Response of the Structure of the Atmosphere to Global Warming

What is robust, what is delicate?

Geoffrey K. Vallis

University of Exeter

J. Kidston, C. Cairns, P. Zurita-Gotor

May, 2015
Just because a problem is important is not a reason for not studying it.
Global Warming

Over last century
(NASA/HadCRUT)

Over last millennium
(Mann et al)

What will be the effects of that warming?
Changes that involve thermodynamics and radiation are `robust'.

Changes that involve dynamics are less certain and possibly less robust.

Robust: If you know the parameters and the forcing you can calculate the response reasonably well. No sensitive dependence on parameters.
Thesis

Changes that involve thermodynamics and radiation are `robust'.

Changes that involve dynamics are less certain and possibly less robust.

Robust: If you know the parameters and the forcing you can calculate the response reasonably well. No sensitive dependence on parameters.

Two practical measures:

- Consistency of response of a variety of models.
- An underlying physical mechanism that is not structurally unstable.

Today, our interests are twofold

1. The vertical structure of the atmosphere.
   - The height of the tropopause and stratospheric cooling.

2. The latitudinal structure of the circulation.
   - Expansion of the Hadley Cell
   - Shifts of the westerlies.
Q. How do we know global warming is not just natural variability?

1. No known natural mechanism that is consistent with the observations.

2. In particular, record of ocean heat content. Ocean is not giving up heat to the atmosphere.

Rather, the ocean is warming **because** it is taking up heat *from* the atmosphere.
Q. How do we know global warming is not just natural variability?

1. No known natural mechanism that is consistent with the observations.

2. In particular, record of ocean heat content. Ocean is not giving up heat to the atmosphere.

   Rather, the ocean is warming because it is taking up heat from the atmosphere.

**Troposphere, Stratosphere and Tropopause Height**

**US standard atmosphere**

- **Troposphere:** A region of fast dynamics in which the stratification is set dynamically.
- **Stratosphere:** The region above that in which stratification is set radiatively.
Warming as function of latitude and height

1. Upper stratospheric cooling.
2. Increase height of tropopause.
3. Surface polar amplification.
4. Extra warming aloft in tropics.

Moist adiabatic lapse rate (critical lapse rate for convection for saturated air)

\[
-\frac{dT}{dz} \approx \frac{g}{c_p} \frac{1 + L_c q_s/(RT)}{1 + L_c^2 q_s/(c_p R T^2)}.
\]
Warming as function of latitude and height

1. Upper stratospheric cooling.
2. Increase height of tropopause
3. Surface polar amplification.
4. Extra warming aloft in tropics

Moist adiabatic lapse rate (critical lapse rate for convection for saturated air)

\[- \frac{dT}{dz} \approx \frac{g}{c_p} \frac{1 + L_c q_s/(RT)}{1 + L_c^2 q_s/(c_p RT^2)}.\]

Lapse rate and its rate of change with temperature:
Model (CMIP5) predicted Temperature and tropopause height changes

Increase in tropopause height is common across models. The change in height is greater than the model standard deviation, especially in low latitudes.
Tropopause Height

→ Incoming solar radiation = outgoing IR
→ Stratosphere in radiative equilibrium
→ Uniform tropospheric stratification
→ Outgoing IR radiation can be written as a function of tropopause temperature only.

Only one choice of \( H(T) \) gives the correct OLR.
→ Incoming solar radiation = outgoing IR
→ Stratosphere in radiative equilibrium
→ Uniform tropospheric stratification
→ Outgoing IR radiation can be written as a function of tropopause temperature only.

Only one choice of $H(T)$ gives the correct OLR.

Tropopause height increases with increased COT.
Even as we add greenhouse gases, the OLR is *fixed* independently of optical depth.

