Measuring Paleo-(Climate)-Sensitivity

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Colloquium "Constraining estimates of future climate variability from geological records "

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In collaboration with
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Outline

1. Conceptual Basics
2. Application 1: Late Pleistocene — last 800 kyr
3. Application 2: Late Cenozoic — Last 20 Myr

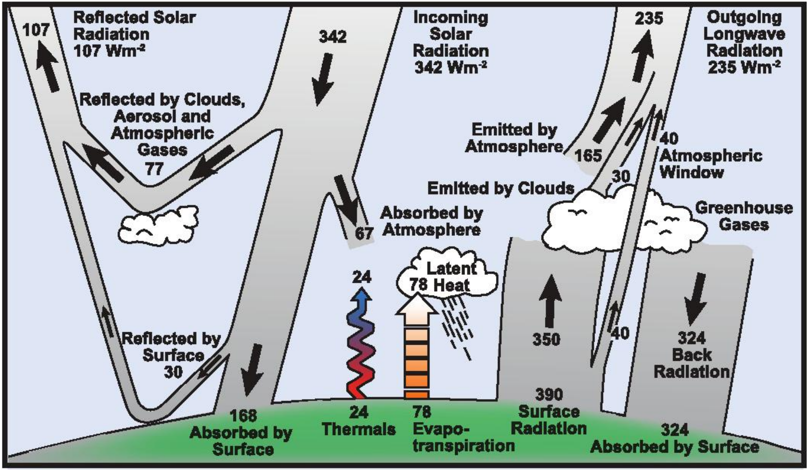
Outline

1 Conceptual Basics

2 Application 1: Late Pleistocene — last 800 kyr

3 Application 2: Late Cenozoic — Last 20 Myr

Present Day Radiative Forcing



(IPCC, 2007, based on Kiehl & Trenberth, 1997)
(revised by Trenberth et al 2009 BAMS (not included here))

Earth's Radiative Budget

Equilibrium temperature change $\Delta T_{\mathrm{E},\infty}$
for a given radiative forcing ΔR (or ΔF or ΔQ)

$$\Delta T_{\mathrm{E},\infty} = \frac{-\Delta R}{\lambda}$$

Climate feedback parameters

$$\lambda = \lambda_{\mathrm{Planck}} + \lambda_{\mathrm{water\ vapour}} + \lambda_{\mathrm{lapse\ rate}} + \lambda_{\mathrm{clouds}} (+\lambda_{\mathrm{albedo}})$$

$$\lambda = \lambda_{\mathrm{Planck}} + \lambda_{\mathrm{else}}$$

Initially: only Planck feedback, others (λ_{else}) ignored

$$\Delta T_{\mathrm{E,P}} = \frac{-\Delta R}{\lambda_{\mathrm{Planck}}} = \frac{-\Delta R}{-3.2\ \mathrm{W\ m^{-2}\ K^{-1}}}$$

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$$\lambda = \lambda_{\mathrm{Planck}} + \lambda_{\mathrm{else}}$$

(specific) climate sensitivity S

$$S = \frac{\Delta T_{\mathrm{E},\infty}}{\Delta R}$$

State-of-the-Art

Climate sensitivity ($\Delta T_{2\times\mathrm{CO_2}}$)

equilibrium temperature change for doubling CO_2 concentration

Problem: When is equilibrium?

Calculated range of $\Delta T_{2\times\mathrm{CO_2}}$ (IPCC 2007 or others): 2.1–4.4 K

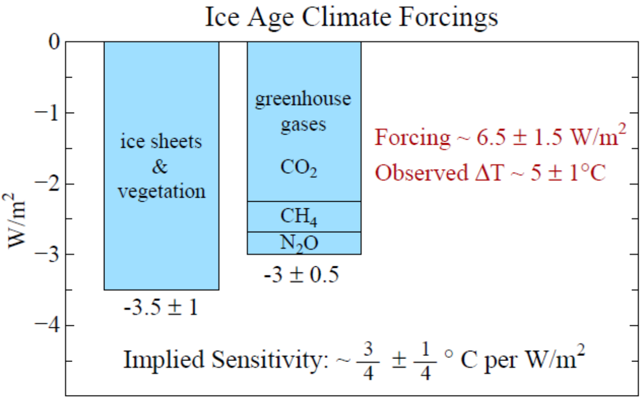
Problem: Factor $\geq$ 2 uncertainty, depending on the climate model

Forcing $\Delta R_{2\times\mathrm{CO_2}} = 5.35\ \mathrm{W\ m^{-2}} \cdot ln(2) = 3.7\ \mathrm{W\ m^{-2}}$.

$$\frac{\Delta T_{2\times\mathrm{CO_2}}}{\Delta R_{2\times\mathrm{CO_2}}} = S_{2\times\mathrm{CO_2}} = 0.6 - 1.2\ \mathrm{K\ (W\ m^{-2})^{-1}}$$

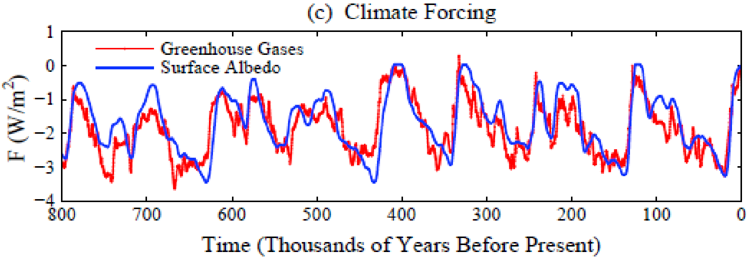
Other Approaches (e.g. Hansen et al, 2007, 2011)

