Towards a theory for Earth’s Climate Sensitivity (ECS)

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40 years of Consensus-building

ECS (K)

1860 1880 1900 1920 1940 1960 1980 2000 2020

1.5 3.0 4.5 6.0

We believe, therefore, that...

Charney

Augustsson and Ramanathan (1977)

Manabe and Wetherald (1967)

AR4

AR5

Likely, $p > 66$

p < 10%

p < 5%

Stefan (1879), Boltzmann (1884)

Arrhenius (1896)

Callendar (1938)
Ansatz

1. Define a best estimate
2. Pose arguments why other best estimates are consistent
3. Explain what it would take for it to be wrong
Robust understanding: 2.6 K

Assumed unbiased best estimates (squares): 2.2-2.8 K

“I believe, therefore, that ECS is 2.6 K”
Best estimates of ECS vary

- **ECS ~ 4-5 K**
- **ECS ~ 2-3 K**
- **ECS ~ 1.6-2.0 K**

CMIP5 models: 3.4 K  
Emergent constraints: ~ 4 K

Illustration modified from Glen Fergus
Best estimates of ECS vary

- ECS ~ 4-5 K
- ECS ~ 2-3 K
- ECS ~ 1.6-2.0 K
- CMIP5 models: 3.4 K
- Emergent constraints: ~ 4 K
Zero-layer diagnostic model:

\[ ECS = F_2 \times \frac{\Delta T}{\Delta F - \Delta Q} \]

Instrumental Record ECS

ECS: 1.79 [1.08 - 4.44]

Instrumental Record ECS

Zero-layer diagnostic model:

\[ ECS = \frac{F_{2\times CO_2}}{\Delta T} \]

- Aerosol forcing (Marvel et al. 2015)
- Time-dependent feedback (Armour 2017)
- Observational issues (Richardson et al. 2016)
- Early period heat uptake (Huybers, in prep)

Underestimated Aerosol Cooling?

- It will take about -1.3 Wm⁻² to get ECS = 2.6 K
- Such strong cooling is being contested
Time Dependent Feedback less Effective

Clouds cooled the Earth:

- Remote warming of troposphere
- Strengthening inversion
- More low-level clouds

Temperature trend 1980–2005 (K per year)

CMIP5 model range, Geoffroy et al. 2013
Zero-layer assumption
Two-layer assumption

Ocean heat uptake efficacy, ε

ECS

Feedback is the rate of change in the Earth's radiation budget with global mean surface temperature change. Climate sensitivity is determined by the Earth's equilibrium climate sensitivity (0.87°C per W m⁻²), which is the global warming that occurs when the atmospheric carbon dioxide has been doubled. The global warming that occurs over a long time is likely to be a function of two factors: the radiative forcing from the increase in atmospheric carbon dioxide and the less certain climate change feedback.

The slow instrumental-record warming is consistent with lower-end climate sensitivity. Simulations and observations now show that changing sea surface temperature patterns could have a role in the Earth's climate. Writing explanations have been proposed. Where sea surface temperatures are kept weak by various atmospheric motions (red curly symbols). Therefore, when the warmest regions warm more than the colder regions (west Pacific, the black and red curves are idealized vertical temperature profiles before and after warming, respectively). Where sea surface temperatures are relatively cool, the lowermost mixed-layer of the atmosphere is also cool, but a sharp rise in temperature (inversion) occurs at 1–2 km height and is apparent. The map shows the trend in annual mean sea surface temperature for the period 1980–2005 from the HadSST3 dataset.

Cloud feedbacks can act as the main source of climate sensitivity variability and clouds. Cloud feedbacks are usually identified as the main source of climate sensitivity variability. However, there is a puzzling result: it is unclear how climate sensitivity in a model can be so crucially different depending on the assumptions of the feedback functions. Feedback may not be constant over the 1950–1970s until now; a change that corresponds to a halving of the climate sensitivity range.

