Cryosphere radiative forcing and climate feedback

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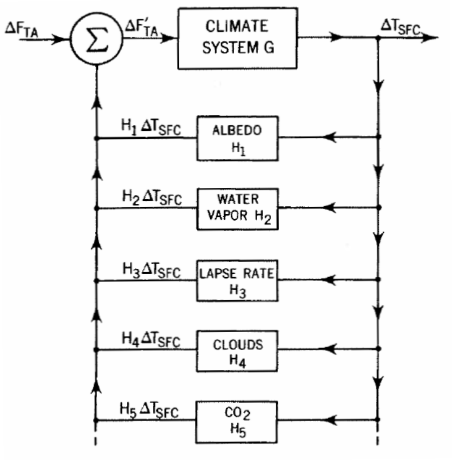
Importance of surface albedo for climate

- A 1% change in Earth's albedo (reflectance) exerts an energy "forcing" almost as large as a doubling of CO_2. Governed by:
  - Clouds and aerosols
  - Snow and vegetation
- Albedo changes with climate and thus provides *feedback* on climate
- Other feedback mechanisms?

Climate feedback theory



Parallel representation of climate feedback assumes individual feedback mechanisms:

1. Are independent
2. Can be added linearly

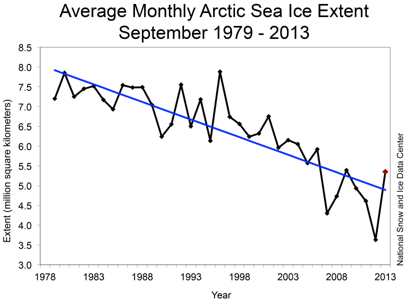
Then, *climate sensitivity*, or change in surface temperature (dT_s) in response to a radiative forcing (dQ) can be represented as:

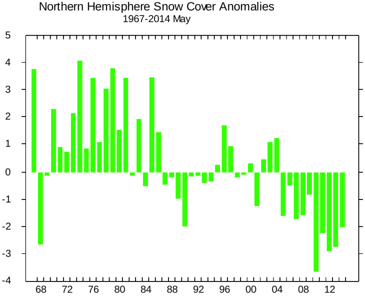
$$\frac{dT_s}{dQ} = -\left(\sum_{i=1}^{N} \frac{\partial x_i}{\partial T_s} \frac{\partial F_{\mathrm{TOA}}}{\partial x_i}\right)^{-1} \quad (1)$$

where each x_i is an environmental state, and F_{TOA} is net top-of-atmosphere energy flux

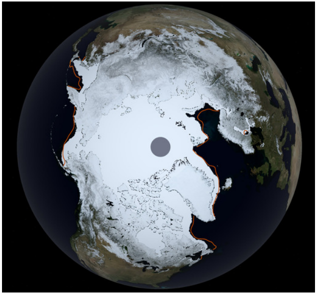
Declining sea-ice and seasonal snow cover

Recent reductions in cryospheric cover are evident





Cryospheric influence on planetary reflectance



- Large capacity for albedo change from Earth's cryosphere
- Questions:
  1. What is the influence of the cryosphere on Earth's solar energy budget?
  2. What has been the radiative impact and associated albedo feedback of recent changes in seasonal snow cover and sea-ice?
- 30+ years of remote sensing observations from which to diagnose the cryospheric contribution to Earth's climate sensitivity

Definition: Cryosphere radiative effect

- *Cryosphere radiative effect* (CrRE): the instantaneous perturbation to Earth's TOA shortwave (solar) energy budget induced by the presence of surface cryospheric components. (Analogous to *cloud radiative effect*).
- Time- (t) dependent CrRE within a region R of area A:

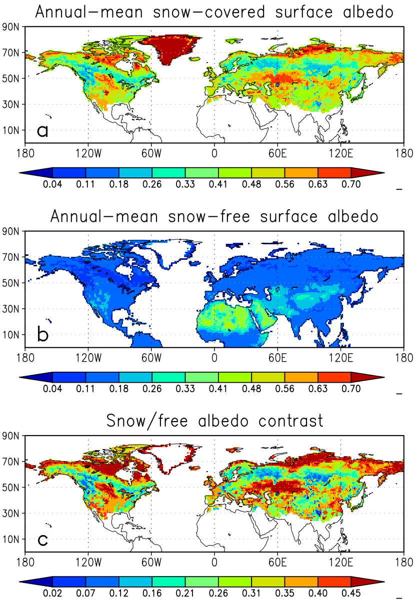
$$\mathrm{CrRE}(t,R) = \frac{1}{A(R)} \int_R \underbrace{S_x(t,r)}_{1} \underbrace{\frac{\partial \alpha}{\partial S_x}(t,r)}_{2} \underbrace{\frac{\partial F}{\partial \alpha}(t,r)}_{3} \, dA(r) \qquad [\mathrm{W\,m^{-2}}] \tag{2}$$