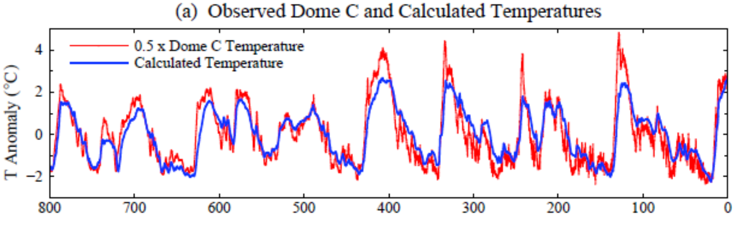
1. Use LGM for ΔT and $\Delta R \Rightarrow S = 0.75 \pm 0.25$ K (W m^{-2})$^{-1}$



Other Approaches (e.g. Hansen et al, 2007, 2011)

1. Use LGM for ΔT and $\Delta R \Rightarrow S = 0.75 \pm 0.25$ K (W m^{-2})$^{-1}$
2. Keep S constant and calculate ΔT out of given ΔR.





Other Approaches (e.g. Hansen et al, 2007, 2011)

1. Use LGM for ΔT and $\Delta R \Rightarrow S = 0.75 \pm 0.25$ K (W m^{-2})$^{-1}$
2. Keep S constant and calculate ΔT out of given ΔR.

Points for improvements:

1. ΔT (observed) is 0.5$\times$ EPICA-Dome-C ΔT which is wrong, because the polar amplification changes with climate / time.
2. There is no global time series of ΔT, urgently wanted.
3. S might depend on climate ($S = f(T)$), thus taking $S_{2\times CO_2}$ for LGM and vice versa might be wrong.
4. More knowledge on changes in albedo available.

Our Approach:

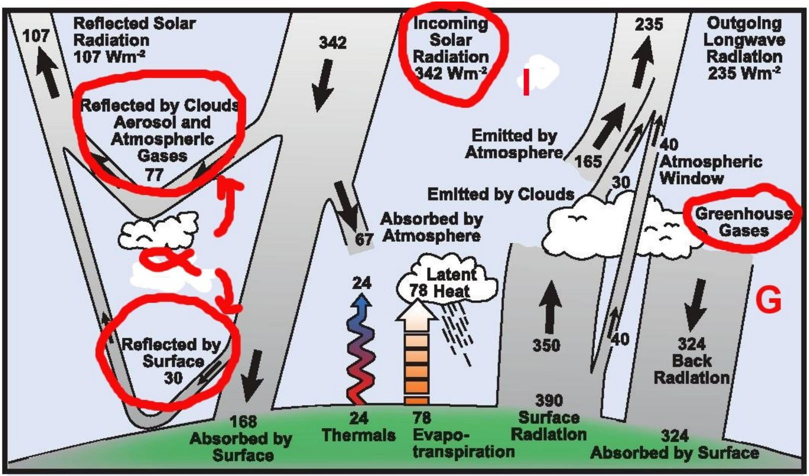
Refine ΔR as far as possible out of data sets.

$$\Delta T_{\mathrm{E},\infty} = \frac{-\Delta R}{\lambda}$$

Two applications:

1. Late Pleistocene:
   (a) Refine ΔR over last 800 kyr
   (b) For LGM: use a ΔT and our $\Delta R \Rightarrow S = \Delta T / \Delta R$.
2. Late Cenozoic:
   (a) Use data-based ΔT and constant S for $\mathrm{CO_2} = f(\Delta R)$
   (b) Use ΔT and ice core CO_2 to calculate variability in S

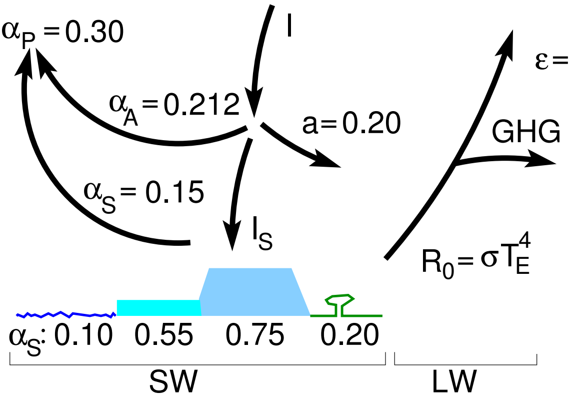
Our Approach



Our Approach: Processes change in the radiative budget based on **data**

Considered paleo changes: *I*: incoming solar radiation; *G*: GHG; α: albedo

Simplified View on the Radiative Budget



Considered paleo changes:

I: incoming solar radiation; *GHG*: greenhouse gases

α_S: surface: land ice, snow, sea ice, vegetation α_A: atmosphere dust

annual mean and zonally averaged view

(Köhler et al., QSR 2010)

Outline

1. Conceptual Basics
2. Application 1: Late Pleistocene — last 800 kyr
3. Application 2: Late Cenozoic — Last 20 Myr

The Processes

Orbital Forcing

GHG Forcing
(CO_2, CH_4, N_2O)

