Measuring Paleo-(Climate)-Sensitivity

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Colloquium "Constraining estimates of future climate variability from geological records"
Royal Netherlands Academy of Arts and Sciences, Amsterdam, 29–31 March 2011

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 $\Delta T_{E,\infty}$ for a given radiative forcing $\Delta R$ (or $\Delta F$ or $\Delta Q$)

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

Climate feedback parameters

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

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

**Initially:** only Planck feedback, others ($\lambda_{\text{else}}$) ignored

$$\Delta T_{E,P} = \frac{-\Delta R}{\lambda_{\text{Planck}}} = \frac{-\Delta R}{-3.2 \text{ W m}^{-2} \text{ K}^{-1}}$$
Equilibrium temperature change $\Delta T_{E,\infty}$ for a given radiative forcing $\Delta R$ (or $\Delta F$ or $\Delta Q$)

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

Climate feedback parameters

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

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

(specific) climate sensitivity $S$

$$S = \frac{\Delta T_{E,\infty}}{\Delta R}$$
Climate sensitivity ($\Delta T_{2 \times CO_2}$)

equilibrium temperature change for doubling CO$_2$ concentration

Problem: When is equilibrium?

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

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

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

$$\frac{\Delta T_{2 \times CO_2}}{\Delta R_{2 \times CO_2}} = S_{2 \times CO_2} = 0.6 - 1.2 \text{ K (W m}^{-2})^{-1}$$
Other Approaches (e.g. Hansen et al, 2007, 2011)

1. Use LGM for $\Delta T$ and $\Delta R \Rightarrow S = 0.75 \pm 0.25 \text{ K (W m}^{-2})^{-1}$
Other Approaches (e.g. Hansen et al, 2007, 2011)

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

(c) Climate Forcing

(a) Observed Dome C and Calculated Temperatures
Other Approaches (e.g. Hansen et al, 2007, 2011)

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

Points for improvements:

1. $\Delta T$ (observed) is $0.5 \times$ EPICA-Dome-C $\Delta T$ which is wrong, because the polar amplification changes with climate / time.
2. There is no global time series of $\Delta T$, 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 $\Delta R$ as far as possible out of data sets.

\[ \Delta T_{E,\infty} = \frac{-\Delta R}{\lambda} \]

Two applications:

1. **Late Pleistocene:**
   (a) Refine $\Delta R$ over last 800 kyr
   (b) For LGM: use a $\Delta T$ and our $\Delta R \Rightarrow S = \Delta T / \Delta R$.

2. **Late Cenozoic:**
   (a) Use data-based $\Delta T$ and constant $S$ for $CO_2 = f(\Delta R)$
   (b) Use $\Delta T$ and ice core $CO_2$ to calculate variability in $S$
Our Approach: Processes change in the radiative budget based on data

Considered paleo changes: $I$: incoming solar radiation; $G$: GHG; $\alpha$: albedo
Simplified View on the Radiative Budget

Considered paleo changes:
- $I$: incoming solar radiation;
- $\alpha_S$: surface: land ice, snow, sea ice, vegetation
- $\alpha_A$: atmosphere dust
- $\alpha_P$: greenhouse gases

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$_2$O)

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

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

(Köhler et al., QSR 2010)
Total radiative forcing

LGM: $-12.4 \pm 3.2 \text{ W m}^{-2}$. $\Delta T_{E,P}: -3.9 \pm 0.8 \text{ K}$. With feedbacks (water vapour, clouds, lapse rate) $\lambda_{\text{today}} = 1.65 \pm 0.49 \text{ W m}^{-2} \text{ K}^{-1}$:

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

$\Rightarrow$ Feedbacks strength $\lambda_{\text{else}}$ and then also $S$ depends on climate.

(Köhler et al., QSR 2010)
Asymmetry in $S$ for warming and cooling (Hargreaves et al., 2007).