**In the troposphere:**

\[ T = T_S - \Gamma z, \quad z \leq H_T \]

**Change in trop. height:**

\[ \Delta H_T = \frac{\Delta T}{\Gamma} \]
Tropopause Height

with Gray Radiation and a `Thin' Stratosphere

\[ \frac{\partial U}{\partial \tau} = U - B \]
\[ \frac{\partial D}{\partial \tau} = B - D, \]

where \( \tau = \tau(z) \), \( U \) is upwards irradiance, \( D \) is downwards irradiance and \( B = \sigma T^4 \).

\[ \frac{\partial}{\partial \tau} (U - D) = U + D - 2B, \quad \frac{\partial}{\partial \tau} (U + D) = U - D \]

Stratosphere in longwave radiative equilibrium:

\[ D = \frac{\tau}{2} OLR, \quad U = \left(1 + \frac{\tau}{2}\right) OLR, \quad B = \frac{1 + \tau}{2} OLR. \]

and if \( \tau \ll 1 \)

\[ D = 0, \quad U = OLR = 2B, \quad B = OLR/2. \]

Stratosphere is isothermal. Tropopause temperature fixed by OLR.
Lapse rate and temperature effects

\[ \Delta H_T = \frac{\Delta T}{\Gamma} - \frac{H_T \Delta \Gamma}{\Gamma} \]

\( T_T \) is the tropopause temperature, \( \Delta T \) is the increase in temperature at a given height in the troposphere, \( \Delta \Gamma \) the change in the lapse rate.

That is:

\[ \frac{\partial H_T}{\partial T} = \frac{1}{\Gamma} - \frac{H_T}{\Gamma} \frac{\partial \Gamma}{\partial T} \]

Change in tropopause height with change in temperature and lapse rate.
Theoretical predictions of tropopause height

Increase in tropopause height with change in lapse rate and with temperature.

\[ \Delta H_T = \frac{\Delta T}{\Gamma} - \frac{H_T \Delta \Gamma}{\Gamma} \]

\[ \frac{\partial H_T}{\partial T} = \frac{1}{\Gamma} - \frac{H_T \partial \Gamma}{\Gamma \partial T} \]

`Temperature effect' = \( \frac{1}{\Gamma} \)

`Lapse rate effect' = \(- \frac{H_T \partial \Gamma}{\Gamma \partial T} \)
Change in tropopause height is correlated to the climate sensitivity, both locally and in the mean.
Analytic Expression for Tropopause Height

\[-\frac{dU}{d\tau} = B - U, \quad \frac{dD}{d\tau} = B - D.\]

Suppose that lapse rate, $\Gamma$, is given up to a height $H_T$, above which the atmosphere is in radiative equilibrium.

Formal solution:

\[
D(\tau') = e^{-\tau'} \left[ D(0) - \int_0^{\tau'} B(\tau)e^{\tau} \, d\tau \right], \quad U(0) = U(\tau')e^{-\tau'} + \int_0^{\tau'} B(\tau)e^{-\tau} \, d\tau
\]

But, these expressions don't give the right answer for outgoing radiation for an arbitrary $\tau'$.

Must adjust $H_T$ so that the equations satisfy the boundary conditions.

\[
H_T = \frac{1}{16\Gamma} \left( C T_T + \sqrt{C^2 T_T^2 + 32\Gamma \tau_s H_a T_T} \right).
\]

where $H_a$ is the scale height of the main absorber and $C = \log 2$. 

Numerical and Analytic Comparison

Numerical solution

Analytical approximation
Stratospheric Cooling

Also robust, but dependent on the presence of ozone!

Basic Mechanism:

- A balance between solar heating and longwave cooling.
- If emissivity (optical depth) increases, a lower temperature will suffice to provide the needed cooling.

Algebra:

In a semi-grey atmosphere one may show...

\[
3\sigma T^3 \frac{\partial T}{\partial z} \approx - \frac{\tau}{2H_a} I + \frac{1}{2\tau} Q
\]

- The robustness of the effect does not depend on the detailed distribution of absorption bands of carbon dioxide or water vapor.
- Effect depends on there being (ozone) heating.
Assume $\tau(z) = \tau_0 \exp[-z/H]$ and that the stratosphere is in radiative equilibrium, with long-wave cooling balancing short wave heating, $Q$.