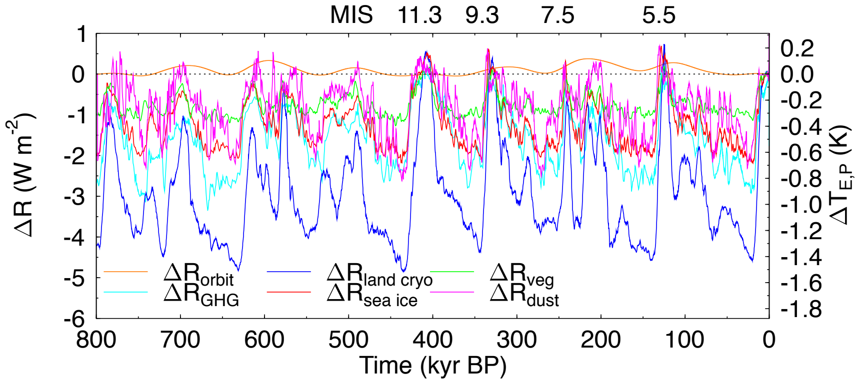
Surface albedo (1): Land Cryosphere
(land ice, sea level, snow cover)

Surface albedo (2): Sea ice

Surface albedo (3): Vegetation

Atmospheric albedo: Aerosols (Dust)

Individual radiative forcings



Considering orbital variation, GHG, surface albedo (land ice sheets, snow, exposed shelves, sea ice) and atmospheric albedo (dust)

(Köhler et al., QSR 2010)

Individual radiative forcings for LGM

Process	Uncertainties	$\Delta R \pm 1\sigma$ (W m^{-2})		upper err (W m^{-2})
Orbit	—	0.01 ± 0.00		
GHG		-2.81 ± 0.25		± 0.37
CO_2	$\sigma_{\mathrm{CO_2}} = 2$ ppmv; $\sigma_R = 10\%$		-2.10 ± 0.22	
CH_4	$\sigma_{\mathrm{CH_4}} = 10$ ppbv; $\sigma_R = 10\%$; $\sigma_{\mathrm{efficacy}} = 5\%$; $\sigma_{\mathrm{interN_2O}} = 0.02$ W m^{-2}		-0.40 ± 0.05	
N_2O	$\sigma_R = 0.1$ W m^{-2}		-0.30 ± 0.10	
land cryosphere		-4.54 ± 0.90		± 1.50
land ice	$\sigma_I = 0.2\%$; $\sigma_{\mathrm{area}} = 10\%$; $\sigma_{\alpha_{LI}} = 0.1$		-3.17 ± 0.63	
sea level	$\sigma_I = 0.2\%$; $\sigma_{\mathrm{area}} = 20\%$; $\sigma_{\alpha_L} = 0.05$		-0.55 ± 0.29	
snow cover	$\sigma_I = 0.2\%$; $\sigma_{\mathrm{area}} = 20\%$; $\sigma_{\alpha_L} = 0.05$		-0.82 ± 0.58	
sea ice		-2.13 ± 0.53		± 0.64
sea ice N	$\sigma_I = 0.2\%$; $\sigma_{\mathrm{area}} = 20\%$; $\sigma_{\alpha_{SI}} = 0.1$		-0.42 ± 0.12	
sea ice S	$\sigma_I = 0.2\%$; $\sigma_{\mathrm{area}} = 20\%$; $\sigma_{\alpha_{SI}} = 0.1$		-1.71 ± 0.51	
vegetation	$\sigma_I = 0.2\%$; $\sigma_{\alpha_L} = 0.05$	-1.09 ± 0.57		
dust	$\sigma_I = 0.2\%$; $\sigma_{\alpha_A} = 50\%$	-1.88 ± 0.94		
subtotal		-12.43 ± 1.39		± 3.19
for 'Charney' sensitivity S_C (no snow cover and sea ice)		-9.48 ± 1.15		± 2.55
Other approaches (e.g. Hansen)		-6.5 ± 1.5		

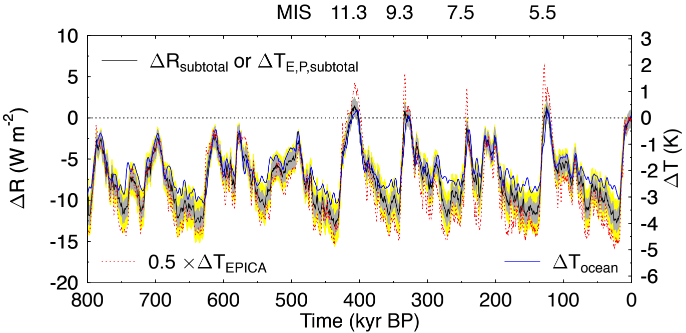
(Köhler et al., QSR 2010)

Total radiative forcing

LGM: -12.4 ± 3.2 W m^{-2}. $\Delta T_{E,P}$: -3.9 ± 0.8 K.

With feedbacks (water vapour, clouds, lapse rate) $\lambda_{else}^{today} = 1.65 \pm 0.49 \text{ W m}^{-2} \text{ K}^{-1}$:

$\Delta T_{E,\infty}^{LGM}$: -8.0 ± 1.6 K.



⇒ Feedbacks strength λ_{else} and then also S depends on climate.

(Köhler et al., QSR 2010)

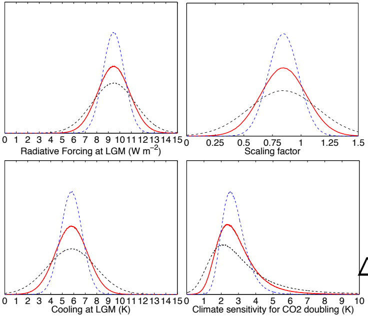
Climate Sensitivity for 2× CO_2 based on LGM data

Asymmetry in *S* for warming and cooling (Hargreaves et al., 2007).