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

$\Delta T = 5.8 \pm 1.4 \text{K}$

(Hargreaves et al., 2007)

$\Delta T_{2 \times \text{CO}_2} = 1.4 - 5.2 \text{ K}$

(Köhler et al., QSR 2010)
Conclusions I

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

Wanted Improvements

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

1. Conceptual Basics

2. Application 1: Late Pleistocene — last 800 kyr

3. Application 2: Late Cenozoic — Last 20 Myr
Necessary Ingredients

Two out of three

\( \Delta T \)

\( \Delta R \)

S

are necessary to calculate the third after

\[ S = \frac{\Delta T}{\Delta R} \]
Deconvolute stacked benthic $\delta^{18}O$ into climate variables ($\Delta T_{\text{deep o}}, \Delta T_{\text{atm}}$ (40–80°N), size of ice sheets, sea level, snow cover)

(Bintanja et al., 2005; de Boer et al., 2011)
\[ \Delta T = f(\text{benthic } \delta^{18}O) \]

(\text{after Bintanja et al., 2005; van de Wal et al., 2011; de Boer et al., 2011})
CO₂: proxy diversity

after van de Wal et al., 2011 CPD
Use data-based $\Delta T = f(\delta^{18}O)$, assume constant $S = \Delta T / \Delta R$ to calculate $CO_2 = f(\Delta R)$

$\Rightarrow \Delta R = \Delta T / S$
Relationship $\Delta T_{NH} - \text{CO}_2$

Graph showing the relationship between $\Delta T_{NH}$ (K) and CO$_2$ (ppmv) with various datasets and correlation coefficients:

- Alkenones (Pagani 2005), $r^2 = 0.24$
- Alkenones (Pagani 2010), $r^2 = 0.03$
- $\delta^{11}B_p$ (Pearson 2000), $r^2 = 0.00$
- Stomata (Kürschner 2008), $r^2 = 0.16$
- B/Ca (Tripati 2009), $r^2 = 0.39$
- Alk+$\delta^{11}B_s$ (Seki 2010), $r^2 = 0.75$
- $\delta^{11}B_h$ (Hönisch 2009), $r^2 = 0.62$
- Ice cores (EDC+Vostok), $r^2 = 0.61$

Graph notes:
- (after van de Wal et al., 2011 CPD)
ΔT_{NH}—CO₂ 1: Empirical Relationship

resampled and binned data in intervals of Δ(ΔT_{NH}) = 1 K

C = 39 ± 4K regression slope from modelled ΔT_{NH} and CO₂ data

(van de Wal et al., 2011, CPD)
$\Delta T_{NH} - \text{CO}_2$ 2: Theoretical Relationship

\[ \Delta T = S \cdot \Delta R \]

\[ \Delta T_{NH} = C \cdot \ln \frac{\text{CO}_2}{\text{CO}_2,\text{ref}} \quad \text{with} \quad C = \frac{\alpha \beta \gamma S_C}{1 - f} \]

**LGM parameters:**

\[ \alpha = \Delta T_{NH}/\Delta T_{global} = 15 \text{ K}/6 \text{ K} = 2.5 \]

\[ \beta = 5.35 \text{ W m}^{-2}: \text{radiative forcing of CO}_2 \]

\[ \gamma = 1.3: \text{enhancement factor for non-CO}_2 \text{ GHG (CH}_4, \text{ N}_2\text{O)} \]

\[ S_C = 0.72 \text{ K (W m}^{-2})^{-1}: \text{Charney climate sensitivity (fast feedbacks: Planck, water vapour, lapse rate, clouds, sea ice, albedo)} \]

\[ f = 0.72: \text{feedbacks of slow processes (land ice, dust, vegetation)} \]

\[ C = 43K \text{ theoretical calculation based LGM data and constant } S \]

For comparision:

\[ \text{pure } S_{\text{Charney}} (f = 0; \gamma = 1; \alpha = 1) \Rightarrow C_C = 3.9 \text{ K and } \Delta T_{global} = 2.7 \text{ K} \]

(van de Wal et al., 2011, CPD)
Develop relationship atmospheric $\Delta T_{NH}$—$CO_2$

\[ \Delta T_{NH} = C \cdot \ln \frac{CO_2}{CO_2,\text{ref}} \text{ with } C = \frac{\alpha \beta \gamma S}{1 - f} \]

Two independent approaches to calculate the slope:

1. $C = 39 \pm 4K$ regression slope from modelled $\Delta T_{NH}$ 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
**Conceptual Basics**