In climate models that use sea surface temperature warming as the main source of uncertainty, remote warming of the troposphere could be the cause of the discrepancy. Simulations and a number of efforts and a review of the range of the model results now show that changing sea surface temperature patterns could have a role in the Earth's climate.
State Dependent Feedback

TOA net imbalance (Wm$^{-2}$) vs. Surface temperature change (K)

Meraner, Mauritsen and Voigt (2013)

ECHAM6.0, mixed-layer ocean
Explanatory model (green):
- Constant RH
- Moist adiabat (RCE)
- Constant tropopause temperature

Meraner, Mauritsen and Voigt (2013)
State Dependent Feedback

- Schmittner et al. (2011), Last glacial maximum
- Hargreaves et al. (2012), Last glacial maximum
- Mauritsen and Pincus (2017), Instrumental record warming
- Ongoing work, Instrumental record variability
- Hargreaves and Annan (2016), Pliocene
- Shaffer et al. (2016), Late Paleocene
- Shaffer et al. (2016), PETM
- Rohling et al. (2012), LGM only
- CMIP5, Climate models
- Brient and Schneider (2016), low clouds
- Charney report (1979)
- Bloch-Johnson et al. (2015)
- Meraner et al. (2013)
I believe, therefore, that... ECS is 2.6 K (2.0-3.5)
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The hypothesis requires the following to be consistent with current evidence:

- ECS rises in a warmer climate
- Feedback must be highly time-dependent
  and/or
- Aerosol cooling strongly negative
- Climate models miss negative feedback(s)
Process-level approach: \( ECS = -\frac{F_{2x}}{\sum \lambda_i} \)
**Multiple Lines of Evidence**

- **Figure 2**: Equilibrium climate sensitivity (°C) with different lines of evidence.
- **Figure 3**: Transient climate response (°C) without ocean heat uptake.

**Equilibrium climate sensitivity (°C)**

- **Figure 2**: Various models and observational data showing the range of sensitivities.
- **Figure 3**: Comparisons between models and observations for lower limit of sensitivity.

**Transient climate response (°C)**

- **Figure 3**: Different models and observational data showing the range of transient responses.

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**Figure 2**:

- Averbeck et al. (2013): mean and likely range.
- Kao and Lin (2015): 90% range of all models.

**Figure 3**: Different models and observational data showing the range of transient responses.

- Cherchi et al. (2016): mode and 90%.
- Koberle et al. (2011): mean and likely range.

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**Table 1**: Summary of climate sensitivity estimates.

<table>
<thead>
<tr>
<th>Model</th>
<th>Method</th>
<th>Time Period</th>
<th>Best Estimate</th>
<th>Likely Range</th>
</tr>
</thead>
<tbody>
<tr>
<td>AR5/6</td>
<td>Houghton (1990)</td>
<td>1900-2000</td>
<td>3°C</td>
<td>2°C-4°C</td>
</tr>
<tr>
<td>AR5/6</td>
<td>Houghton (1990)</td>
<td>1970-2000</td>
<td>2°C</td>
<td>1.5°C-2.5°C</td>
</tr>
<tr>
<td>AR5/6</td>
<td>Houghton (1990)</td>
<td>Present</td>
<td>1.5°C</td>
<td>1°C-2°C</td>
</tr>
<tr>
<td>AR5/6</td>
<td>Houghton (1990)</td>
<td>Future</td>
<td>1°C</td>
<td>0.5°C-1.5°C</td>
</tr>
</tbody>
</table>

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**Figure 1**: Range of observed and modeled climate sensitivities.

- Different models and observational data showing the range of observed sensitivities.

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**Background Information**

- Friedlingstein et al. (2014): best estimate.

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**Discussion**

- Knutti et al. (2017): mean and likely range.
- leverett et al. (2016): mean and standard deviation.

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**Conclusion**

- Future work will be needed to further refine our understanding of climate sensitivities.
- Continued research is essential to bridge the gap between current understanding and future projections.
Time- and State Dependence

![Graphs showing temperature vs. imbalance with different curves for varying states and time periods.](image-url)
Low-level cloud emergent constraints

\[ \lambda = \lambda_T + \lambda_W + \lambda_C + \lambda_A + \ldots \]

\[ r = 0.68/0.70 \]

\( \text{LTMI, } (S + D) \)

\( \text{Climate sensitivity} \)

\( \text{ECS (K)} \)

\( \text{Deseasonalized } \delta T \)

\( \text{Obs. PDF} \)

\( \text{HS models} \)

\( \text{LS models} \)

Sherwood et al. (2014)

Brient and Schneider (2016)
A missing iris-effect

Figure 1 | Illustration of the tropical atmospheric circulation.