  - S_x: snow or sea-ice cover fraction
  - α: surface albedo
  - F: TOA net solar flux
- Contributions separated from:
  - Land snow and ice
  - Sea-ice

Methods and Data

- S_{snow}: NOAA/Rutgers binary snow cover product (1979–2008), derived from AVHRR data (*Robinson and Frei*, 2000)
- S_{ice}: Sea-ice concentrations (1979–2008) derived from passive microwave sensing (*Cavalieri et al.*, 2008, NSIDC)
- $\Delta\alpha_{\text{snow}}$: Snow-covered albedo: 2000–2008 MODIS surface albedo, filtered with NOAA/Rutgers binary snow cover
- Minimum and maximum $\Delta\alpha_{\text{snow}}$ products created using albedo variance by land-class
- Sea-ice and associated $\Delta\alpha_{\text{ice}}$ partitioned into first-year and multi-year components (*Fowler et al.*, 2004; *Perovich et al.*, 2002; *Tschudi et al.*, 2010)
- $\partial F/\partial\alpha$: *Radiative kernels* derived from NCAR CAM3 and GFDL AM2 models (*Shell et al.*, 2008; *Soden et al.*, 2008), and radiative transfer modeling using global cloud products (ISCCP, APP-x)
- WCRP CMIP3 model archive

Snow-covered / snow-free albedo contrast ($\Delta\alpha_{snow}$)

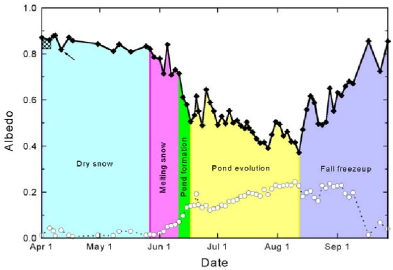


- Reduced snow albedo impact over mature forests
- Large $\Delta\alpha_{snow}$ over grasslands and tundra
- NOAA/Rutgers "snow-covered" surfaces can be up to 50% snow-free

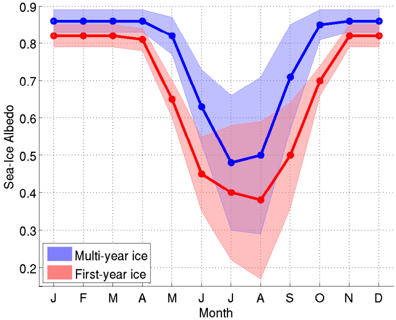
MODIS albedo and variability by land class

UMD Land Class	Snow-covered albedo μ	Snow-covered albedo σ	Snow-free albedo μ	Snow-free albedo σ
Evergreen Needleleaf forest	0.31	0.07	0.10	0.02
Evergreen Broadleaf forest	–	–	0.14	0.01
Deciduous Needleleaf forest	0.36	0.06	0.12	0.01
Deciduous Broadleaf forest	0.35	0.08	0.14	0.02
Mixed forest	0.34	0.09	0.12	0.02
Closed shrublands	0.59	0.06	0.15	0.02
Open shrublands	0.60	0.12	0.19	0.05
Woody savannas	0.42	0.08	0.14	0.02
Savannas	0.49	0.09	0.16	0.02
Grasslands	0.55	0.13	0.19	0.04
Croplands	0.55	0.10	0.16	0.03
Urban and built-up	–	–	0.12	0.02
Barren or sparsely vegetated	0.48	0.14	0.26[c]	0.07
Greenland	0.76	0.07	–	–

Sea-ice albedo



(a) Perovich et al (2002)

