The inconstancy of transient climate response

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Results mostly from Gregory, Andrews and Good (2015)  
and Gregory and Andrews (2016)
Climate sensitivity, forcing and feedback

Equilibrium climate sensitivity applies to a particular forcing (2×CO₂). It is more generally useful because of the separation of forcing and feedback:

\[ T \propto F \quad \text{or} \quad F = \alpha T \]

Radiative forcing is the net heat flux into the climate system in the presence of the forcing agent, before climate change has occurred.

- **Radiative forcing**
  - \( F \) (W m\(^{-2}\))
  - depends on the forcing agent

- **Climate feedback parameter**
  - \( \alpha \) (W m\(^{-2}\) K\(^{-1}\))
  - is a property of the climate system
Radiative forcing is time-dependent

- AR5 anthropogenic + natural
- AR5 volcanic

Radiative forcing $F$ (W m$^{-2}$)

Year

- Krakatoa
- Santa Maria
- Katmai
- Agung
- El Chichon
- Pinatubo

CO$_2$ is the largest contributor to the net radiative forcing

**IPCC AR5 Fig SPM5**

<table>
<thead>
<tr>
<th>Emitted compound</th>
<th>Resulting atmospheric drivers</th>
<th>Radiative forcing by emissions and drivers</th>
<th>Level of confidence</th>
</tr>
</thead>
<tbody>
<tr>
<td>CO$_2$</td>
<td>CO$_2$</td>
<td>1.68 [1.33 to 2.03]</td>
<td>VH</td>
</tr>
<tr>
<td>CH$_4$</td>
<td>CO$_2$, H$_2$O, O$_3$, CH$_4$</td>
<td>0.97 [0.74 to 1.20]</td>
<td>H</td>
</tr>
<tr>
<td>Halocarbons</td>
<td>O$_3$, CFCs, HCFCs</td>
<td>0.18 [0.01 to 0.35]</td>
<td>H</td>
</tr>
<tr>
<td>N$_2$O</td>
<td>N$_2$O</td>
<td>0.17 [0.13 to 0.21]</td>
<td>VH</td>
</tr>
<tr>
<td>Anthropogenic</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>CO</td>
<td>CO$_2$, CH$_4$, O$_3$</td>
<td>0.23 [0.16 to 0.30]</td>
<td>M</td>
</tr>
<tr>
<td>NMVOC</td>
<td>CO$_2$, CH$_4$, O$_3$</td>
<td>0.10 [0.05 to 0.15]</td>
<td>M</td>
</tr>
<tr>
<td>NO$_x$</td>
<td>Nitrate, CH$_4$, O$_3$</td>
<td>-0.15 [-0.34 to 0.03]</td>
<td>M</td>
</tr>
<tr>
<td>Aerosols and precursors</td>
<td>Mineral dust, Sulphate, Nitrates, Organic carbon, Black carbon</td>
<td>-0.27 [-0.77 to 0.23]</td>
<td>H</td>
</tr>
<tr>
<td></td>
<td>Cloud adjustments due to aerosols</td>
<td>-0.55 [-1.33 to -0.06]</td>
<td>L</td>
</tr>
<tr>
<td></td>
<td>Albedo change due to land use</td>
<td>-0.15 [-0.25 to -0.05]</td>
<td>M</td>
</tr>
<tr>
<td>Natural</td>
<td>Changes in solar irradiance</td>
<td>0.05 [0.00 to 0.10]</td>
<td>M</td>
</tr>
<tr>
<td></td>
<td>Total anthropogenic RF relative to 1750</td>
<td>2.29 [1.13 to 3.33]</td>
<td>H</td>
</tr>
<tr>
<td></td>
<td></td>
<td>1.25 [0.64 to 1.86]</td>
<td>H</td>
</tr>
<tr>
<td></td>
<td></td>
<td>0.57 [0.29 to 0.85]</td>
<td>M</td>
</tr>
</tbody>
</table>

W m$^{-2}$ relative to 1750
$N = F - \alpha T$

$N$ is the net heat flux at the TOA i.e. into the climate system.

$N \approx$ net heat flux into the ocean.

In the unperturbed steady state $N = F = 0$ and $T = 0$.

While the climate is changing, $N \neq 0$.

Ocean heat uptake mitigates the rate of surface climate change, $T = (F - N)/\alpha$.

In the perturbed steady state $N = 0$ and $F = \alpha T$.