Our $\Delta R_{\text{Charney}} = -9.5 \pm 1.2$ W m^{-2}

factor $f = 0.85 \pm 0.2$

(Hargreaves et al., 2007)



$\Delta T = 5.8 \pm 1.4$K

(Schneider v. Deimling et al., 2006)

$\Delta T_{2\times CO_2} = 1.4 - 5.2$ K

$$\Delta T_{2\times CO_2} = \frac{\Delta T}{\Delta R} \cdot \frac{\Delta R(2\times\mathrm{CO_2})}{\text{scale}} = \frac{5.8\mathrm{K}}{9.46\mathrm{Wm^{-2}}} \cdot \frac{3.71\mathrm{Wm^{-2}}}{0.85} \approx 2.6\mathrm{K}$$

(Köhler et al., QSR 2010)

Conclusions I

1. Improved $\Delta R \Rightarrow \Delta T_{\mathrm{E,P}}^{LGM} = -3.9 \pm 0.8$ K without feedbacks.
2. Feedback strength is climate dependent.
3. $\Delta T_{2\times CO_2}$ based on our LGM ΔR compilation is 2.4 K (1.4– 5.2 K).

Wanted Improvements

1. Global mean surface ΔT time series over 800 kyr wanted.
2. Inconsistent picture of ΔT between deep ocean and ice cores in the warmer-than-Holocene-interglacials.
3. Understand the climate dependency of S.

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3. Application 2: Late Cenozoic — Last 20 Myr

Necessary Ingredients

Two out of three

ΔT

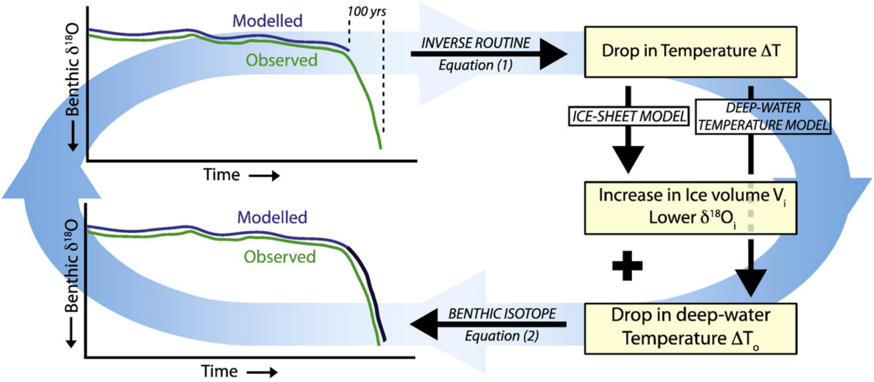
ΔR

S

are necessary to calculate the third after

$$S = \Delta T / \Delta R$$

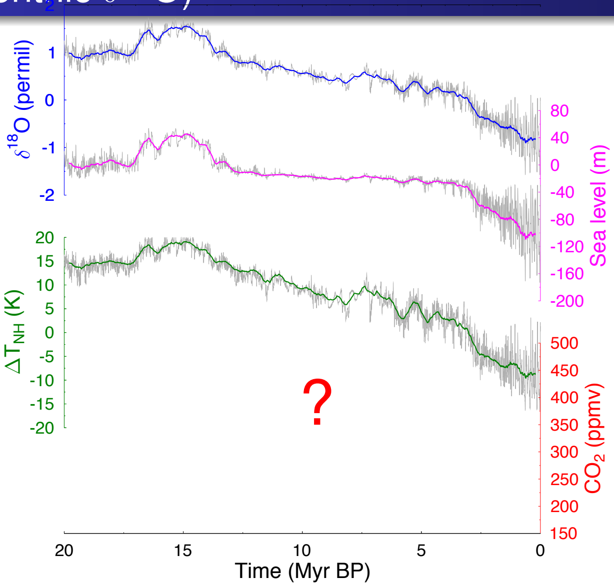
Northern Hemispheric Ice Sheets and benthic δ^{18}O



Deconvolute stacked benthic δ^{18}O into climate variables ($\Delta T_{\mathrm{deep\ o}}$, $\Delta T_{\mathrm{atm\ (40-80°N)}}$), size of ice sheets, sea level, snow cover)

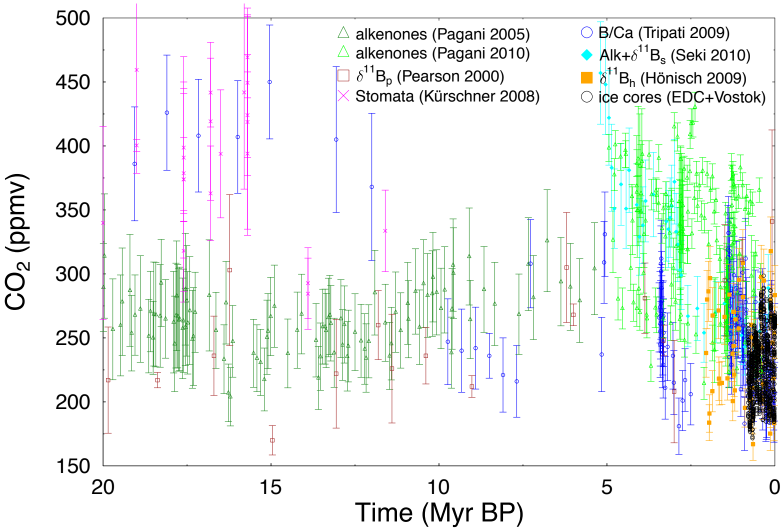
(Bintanja et al., 2005; de Boer et al., 2011)