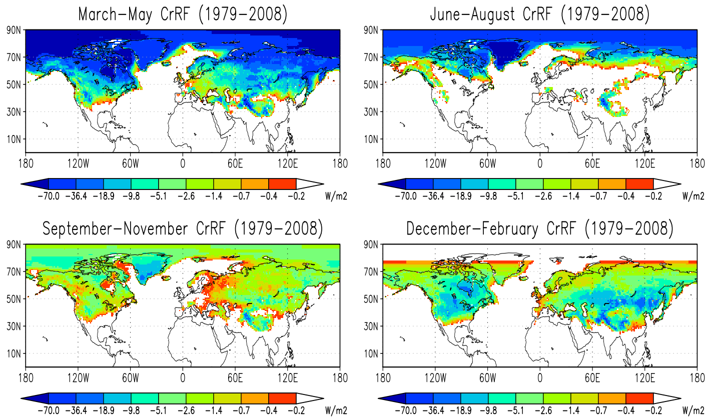


(b) parameterization

- Ranges indicate variability applied in min/max $\Delta\alpha_{\text{ice}}$ scenarios
- Substantial darkening during summer melt
- First-year ice tends to be darker because of greater morphological susceptibility to ponding and tendency to be thinner

1979–2008 mean cryosphere radiative effect

- Annual-mean N. Hemisphere CrRE over land: $-2.0 \pm 0.8\,\mathrm{W\,m^{-2}}$
- Over ocean/sea-ice: $-1.3 \pm 0.4\,\mathrm{W\,m^{-2}}$
- Sea-ice estimate from *Hudson* (2011): $-1.36\,\mathrm{W\,m^{-2}}$



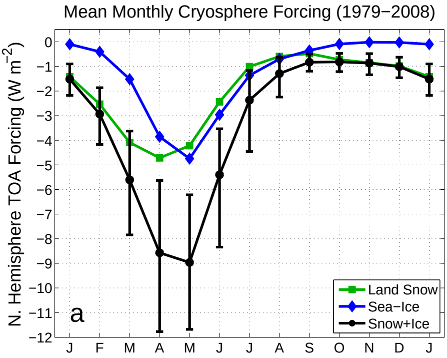
CrRE produced with different methods

Table : N. Hemisphere CrRE [$\mathrm{W\,m^{-2}}$] averaged over 1979–2008

Kernel ($\partial F/\partial\alpha$)	Albedo contrast ($\Delta\alpha$)		
	Low	Central	High
CAM3	–2.3	–3.1	–3.9
AM2	–2.7	–3.6	–4.4
ISCCP	–2.2	–3.1	–4.0
APP-x	–2.6	–3.6	–4.6
CAM3 clear-sky	–4.5	–6.1	–7.7
AM2 clear-sky	–4.3	–5.7	–7.1

- Clouds mask slightly less than half of the cryosphere radiative impact. Consistent with *Qu and Hall* (2005, 2007).

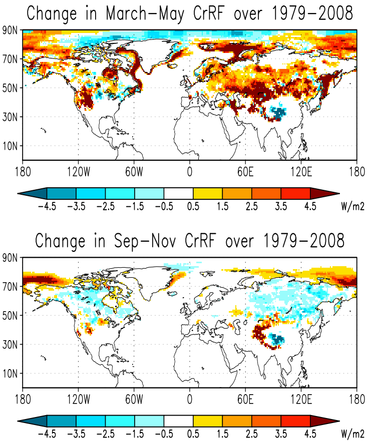
Seasonal cycle of CrRE



- Peak season for land CrRE: March–May
- In May, the Northern Hemisphere reflects an additional $\sim 9\,\mathrm{W\,m^{-2}}$ to space because of the cryosphere
- Larger sea-ice effect in May than June because:
  1. Larger areal coverage
  2. Ice is more reflective (snow cover)

1979–2008 change in cryosphere radiative effect

- 30-year change in land CrRE: $+0.22\ (0.11 - 0.41)\,\mathrm{W\,m^{-2}}$
- 30-year change in sea-ice CrRE: $+0.22\ (0.15 - 0.32)\,\mathrm{W\,m^{-2}}$



1979–2011 CrRE change from Arctic sea-ice

- Recently, *Pistone et al.*, (2014, PNAS) derived a larger estimate ($0.43\,\mathrm{W\,m^{-2}}$) of the NH-averaged 1979–2011 change in CrRE due to Arctic sea-ice loss.