Equilibrium climate sensitivity $ECS = F(2\times CO_2)/\alpha$ is likely within 1.5 to 4.5 K.
Two-layer model for transient climate response

\[ N = F - \alpha T \]

\[ N = C_u \, dT/dt + \gamma (T - T_d) \]

Upper ocean \( T \) of small heat capacity

Deep ocean \( T_d \) of large heat capacity

Gregory et al. (2000)
Held et al. (2010)

Geoffroy et al. (2013)
Simpler model for transient climate response

$N = C_u \frac{dT}{dt} + \gamma (T - T_d)$

$N = \kappa T = F - \alpha T$

$F = (\alpha + \kappa) T$

Transitional climate response $TCR = F(2\times CO_2)/(\alpha + \kappa)$

The transitional climate response (TCR) is likely in the range 1.0 to 2.5 K.

Transient climate response parameter $TCRP = 1/(\alpha + \kappa)$ in K (W m$^{-2}$)$^{-1}$

(Held et al., 2010; Gregory et al., 2015)
TCRP increases under 1pctCO2

If $F \propto T$ and
$F \propto \log(\text{CO}_2) \propto t$,
we expect $T \propto t$
Treat $F(t)$ for 1pctCO2 as a succession of steps
Step (Good et al., 2011) model

[Graph showing temperature change over years]

Good et al. (2013)
Inconstancy of TCRP partly predicted by step model
Inconstancy of TCRP partly predicted by step model
Inconstant $dR/dT$ in 1pctCO2
Climate feedback may vary under constant $F$

$$N = F - \alpha T$$

predicts a straight line for constant $\alpha$

The blue and green slopes are quite similar

Andrews et al. (2015)
Historical $\alpha$ from SST-forced AGCM experiments

$$N = F - \alpha T$$

$F$\hspace{0.5cm} $\alpha T$

Upper ocean $T$ of small heat capacity

Supplied via SST forcing

AOGCM

AGCM

$N = -\alpha T$

$\alpha T$

Deep ocean

amip-piForcing is an AGCM experiment for CFMIP3/CMIP6 (coordinated by Tim) with AMIP sea-surface BCs for 1870-present and constant pre-industrial forcing.
Historical $T(t)$ from obs and amipPiForcing expts
Climate feedback from amipPiForcing expts

\[ N = -\alpha T \]
\[ -\frac{dN}{dT} \text{ from HadGEM2-A amipPiForcing} \]
$-dN/dT$ from HadCM3-A amipPiForcing
- $\frac{dN}{dT}$ from HadCM3-A amipPiForcing
\(-dN/dT\) from ECHAM6 amipPiForcing
Components of $-\frac{dN}{dT}$ for amipPiForcing
Climate feedback from HadCM3-A amipPiForcing

- **Change in global mean net downward radiative flux** $N$ (W m$^{-2}$)

- **Change in global mean surface air temperature** $T$ (K)

Legend:
- **All years**: $1.66\pm0.01$ W m$^{-2}$ K$^{-1}$
- **1926-1955**: $1.16\pm0.07$ W m$^{-2}$ K$^{-1}$
- **1979-2008**: $2.21\pm0.08$ W m$^{-2}$ K$^{-1}$
Regression slope of local $T$ versus global $T$ (K K$^{-1}$)

AMIP SST dataset

1926-1955

1978-2008
\(-dN/dT\) from HadCM3-A amipPiForcing
Regression slope of local $T$ versus global $T$ (K K$^{-1}$)

AMIP SST dataset
1926-1955

1978-2008

abrupt4xCO2 years 1-20
HadCM3

CMIP5 mean
\(-dN/dT\) from HadCM3-A amipPiForcing
Climate feedback from AMIP and abrupt4xCO2

Using $\alpha T = F - N$ for AMIP
Conclusions

Under 1pctCO2, The transient climate response parameter (TCRP increases), by about 20% over 140 years, because

- ocean heat uptake efficiency declines as time passes.
- climate feedback decreases, or forcing rises more rapidly than logarithmically, as CO$_2$ increases.

Consequently scaling the TCR gives an underestimate of projected $T$.

Under abrupt4xCO2, climate feedback decreases (climate sensitivity increases) as $T$ rises and time passes, due mostly to the effect of changing SST patterns on SW cloud feedback, but this is a small effect for TCRP.

Time-dependent historical SSTs produce an effective climate sensitivity of ~2 K in three AGCMs, having large decadal variation that is partly model-dependent, explanations not yet clear (perhaps varying forcing or unforced variability).

These and other AMIP AGCMs give a historically unusually low effective climate sensitivity of ~1.5 K for 1979-2008, considerably less than for abrupt4xCO2.

These results may help to relieve the apparent contradiction in the AR5 between the larger values of effective climate sensitivity diagnosed from AOGCMs and the smaller values inferred from historical climate change.