ΔT: =f(benthic $\delta^{18}O$)



(after Bintanja et al., 2005; van de Wal et al., 2011; de Boer et al., 2011)

CO_2: proxy diversity



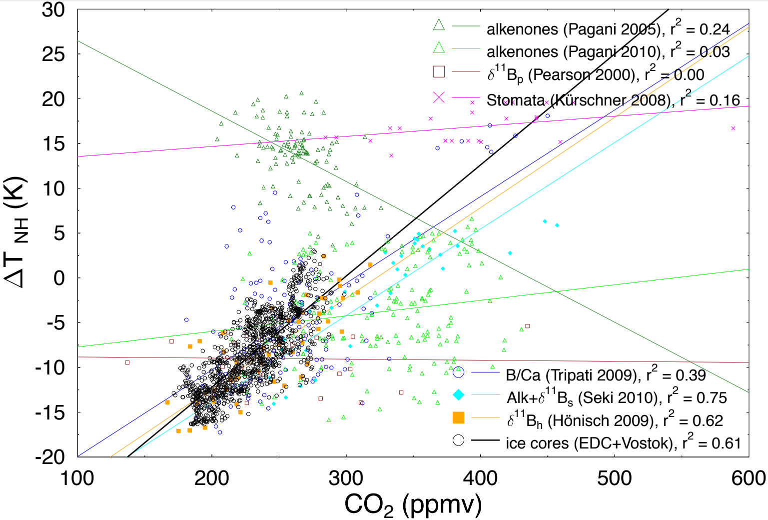
after van de Wal et al., 2011 CPD

Initial Approach

Use data-based $\Delta T = f(\delta^{18}O)$,
assume constant $S = \Delta T / \Delta R$
to calculate $\mathrm{CO_2} = f(\Delta R)$

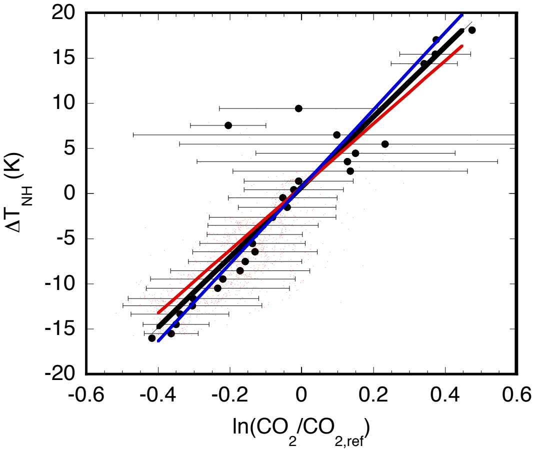
$$\Rightarrow \Delta R = \Delta T / S$$

Relationship ΔT_{NH}—CO_2



(after van de Wal et al., 2011 CPD)

ΔT_{NH}—CO_2 1: Empirical Relationship



resampled and binned data in intervals of $\Delta(\Delta T_{NH}) = 1$ K

$C = 39 \pm 4K$ regression slope from modelled ΔT_{NH} and CO_2 data

(van de Wal et al., 2011, CPD)

ΔT_{NH}—CO_2 2: Theoretical Relationship

$$\Delta T = S \cdot \Delta R$$

$$\Delta T_{NH} = C \cdot \ln \frac{\mathrm{CO_2}}{\mathrm{CO_{2,ref}}} \text{ with } C = \frac{\alpha\beta\gamma S_C}{1-f}$$

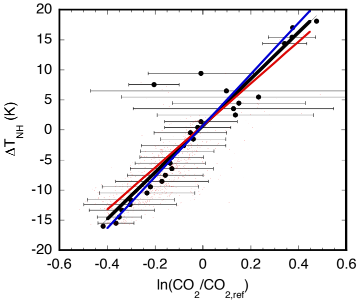
LGM parameters:

$\alpha = \Delta T_{NH} / \Delta T_{global} = 15\ \mathrm{K}/6\ \mathrm{K} = 2.5$
$\beta = 5.35$ W m^{-2}: radiative forcing of CO_2
$\gamma = 1.3$: enhancement factor for non-CO_2 GHG (CH_4, N_2O)
$S_C = 0.72$ K (W m^{-2})$^{-1}$: Charney climate sensitivity (fast feedbacks: Planck, water vapour, lapse rate, clouds, sea ice, albedo)
$f = 0.72$: feedbacks of slow processes (land ice, dust, vegetation)

$C = 43K$ theoretical calculation based LGM data and constant S

For comparision:
pure S_{Charney} ($f = 0; \gamma = 1; \alpha = 1$) $\Rightarrow C_C = 3.9$ K and $\Delta T_{global} = 2.7$ K
(van de Wal et al., 2011, CPD)