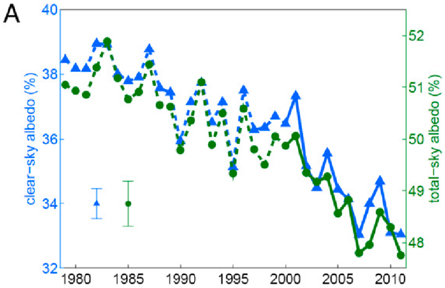


Figure : Pistone et al. (2014)

1979–2008 change in CrRE: Seasonal cycle

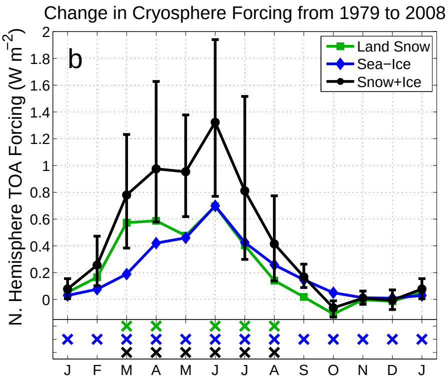


Figure : 'X' indicates month of statistically-significant change ($p = 0.01$)

- Sea-ice peak change occurs in **summer**
- June peak in land snow change is sensitive to mountain snow cover estimates (Himalaya, Tien Shan)

Do cloud changes enhance or reduce the CrRE change?

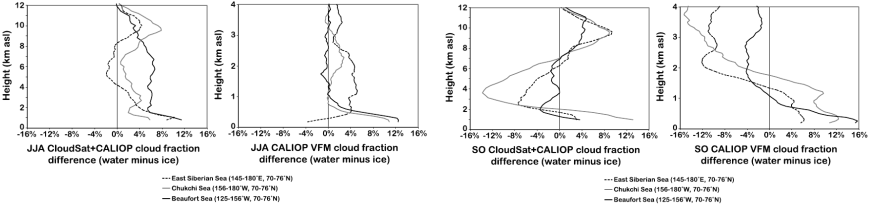
Table : Change in N. Hemisphere CrRE [$\mathrm{W\,m^{-2}}$] from 1979 to 2008.
Annually-repeating kernel. Annually-varying kernel.

Kernel ($\partial F/\partial\alpha$)	Albedo Contrast ($\Delta\alpha$)		
	Low	Central	High
CAM3	0.26	0.38	0.48
AM2	0.29	0.40	0.49
ISCCP	0.40	0.57	0.72
APP-x	0.31	0.48	0.59
CAM3 clear-sky	0.58	0.82	1.00
AM2 clear-sky	0.58	0.77	0.97

- CrRE *changes* are greater with annually-varying cloud conditions (ISCCP and APP-x) than with single-year model kernels

Coupled cryosphere–cloud evolution

Cloud changes over areas of ice loss

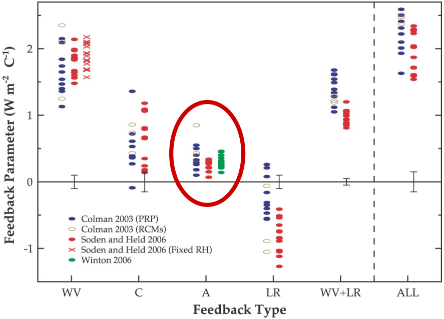


- *Kay and Gettleman* (2009): analysis with CloudSat and CALIOP
- Little change in cloud cover over newly open water during summer months because weak surface–air temperature contrast limits atmosphere–ocean coupling
- In fall, low clouds increase over newly open water because of lower static stability

Climate feedback components

- Combined strength of feedback mechanisms (i) governs climate sensitivity:

$$\lambda \approx -\left(\sum_{i=1}^{N} \frac{\partial F_i}{\partial T_s}\right)^{-1} \qquad [\mathrm{K\,(W\,m^{-2})^{-1}}] \qquad (3)$$



- IPCC AR4 figure 8.14
- WCRP CMIP3 global albedo feedback: $\sim 0.3\,\mathrm{W\,m^{-2}\,K^{-1}}$ (*Winton*, 2006; *Soden et al.*, 2008; *Shell et al.*, 2008)

Cryosphere albedo feedback

What is $\Delta F_{cryo}/\Delta T_s$?

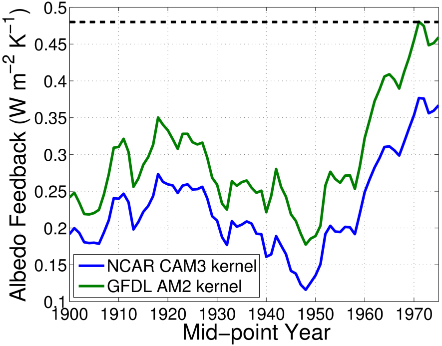
- 1979–2008 boreal feedback (from observations): $\mathbf{0.62}$ $(0.3 - 1.1)\,\mathrm{W\,m^{-2}\,K^{-1}}$
- CMIP3 1980–2010 N. Hemisphere albedo feedback (18 models): $0.25 \pm 0.17\,\mathrm{W\,m^{-2}\,K^{-1}}$
- Global cryosphere feedback is less because SH sea-ice extent has increased (e.g., *Cavalieri and Parkinson*, 2008). If S. Hemisphere cryosphere did not change, global feedback was: $0.48\,\mathrm{W\,m^{-2}\,K^{-1}}$
- Why has boreal albedo feedback been so strong?