CO₂ reconstructions, the last 20 Myr

- alkenones (Pagani 2005)
- alkenones (Pagani 2010)
- δ¹¹Bᵢ (Pearson 2000)
- Stomata (Kürschner 2008)
- B/Ca (Tripati 2009)
- Alk+δ¹³Bₛ (Seki 2010)
- δ¹¹Bₜ (Hönisch 2009)
- ice cores (EDC+Vostok)

Glacial/interglacial amplitudes captured, details wrong

after van de Wal et al., 2011 CPD
Assumption: relation $\text{CO}_2 - \Delta T$ unchanged with time!!!

after van de Wal et al., 2011 CPD
Use data-based $\Delta T = f(\delta^{18}O)$, and the best constrained $\Delta R = f(CO_2, \text{ice core})$ to calculate the variability in $S = f(T)$

$\Rightarrow S = \Delta T / \Delta R$
Alternative: $S = f(T)$ based on ice core data

![Graph showing correlation between temperature and CO2 concentration](image)

- Alkenones (Pagani 2005), $r^2 = 0.24$
- Alkenones (Pagani 2010), $r^2 = 0.03$
- $\delta^{11}B_p$ (Pearson 2000), $r^2 = 0.00$
- Stomata (Kürschner 2008), $r^2 = 0.16$
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- Ice cores (EDC+Vostok), $r^2 = 0.61$

(after van de Wal et al., 2011 CPD)
Alternative: $S = f(T)$ based on ice core data

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

interglacials: $\Delta T \sim 0$; $\Delta R \sim 0$; synchronisation!!!
Revise Theoretical Relationship $\Delta T_{NH} - CO_2$

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

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

$C = 43K$ theoretical calculation based LGM data and constant $S$
Revise Theoretical Relationship $\Delta T_{NH} - CO_2$

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

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

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

Revision:

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

$\Rightarrow$ Ice cores suggest a variability in $S$ of $\pm 26\%$, thus $C = 43 \pm 11K$
Revise CO₂ 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$^{-2}$

$\Delta T = 5.8 \pm 1.4$ K

(Schneider v. Deimling et al., 2006)

factor $f = 0.85 \pm 0.2$

(Hargreaves et al., 2007)

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

Asymmetry in $S$ (scaling factor $f$) not considered so far.
Revise CO$_2$, consider unsymmetry in S (here: $f = 0.85$)
Conclusions II

1. If one assumes constant $S \Rightarrow \text{CO}_2$ can be calculated out of $\Delta T$.
2. Alternatively, if we believe in a $\Delta T$ we can obtain a climate-dependent $S$ from the ice core $\text{CO}_2$.
3. For which forcing $\Delta R$ is $S$ calculated? e.g. $S_{\text{Charney}}, S_{\text{all}}, S_{\text{CO}_2}$.
4. Approach is weak in MIS 5, 7, 9, 11 with both $\Delta R$ 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

\[ \lambda^{-1} = \frac{\Delta T}{\Delta R} \]

OLS

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

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

minimized

Draper & Smith 1981

Mudelsee et al., unpublished
Open Questions:
Uncertainties versus slope of regression

Climate Sensitivity

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

WLSXY

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

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

minimized


Mudelsee et al., unpublished
The End
Implicit versus explicit

Equilibrium temperature change $\Delta T_{E,\infty}$ for a given radiative forcing $\Delta R$ (or $\Delta F$ or $\Delta Q$)

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

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

- glacial/interglacial: GHG ($\Delta R_{GHG}$), but GHG are NOT the underlying cause for the temperature change, they contribute to it by changing the radiative budget.
- pure GHG forcing ($\Delta R_{GHG}$) 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 (implicitly).
Climate sensitivity after IPCC: $\Delta T_{2\times CO_2}$ (K)
equilibrium temperature change for doubling $CO_2$ concentration

Specific climate sensitivity $S$ (K $(W \ m^{-2})^{-1}$)

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

(or specific paleo climate sensitivity or Earth system sensitivity)

3 information wanted: which forcing $\Delta R$, temperature $\Delta T$, time slice

$$time \ S_{\Delta T}^{\Delta R} \quad or \quad S_{\Delta R}^{\Delta T}@time$$

Example: ice core CO$_2$ over 800 kyr with $\delta^{18}O$-model-inverted $\Delta T$

$$S_{CO_2}^{f(\delta^{18}O)} \quad or \quad S_{CO_2}^{f(\delta^{18}O)}@Pleistocene$$