Ice-albedo feedback - Apr 5
Known about since at least the time of James Croll who argued that ice-albedo feedback amplified solar variations from orbital changes. Later the theory was advanced mathematically with an emphasis on summer season and came to be known as “Milankovitch forcing”
Surface Warming with ice-albedo Feedback and without it
“Polar Amplification”

Zonal Mean Surface Air Temperature
Transient Change in IPCC AR4 runs

A1B Scenario
2080-2099 minus
1980-1999

Thanks to
Marika Holland
"Polar Amplification"

Zonal SAT change normalized by the global mean

**AR4 Models**

**TAR Models**

From Holland and Bitz 2003
“Polar Amplification” only in the Northern Hemisphere
A number of papers find no significant “polar amplification” during the 20th century in Arctic observations (Przybylak 2000; Polyakov et al 2002)

Why do we see “polar amplification” in future scenarios?

Expect higher equilibrium climate sensitivity in polar regions, but the climate is not in equilibrium

We can only measure “transient polar amplification”
= SAT trend in the Arctic / SAT trend for the NH
Observed SAT Anomalies from **Land Stations** only

Shading = sampling error of pentads

Anomaly Reference Period
Transient simulations with climate models also do not have significant polar amplification in individual ensemble members of the 20th century. There is no real difference between models and observations on this matter.

Wang et al 2007 concluded 5 of 20 models were able to reproduce variability like that which is observed in the early 20th century, though not its timing. Thus they argue the early warming is natural variability. However, they were very generous in what they considered a good match between model and observations. I chose a more strict requirement and found no satisfactory comparison ever in CCSM3 in 1500 years of a modern day “control” (annually periodic climate forcing)

One of the first one-dimensional (latitude) energy balance climate models

Key results:

1) Volcanic dust could account for ice ages, but orbital variations could not because the change to solar radiation is too small (oops?)

2) There is a threshold latitude of ice cover on the planet, if ice advances to this latitude, it will then quickly reach the equator (i.e., snowball earth). He estimated this latitude is about 50 deg.
When Solar constant was decreased to 1.6% below modern, The planet plunged into total ice cover!

AND the ice cover does not go away unless the solar constant is massively increased!

Hence Budyko realized ice-albedo feedback could cause irreversible climate change
He also experimented with removing ice (via darkening the surface) and noted that it warmed the poles by 10deg C but did little to the lower latitudes.

A message that has been slow to take hold...
Sensitivity of a Global Climate Model to an Increase of CO₂ Concentration in the Atmosphere

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This study investigates the response of a global model of the climate to the quadrupling of the CO₂ concentration in the atmosphere. The model consists of (1) a general circulation model of the atmosphere, (2) a heat and water balance model of the continents, and (3) a simple mixed layer model of the oceans. It has a global computational domain and realistic geography. For the computation of radiative transfer, the seasonal variation of insolation is imposed at the top of the model atmosphere, and the fixed distribution of cloud cover is prescribed as a function of latitude and of height. It is found that with some exceptions, the model succeeds in reproducing the large-scale characteristics of seasonal and geographical variation of the observed atmospheric temperature. The climatic effect of a CO₂ increase is determined by comparing statistical equilibrium states of the model atmosphere with a normal concentration and with a 4 times the normal concentration of CO₂ in the air. It is found that the warming of the model
1) Strong polar amplification in both hemispheres for a slab-ocean model

2) Polar warming is surface trapped
3) Increase in absorbed shortwave radiation at the Top of Atmosphere is comparable to the radiative forcing of the CO2 increase (which is about 8W/m2)
4) When broken down by month, the increase in absorbed shortwave radiation at the Top of Atmosphere over land occurs primarily in SPRING but over ocean it is from SPRING THROUGH FALL.
5) When broken down by month, the surface warming is maximum in FALL – which was explained by Manabe and Stouffer as the time when anomalously thin sea ice has greatest influence on warming.
Why isn’t the warming in summer? Because ice and snow can’t warm above 0 degC, and ocean warms slowly.
F_c = conductive flux
F_{net} = atmosphere net flux

For simplicity, assume a linear temperature profile and let ocean temperature be 0 deg C:

h = thickness
k = conductivity
T_s = Top surface temperature in deg C

F_c \sim - k \ T_s / h

F_{net} \sim A + BT_s \ (\text{very crudely, assuming no sunlight in fall, A & B are constants})

T_s \sim -hA / (k+hB)
Hence small changes in thin ice have a MUCH bigger influence on surface temperature (in cold season).
Difference is about the same all year, but because ice is thinner in Fall, it will have a bigger affect on warming in Fall. Snow is another factor that adds to insulation and diminishes the effect of thinning as it piles up.