Possible explanations of model–observation discrepency

1. Internal variability. Models can simulate a 30-year boreal albedo feedback as strong as observed during 1979–2008, but they simply failed to do so over this particular timeframe. Could indicate that actual feedback was anomalously strong during the past 3+ decades.
2. Models have a strong contribution to albedo change from non-cryospheric sources, thus biasing the comparison with cryospheric observations
3. Models are generally incapable of simulating as strong of boreal albedo feedback as observed during 1979–2008 over any 30-year window representative of current climate, indicating a systematic problem in the representation of processes responsible for rapid albedo change
4. The observations are flawed and the actual feedback is not as strong

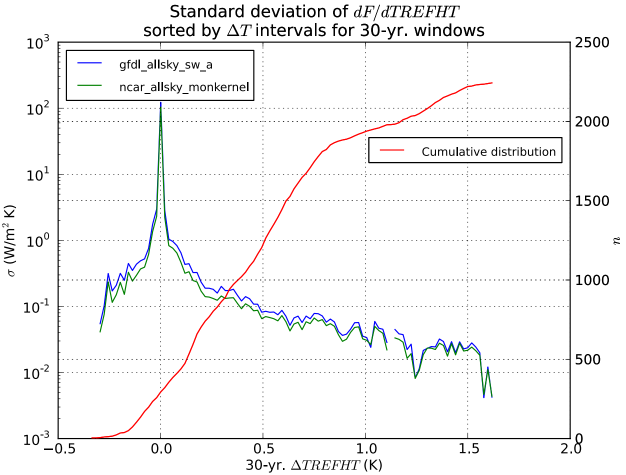
Possibility #1: Temporal variabilty in feedback

- Temporal evolution in feedback strength in a CCSM4 20th century simulation (preliminary work by Adam Schneider):



- Is this robust across models and ensemble members?
- Why might the cryosphere respond more sensitively to a unit warming now than in the past?
- When will feedback peak, and what governs timing of inflection?

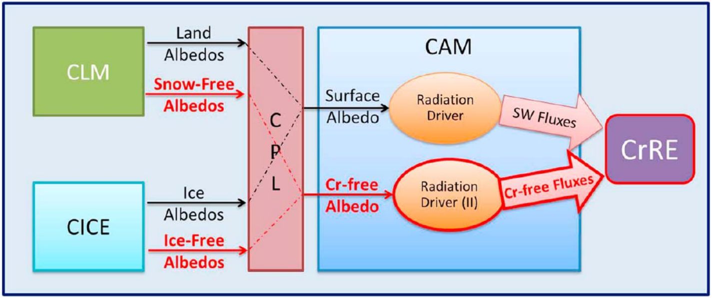
Possibility #1: Temporal variabilty in feedback



- How much ΔT and/or Δt is needed to derive a meaningful feedback estimate?

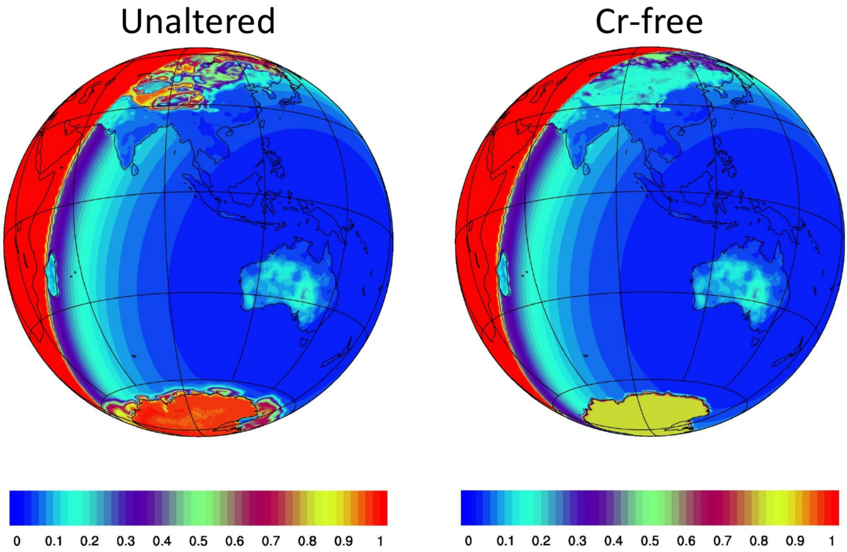
Diagnostic CrRE in CESM

Implemented by Justin Perket (*Perket et al*, 2014, JGR)