Develop relationship atmospheric ΔT_{NH}—CO_2



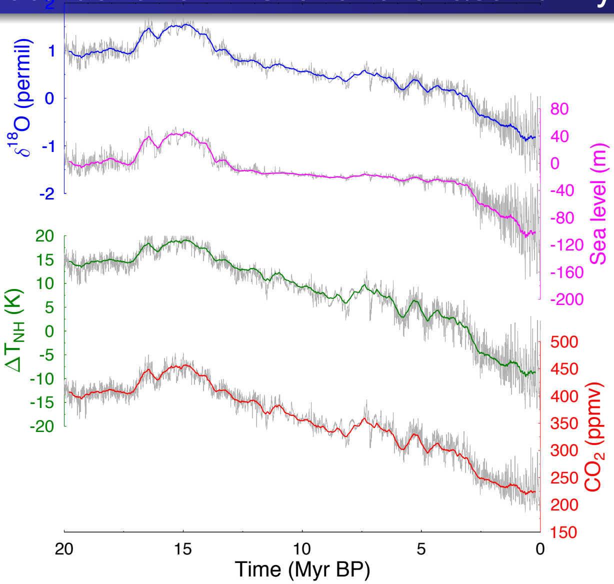
$$\Delta T_{NH} = C \cdot \ln \frac{\mathrm{CO_2}}{\mathrm{CO_{2,ref}}} \text{ with } C = \frac{\alpha\beta\gamma S}{1-f}$$

Two independent approaches to calculate the slope:

1. $C = 39 \pm 4K$ regression slope from modelled ΔT_{NH} and CO_2 data
2. $C = 43K$ theoretical calculation based LGM data and constant S

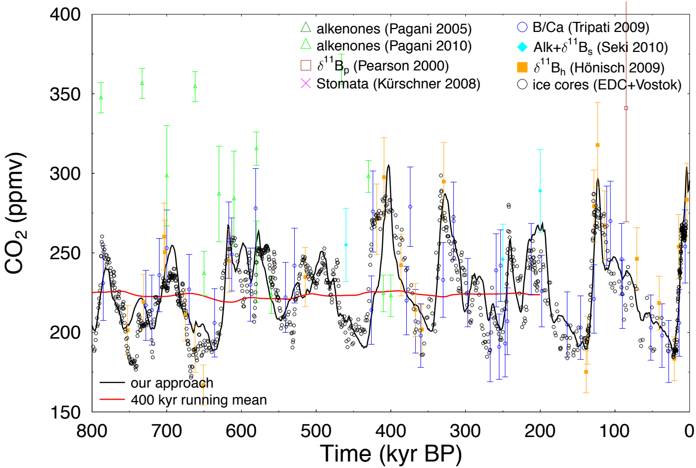
(van de Wal et al., 2011, CPD)

Deconvolute benthic $\delta^{18}O$ over the last 20 Myr



after van de Wal et al., 2011 CPD

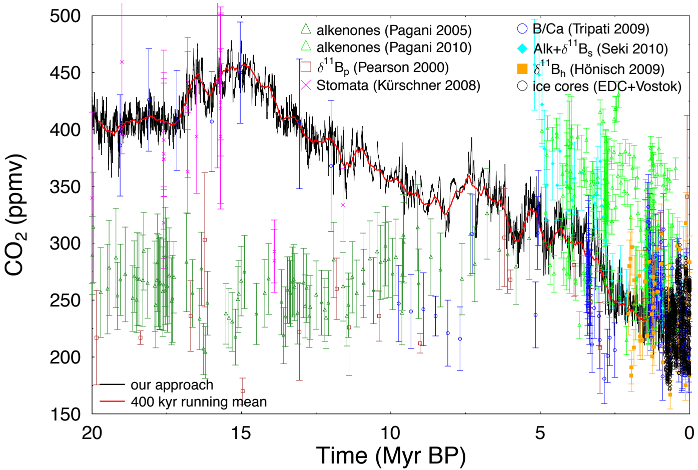
CO_2 reconstructions, the last 20 Myr



Glacial/interglacial amplitudes captured, details wrong

after van de Wal et al., 2011 CPD

CO_2 reconstructions, the last 20 Myr



Assumption: relation CO_2-ΔT unchanged with time!!!

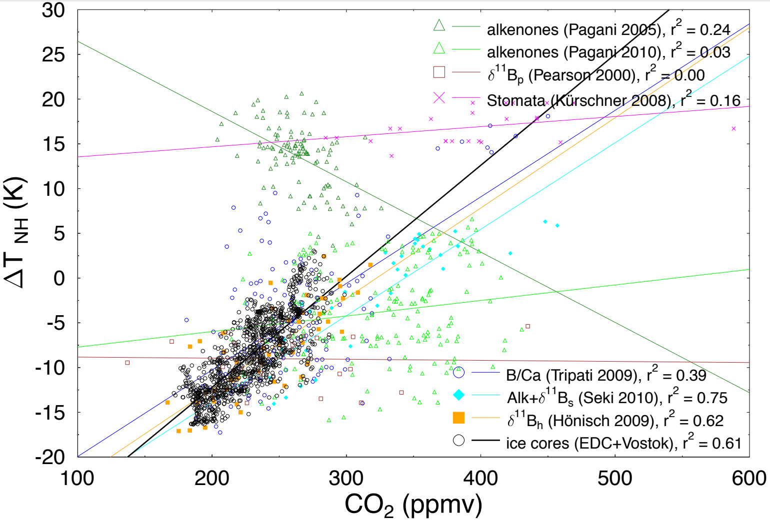
after van de Wal et al., 2011 CPD

Alternative Approach

Use data-based $\Delta T = f(\delta^{18}O)$,
and the best constrained $\Delta R = f(\mathrm{CO_2}$, ice core)
to calculate the variability in $S = f(T)$