- Strong OLR
- Weak OLR
- Iris expansion
- Rising tropopause
- Moist and cloudy
- Dry and clear
- Radiative cooling
- Latent heating

Mauritsen and Stevens (2015)
A missing iris-effect

Mauritsen and Stevens (2015)
Climate sensitivity is determined by a series of processes in the Earth system [20]. A forced system will experience an increase in temperature due to the warming of the troposphere relative to the surface. To estimate this increase, we can use the ECS (Equilibrium Climate Sensitivity) parameter, which is defined as:

\[ \text{ECS} = \frac{-F_{2x}}{\lambda} \]

where \( \lambda \) is the sum of various feedback terms:

\[ \lambda = \lambda_T + \lambda_W + \lambda_C + \lambda_A + \ldots \]

Candidates for biased/missing feedbacks are:

- Ozone feedback, about -0.1 Wm^-2/K
- Iris-effect, about -0.3 Wm^-2/K

Together, these feedbacks are enough to move the CMIP5 model collection mean from 3.4 K to 2.5 K.

Nowack et al. (2014), Chiodi and Polvani (2016)
Mauritsen and Stevens (2015), Williams and Pierrehumbert (2017)
distribution is strongly asymmetric, we use the gamma distribution to represent the probability density function (pdf). This approach allows us to calculate the 95% confidence interval for the pdf, which is an area around the central value where 95% of the probability mass lies. This notation, used throughout this paper, the central value indicates the maximum likelihood estimate in degrees Celsius and the outer values represent the limits of the interval.

We use this distribution as a typical representative of this class of models. It is important to note that the net forcing is small, then climate sensitivity would have to be very high to explain the observed warming. Given this constraint, we consider the issue of model error, which suggests that the uncertainties on the three estimates may not be wholly independent. On the other hand, moreover implies that the uncertainties on the three estimates could themselves be combined.

The shape of the likelihood function (see Figure 1). The distribution is described by (1.5, 3, 6). We take this distribution as our prior with which additional information in the form of likelihood functions will be combined into an estimate which has significantly tighter limits and scale parameters 3.2 and 1.36 (see Figure 1). We take this distribution as a parsimonious representation, using shape and scale parameters 3.2 and 1.36 (see Figure 1). We take this distribution as a parsimonious representation, using shape and scale parameters 3.2 and 1.36 (see Figure 1). We take this distribution as a parsimonious representation, using shape and scale parameters 3.2 and 1.36 (see Figure 1). We take this distribution as a parsimonious representation, using shape and scale parameters 3.2 and 1.36.

## A Statistics Approach

### 3.1. 20th Century Warming

Figure 1. volcanic cooling (1.5, 3, 6). Blue dot-dashed line: 20th century warming (1, 3, 10). Blue dotted line: combination of volcanic and anthropogenic forcing. Red solid line: combination of volcanic and anthropogenic forcing with additional information from paleoclimate records.

### 3.2. Volcanic Cooling

Figure 2. Short-term cooling following volcanic eruptions ranging from 5.2–7.7°C for each individual eruption which in each case gives a high likelihood to values close to 3°C. A comparison with the observed cooling produces a plausible range of about (1.8, 2.8, 4.4). However, their analysis does not combined into an estimate which has significantly tighter limits and scale parameters 3.2 and 1.36 (see Figure 1).

### 3.3. Last Glacial Maximum

The short-term large-scale cooling following volcanic eruptions has also recently been used to estimate climate sensitivity based on various observational constraints. Blue dashed line: 20th century warming (1, 3, 10). Blue solid line: combination of volcanic and anthropogenic forcing with additional information from paleoclimate records. Blue dot-dashed line: volcanic cooling (1.5, 3, 6). Blue dotted line: combination of volcanic and anthropogenic forcing with additional information from paleoclimate records.

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### 3.4. Observational Constraints

The short-term large-scale cooling following volcanic eruptions has also recently been used to estimate climate sensitivity based on various observational constraints. Blue dashed line: 20th century warming (1, 3, 10). Blue solid line: combination of volcanic and anthropogenic forcing with additional information from paleoclimate records. Blue dot-dashed line: volcanic cooling (1.5, 3, 6). Blue dotted line: combination of volcanic and anthropogenic forcing with additional information from paleoclimate records.

### 3.5. Model Error

The short-term large-scale cooling following volcanic eruptions has also recently been used to estimate climate sensitivity based on various observational constraints. Blue dashed line: 20th century warming (1, 3, 10). Blue solid line: combination of volcanic and anthropogenic forcing with additional information from paleoclimate records. Blue dot-dashed line: volcanic cooling (1.5, 3, 6). Blue dotted line: combination of volcanic and anthropogenic forcing with additional information from paleoclimate records.