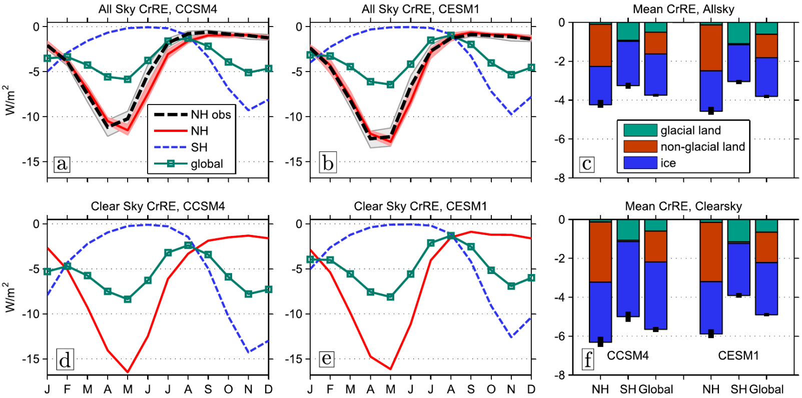
- Atmospheric radiative transfer calculations performed every timestep with and without cryospheric cover
- Calculation does not alter simulated climate state
- Implemented in two versions of CESM

Visible direct albedo (single timestep)



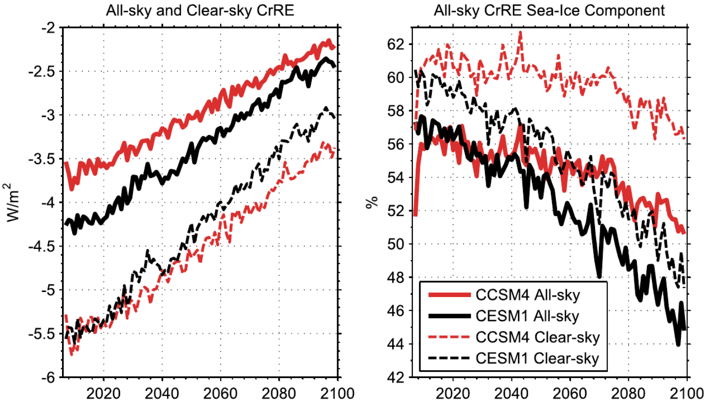
Diagnostic CrRE in CESM

Seasonal cycles of NH/SH CrRE and their contributions:



Diagnostic CrRE in CESM

21st century evolution of CrRE under the RCP8.5 scenario:



- By 2100, all-sky global CrRE changes by $+1.4\,\mathrm{W\,m^{-2}}$ in CCSM4 (38% reduction in magnitude) and by $+1.7\,\mathrm{W\,m^{-2}}$ in CESM1 (46% reduction). Sea-ice reductions are larger in CESM1

Conclusions

- N. Hemisphere cryosphere radiative effect is $-3.3 \pm 1.2\,\mathrm{W\,m^{-2}}$, peaking in May at $\sim -9\,\mathrm{W\,m^{-2}}$
- Boreal cryospheric cooling decreased by at least $0.45\,\mathrm{W\,m^{-2}}$ between 1979 and 2008, with roughly equal contributions from land snow and sea-ice reductions. Change in sea-ice CrRE may have been up to $2\times$ greater than our estimates (*Pistone et al*, 2014).
- Boreal cryosphere albedo feedback is currently *at least* $0.6\ (0.3 - 1.1)\,\mathrm{W\,m^{-2}\,K^{-1}}$, more than double the mean feedback ($0.25\,\mathrm{W\,m^{-2}\,K^{-1}}$) simulated by CMIP3 models over 1980–2010
- Climate simulations indicate that global CrRE will increase (become less negative) by 1.4–1.7 $\mathrm{W\,m^{-2}}$ by century's end under the RCP8.5 scenario