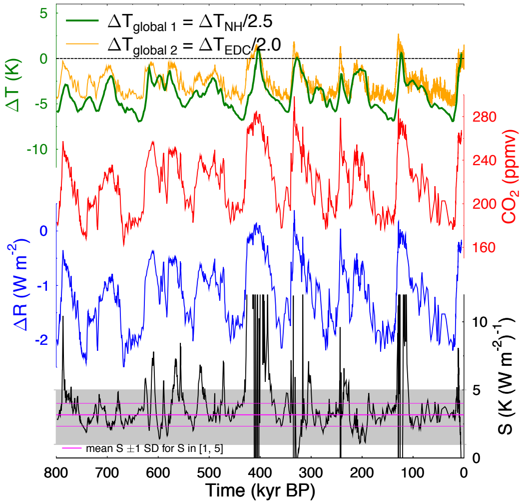
$$\Rightarrow S = \Delta T / \Delta R$$

Alternative: $S = f(T)$ based on ice core data



(after van de Wal et al., 2011 CPD)

Alternative: $S = f(T)$ based on ice core data



$S_{CO_2} = 3.2 \pm 0.8$ K (W m^{-2})$^{-1}$, CV = 26%

interglacials: $\Delta T \sim 0$; $\Delta R \sim 0$; synchronisation!!!

Revise Theoretical Relationship ΔT_{NH}—CO_2

$$\Delta T_{NH} = C \cdot \ln \frac{\mathrm{CO_2}}{\mathrm{CO_{2,ref}}} \text{ with } C = \frac{\alpha\beta\gamma S_C}{1-f}$$

$$\Rightarrow: S^{\mathrm{LGM}}_{\mathrm{CO_2}} = \frac{\Delta T^{\mathrm{LGM}}_{global}}{\Delta R^{\mathrm{LGM}}_{CO_2}} = \frac{\gamma S_C}{1-f} = \frac{1.3 \cdot 0.72}{1-0.72} = 3.3 \text{ K (W m}^{-2})^{-1}$$

$C = 43K$ theoretical calculation based LGM data and constant S

Revise Theoretical Relationship ΔT_{NH}—CO_2

$$\Delta T_{NH} = C \cdot \ln \frac{\mathrm{CO_2}}{\mathrm{CO_{2,ref}}} \text{ with } C = \frac{\alpha\beta\gamma S_C}{1-f}$$

$$\Rightarrow: S_{\mathrm{CO_2}}^{\mathrm{LGM}} = \frac{\Delta T_{global}^{\mathrm{LGM}}}{\Delta R_{CO_2}^{\mathrm{LGM}}} = \frac{\gamma S_C}{1-f} = \frac{1.3 \cdot 0.72}{1-0.72} = 3.3 \text{ K (W m}^{-2})^{-1}$$

$C = 43K$ theoretical calculation based LGM data and constant S

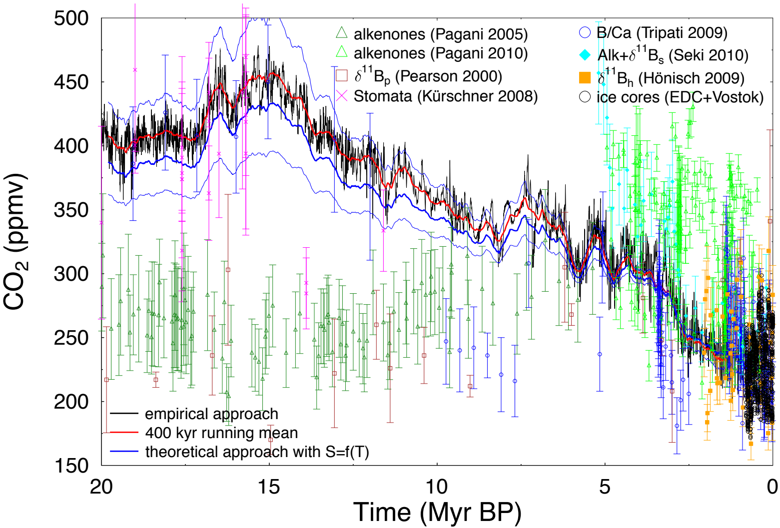
Revision:

$S_{\mathrm{CO_2}}^{\text{ice cores}} = 3.2$ K (W m^{-2})$^{-1} \pm 26\%$

$\Rightarrow$ Ice cores suggest a variability in S of $\pm 26\%$,

thus $C = 43 \pm 11K$

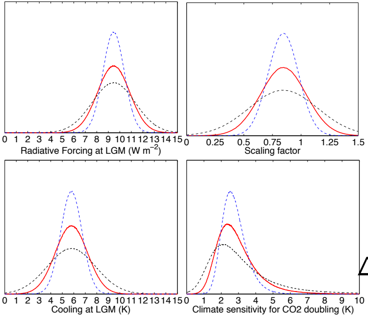
Revise CO_2 reconstructions, the last 20 Myr



after van de Wal et al., 2011 CPD

Open Questions: Asymmetry in S for cooling and warming

Our $\Delta R_{\text{Charney}} = -9.5 \pm 1.2$ W m^{-2}



factor $f = 0.85 \pm 0.2$

(Hargreaves et al., 2007)