### 3.6. Summary

The short-term large-scale cooling following volcanic eruptions has also recently been used to estimate climate sensitivity based on various observational constraints. Blue dashed line: 20th century warming (1, 3, 10). Blue solid line: combination of volcanic and anthropogenic forcing with additional information from paleoclimate records. Blue dot-dashed line: volcanic cooling (1.5, 3, 6). Blue dotted line: combination of volcanic and anthropogenic forcing with additional information from paleoclimate records.

### 3.7. Conclusion

The short-term large-scale cooling following volcanic eruptions has also recently been used to estimate climate sensitivity based on various observational constraints. Blue dashed line: 20th century warming (1, 3, 10). Blue solid line: combination of volcanic and anthropogenic forcing with additional information from paleoclimate records. Blue dot-dashed line: volcanic cooling (1.5, 3, 6). Blue dotted line: combination of volcanic and anthropogenic forcing with additional information from paleoclimate records.

### 3.8. References

- Annan and Hargreaves (2006)
- Stevens et al. (2016)
- Knutti et al. (2017)
- Sherwood et al., in preparation
A Mechanistic View

• How does ECS depend on State?
• How do feedbacks vary with Time?
• What are models missing?

Approximate temperature relative to present (K)

ECS (K)

Schmittner et al. (2011), Last glacial maximum
Hargreaves et al. (2012), Last glacial maximum
Mauritsen and Pincus (2017), Instrumental record warming
Ongoing work, Instrumental record variability
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Rohling et al. (2012), LGM only
CMIP5, Climate models
Brient and Schneider (2016), low clouds
Charney report (1979)

Blein-Johnson et al. (2015)
Meraner et al. (2013)
4.1 Data uncertainty

To calculate an estimate for equilibrium climate sensitivity, we combine the model estimates for climate sensitivity and the warming at the mPWP, together with the PRISM3 estimate of tropical ocean temperature change, using the approach described by Hargreaves et al. (2013). As in Hargreaves and Annan (2016), we assume a linear relationship. The PRISM3 reconstruction does not include an estimate of uncertainty in the reconstruction. Initially we take a value of 0.4 as our initial value of uncertainty in the PlioMIP Experiment 1 SST field has not been objectively estimated, and our initial value of 0.4 is at the low end of (and extending to) the full range of models that contributed to the resulting estimate for the equilibrium climate sensitivity of 1.9–3.7°C.

It is of course essential to test the sensitivity of our result to data uncertainty. As mentioned above, however, the size of the uncertainty estimate substantially to 1.0 (dot-dashed blue and black lines in Fig. 2). As these tend to be low sensitivity models, the uncertainty estimate at this level of sensitivity range covers the full range of model values with an extremely low value has relatively little effect on the resulting sensitivity estimate (which only narrows marginally with values of 0.1 and 1.0 (solid), and 0.4 (dot-dashed). Black arrows of the corresponding type show the resulting sensitivity estimates.

4.2 Forcing uncertainty

To focus on averages over larger spatial scales where we can reasonably expect the models to have some skill (since they fail to do this for other time periods of palaeoclimate), it seems reasonable to conclude that much of the model–data discrepancy here is due to uncertainties in the analysis of the data points. Furthermore, we do not expect models to be able to reliably estimate spatial variation to some extent, it seems reasonable to conclude that much of the model–data discrepancy is at least in the tropics. It would be very useful to have more reconstructions for the mPWP.

A major issue in simulating the mPWP is that the atmosphere–ocean coupling is likely to be weaker in the tropical Pacific during the mPWP compared to today. As tropical Pacific SSTs are strongly affected by the simulations of tropical Pacific SSTs and tropical Pacific wind stress coupled in the models, we focus on the tropical Pacific SSTs used in the PRISM3 reconstructions for the mPWP.

Further, we combine the model estimates for climate sensitivity and the warming at the mPWP, together with the PRISM3 estimate of tropical ocean temperature change, using the approach described by Hargreaves et al. (2013). As in Hargreaves and Annan (2016), we assume a linear relationship. The PRISM3 reconstruction does not include an estimate of uncertainty in the reconstruction. Initially we take a value of 0.4 (solid), and 1.0 (dot-dashed). Black arrows of the corresponding type show the resulting sensitivity estimates.