Hence ice-albedo feedback is tricky. It only “operates” when there is sunlight, but it affects surface temperature primarily when the sun is down...
As Norbert Untersteiner is fond of saying... (paraphrasing)

All the positive feedback occurs in summer and all the negative feedback occurs in winter!

The negative feedbacks are from warming causing increased outgoing longwave radiation and the fact that ice re-grows and snow re-accumulates in winter.
Amplified Arctic climate change: What does surface albedo feedback have to do with it?

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Found that the surface albedo feedback, while important, is a lesser factor, than local “net TOA flux forcing” (or surface flux plus lateral heat transport change) and longwave feedback.

What is this net TOA flux forcing (TOA=top of atmosphere)? Good question! It exemplifies our difficulty with articulating feedbacks over limited areas.
The surface albedo feedback is estimated as:

\[ f_{SAF} = \frac{\Delta S^{\text{cyl}}}{\Delta T} \]  

where \( \Delta S^{\text{cyl}} \) represents the change in the top-of-atmosphere shortwave due to replacing the control run surface albedo, \( \alpha_0 \), with the perturbation run surface albedo, \( \alpha' \). The method used for this replacement actually estimates the surface change but has been shown for the GFDL model to be close to the top-of-atmosphere value [Winton, 2005a, 2005b].

\[ f_{\text{non-SAF-SW}} = \frac{\Delta S}{\Delta T} - f_{SAF} \]

\[ f_{\text{LW}} = -\frac{F_{CO2} + \Delta OLR}{\Delta T} \]

and respectively.
Surface Warming with ice-albedo Feedback

and without it

When we “knock out” a feedback have we really quantified its affect, or are feedbacks nonlinear?
Temperature feedback is approximately what Winton refers to as longwave feedback (although clouds and water vapor contribute a little in the polar regions too)

Mark Zelinka, PhD Dissertation kernel feedback estimates transient warming climate runs
\[
T_1 = \frac{1}{\pi a^2 \rho c_w H} (S_1 - A - BT_1 - F)
\]
\[
T_2 = \frac{1}{\pi a^2 \rho c_w H} (S_1 - A - BT_2 + F)
\]
Values of $S_1, S_2,$ and $A$ are tuned to give the equilibrium $(\bar{T}_1, \bar{T}_2) = (298, 278) K$, and $F$ is parameterized as

$$F = \bar{F} + (T'_1 - T'_{i}) \gamma_1 + T'_1 \gamma_2,$$

where $T'_{i} = T_i - \bar{T}_i:

$$\dot{T}_1' = \frac{1}{\pi a^2 \rho c_w H} (-B T'_1 - (T'_1 - T'_2) \gamma_1 - T'_1 \gamma_2)\quad B$$

$$\dot{T}_2' = \frac{1}{\pi a^2 \rho c_w H} (B T'_2 + (T'_1 - T'_2) \gamma_1 + T'_1 \gamma_2)$$

or

$$\begin{pmatrix} \dot{T}_1' \\ \dot{T}_2' \end{pmatrix} = \frac{1}{C} \begin{pmatrix} -B - \gamma_1 - \gamma_2 \\ \gamma_1 + \gamma_2 \end{pmatrix} \begin{pmatrix} T'_1 \\ T'_2 \end{pmatrix}$$
Fast mode

$$\lambda_f = -\frac{B + 2\gamma_1 + \gamma_2}{C}, \quad \tau_f = \frac{C}{B + 2\gamma_1 + \gamma_2}$$

$$\mathbf{v_f} = \begin{pmatrix} 1 \\ -1 \end{pmatrix}$$

Slow mode

$$\lambda_s = -\frac{B}{C}, \quad \tau_s = \frac{C}{B}$$

$$\mathbf{v_s} = \begin{pmatrix} 1 \\ 1 + \frac{\gamma_2}{\gamma_1} \end{pmatrix}$$

Could do something bizarre and make gamma’s negative so timescales are swapped...
\[
\left( \frac{\dot{T}_1}{T_1}, \frac{\dot{T}_2}{T_2} \right) = \frac{1}{C} \left( \frac{\gamma_1}{\gamma_1 + \gamma_2} \frac{\gamma_1}{-B - \gamma_1} \right) \left( \frac{\delta b_1}{\delta b_2} \right) + \left( \delta b_1 \right) = 0
\]

For uniform forcing, \( \delta b_1 = \delta b_2 \):

\[
\frac{\delta T_2}{\delta T_1} = \frac{1}{\frac{\gamma_1}{\gamma_1 + \gamma_2} \frac{\gamma_1}{\gamma_1 B + 2 \gamma_1^2}} = \frac{\gamma_2 B}{1 + \frac{\gamma_2}{\gamma_1}} \quad \text{for} \quad \tau_s \gg \tau_f
\]