td>
<td></td>
</tr>
<tr>
<td>level-0</td>
<td>-16</td>
<td>11</td>
</tr>
<tr>
<td>level-3</td>
<td>-23</td>
<td>5</td>
</tr>
<tr>
<td>Thorndike</td>
<td>-20</td>
<td>20</td>
</tr>
</tbody>
</table>

Fig. 2.4.4.5: Thickness of Arctic sea ice as a function of the poleward atmospheric heat transport as derived from a simple model by Thorndike (1992).
Ice-albedo feedback

- Quantitative theory of ice-albedo feedback derived from engineering (for details see Peixoto & Oort, 1992)
- Applied in a number of theoretical and modeling studies
Ice-albedo feedback: Simple or complicated?

- Atmospheric warming, LW balance
- Solar short-wave flux

Sea Ice

Ocean
Ice-albedo feedback

Fig. 2.4.1.2: Zonal averages of surface and planetary albedo for a standard and a simulated perturbed case (sea-ice albedo equal to that of open water, for details see text) and the resulting changes in solar flux absorbed by earth/atmosphere system (Covey et al., 1991).
Ice-albedo feedback in GCMs

Hall, 2004

Fig. 2. (left column) The geographical distribution of the quasi-equilibrium SAT (°C) increase occurring as a result of CO₂ doubling in the VA experiment. As noted in section 3, the climate variables averaged over the fifth century of the CO₂-doubling experiments were compared to the climate variables averaged over the fifth century of the internal variability experiments to assess the quasi-equilibrium response to doubled CO₂. Results from all four seasons are shown. (right column) As in the left column, except for the FA model.
Ice-albedo feedback in GCMs

Fig. 4. Seasonal breakdown in the VA model of the quasi-equilibrium reduction in solar radiation (W m\(^{-2}\)) at the surface due to CO\(_2\) doubling averaged over the polar caps bounded at 30°N and 30°S. (left) The NH polar cap, (right) its SH counterpart. The black bars denote the contribution to the total reduction from snow on land, while the white bars denote the contribution from sea ice. The reduction in reflected solar radiation at the surface is estimated by multiplying the surface albedo change due to CO\(_2\) doubling by the climatological downward shortwave radiation in the unperturbed VA model at each grid point. This plot therefore represents the change in net incoming radiation at the surface assuming there is no change in the overlying atmosphere, including its cloud distribution.

Hall, 2004
Ice-albedo feedback in GCMs

Fig. 9. Annual-mean time series from the scenario run of SAT and surface albedo averaged over the NH and SH polar caps bounded by 30° lat.

Hall, 2004
Ice-albedo feedback: The role of the ice cover

Fig. 2.4.2.1: Annual cycle of surface temperature, surface albedo and ice thickness for a 1-class and a 10-class model with a $\pm 5$ W m$^{-2}$ perturbation in the longwave flux (Curry et al., 1995).