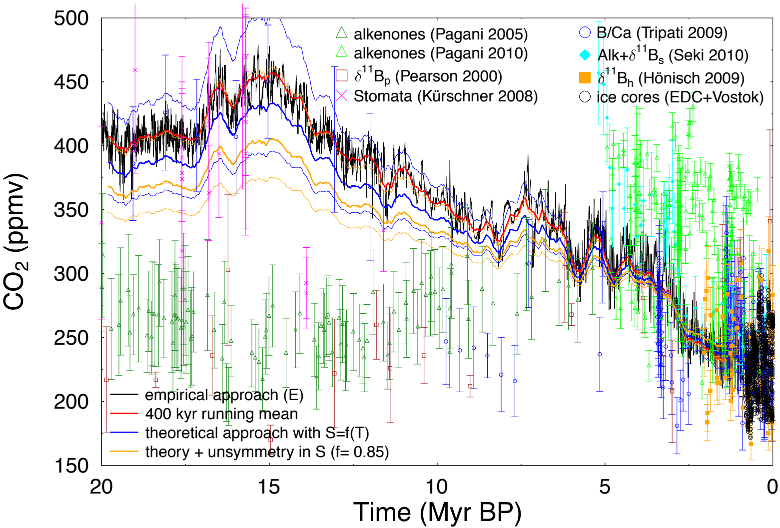
$\Delta T = 5.8 \pm 1.4$K

(Schneider v. Deimling et al., 2006)

$\Delta T_{2\times CO_2} = 1.4 - 5.2$ K

Asymmetry in S (scaling factor f) not considered so far.

Revise CO_2, consider unsymmetry in S (here: f= 0.85)



after van de Wal et al., 2011 CPD

Conclusions II

1. If one assumes constant $S \Rightarrow CO_2$ can be calculated out of ΔT.
2. Alternatively, if we believe in a ΔT we can obtain a climate-dependent S from the ice core CO_2.
3. For which forcing ΔR is S calculated? e.g. S_{Charney}, S_{all}, $S_{\mathrm{CO_2}}$.
4. Approach is weak in MIS 5, 7, 9, 11 with both ΔR and $\Delta T \sim 0$.
5. Asymmetry in S (scaling factor f) not considered so far.
6. We need to agree on global temperature records!.
7. (Precise uncertainty treatment will change slope of regression.)

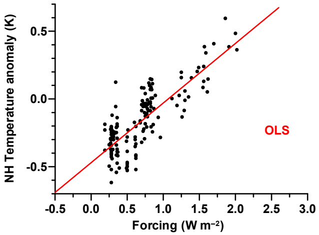
Open Questions: Uncertainties versus slope of regression

Climate Sensitivity $\lambda^{-1} = \Delta T / \Delta R$

OLS

$$Y(i) = a + \lambda^{-1} X(i) \qquad + S_Y(i) Y_{noise}(i)$$

$$SSQ(a, \lambda^{-1}) = \sum [y(i) - a - \lambda^{-1} x(i)]^2$$



minimized

Draper & Smith 1981

Mudelsee et al., unpublished

Open Questions: Uncertainties versus slope of regression

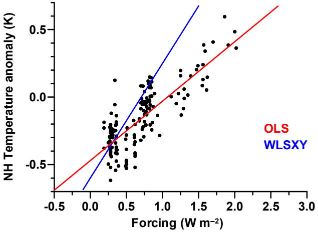
Climate Sensitivity $\lambda^{-1} = \Delta T / \Delta R$

WLSXY

$$Y(i) = a + \lambda^{-1}[X(i) - S_X(i) X_{noise}(i)] + S_Y(i) Y_{noise}(i)$$

$$SSQWXY(a, \lambda^{-1}) = \frac{\sum [y(i) - a - \lambda^{-1} x(i)]^2}{S_Y(i)^2 + \lambda^{-1} S_X(i)^2}$$

minimized



Draper & Smith 1981, Deming 1943, York 1966, Press et al. 1992, Mudelsee 2010

Mudelsee et al., unpublished

The End

Implict versus explicit

Equilibrium temperature change $\Delta T_{\mathrm{E},\infty}$
for a given radiative forcing ΔR (or ΔF or ΔQ)

$$\Delta T_{\mathrm{E},\infty} = \frac{-\Delta R}{\lambda}$$

Here, forcings & feedbacks are only used to calculate $\Delta T_{\mathrm{E},,\infty}$.
They say nothing about CAUSE and EFFECT (leads and lags).

- glacial/interglacial: GHG (ΔR_{GHG}), but GHG are NOT the underlying cause for the temperature change, they contribute to it by changing the radiative budget.
- pure GHG forcing (ΔR_{GHG}) and ice sheet albedo feedback does NOT imply the GHG is causing the changes in the ice sheets.

Because of that we can use whatever we want to (can provide by data) of forcing (explicitly) and everything else as feedback (implictly).

Names

1. Climate sensitivity after IPCC: $\Delta T_{2\times\mathrm{CO}_2}$ (K)
   equilibrium temperature change for doubling CO_2 concentration
2. specific climate sensitivity S (K (W m^{-2})$^{-1}$)

$$S = \frac{\Delta T_{\mathrm{E},\infty}}{\Delta R}$$

(or specific paleo climate sensitivity or Earth system sensitivity)

3 information wanted: which forcing ΔR, temperature ΔT, time slice

$${}^{\mathrm{time}}S^{\Delta T}_{\Delta R} \quad \text{or} \quad S^{\Delta T}_{\Delta R}\text{@time}$$

Example: ice core CO_2 over 800 kyr with δ^{18}O-model-inverted ΔT

$${}^{\mathrm{Pleis}}S^{f(\delta^{18}O)}_{CO_2} \quad \text{or} \quad S^{f(\delta^{18}O)}_{CO_2}\text{@Pleistocene}$$