After manipulation we find:

$$\frac{\partial B}{\partial z} \approx -\frac{\tau}{2H} I + \frac{1}{2\tau} Q$$

If $\tau$ increases the heating term diminishes, and the temperature increase with height falls.

Before: ---

After: ---
The Overturning Circulation: Hadley and Ferrel Cells
Assume flow is axi-symmetric. Outflow is angular momentum conserving:

$$U = \Omega a \frac{\sin^2 \vartheta}{\cos \vartheta}$$

Temperature from thermal wind balance:

$$T = T(0) - \frac{T_0 \Omega^2 \vartheta^4}{2 g H a^2}$$

Temperature falls rapidly with latitude.

Width of Hadley Cell is constrained by thermodynamics:
Air gets too cold and sinks.
1. Held--Hou theory (axi-symmetric)

\[ \phi_H = \left( \frac{5 \Delta \theta_h g H}{3 a^2 \Omega^2 \theta_0} \right)^{1/2} \propto H^{1/2}. \]

- Increase in height of tropopause leads to Hadley Cell expansion.

\[ \frac{\Delta \phi_H}{\phi_0} = \frac{\Delta H}{2H_0} = \frac{\Delta T}{2H_0 \partial T/\partial z} \sim \frac{1}{50} \text{ per } ^\circ \text{C} \]

- Assumes other factors stay the same.
- The atmosphere is not Boussinesq.

2. But baroclinic eddies are likely important. Hadley Cell extends until it feels the effect of baroclinic instabilities.
Hadley Cell Width: with and without eddies

(a) Zonally symmetric simulation

(b) 3D simulation. Hadley cell is narrower and stronger.

Courtesy C. Walker (c.f., Walker & Schneider).
Critical shear in two-layer model:

$$U_s = \frac{1}{4} \beta L^2 \quad \text{where} \quad L^2 = \frac{NH}{f}$$

gives critical latitude:

$$\phi_c \approx \left( \frac{N^2 H^2}{\Omega^2 a^2} \right)^{1/4} = \left( \frac{g \Delta \theta H}{\theta_0 \Omega^2 a^2} \right)^{1/4}$$

Dependence on tropopause height and stratification.

**Hadley Cell expands if:**

1. Tropopause height increases.
2. Stratification increases (which stabilizes the flow)

**But note**

The atmosphere is not a two-level model!
Other formulations are possible, but quantitative predictions will necessarily be uncertain.
The Hadley Cell terminates where Rossby wave breaking occurs, not at the latitude of baroclinic instability.

The real Hadley Cell is probably a combination of the above mechanisms. Different GCMs may have different combinations, and different scalings.
• Hadley cell expands in most models
• Significant scatter.
• Southern expansion is weakly correlated with Northern expansion
• Expansion is not correlated with degree of warming of a model.
The Mid-latitudes: A still harder problem?

• The atmospheric mid-latitude circulation is a problem in weak turbulence (eddy--mean-flow interaction) and so a difficult problem.
• A small shift in the surface winds could have large effects on the climate in mid-latitudes.
• Surface winds approximately obey the eddy--mean-flow balance, in QG approximation and in the steady state,

$$r \bar{u}_s \approx \int \frac{\partial u'v'}{\partial y} \, dz = \int v'q' \, dz$$

where $r$ is a surface friction parameter.
Changes in Surface Winds

Ensemble mean surface wind

Changes in wind of individual models

Small and inconsistent differences in general, larger in Southern Hemisphere
Shift of the surface westerlies

**Summer**

- **1pctCO2**
  - SH Summer
  - NH Summer

**Winter**

- **1pctCO2**
  - SH Winter
  - NH Winter
Shift of the surface westerlies
vs increase in temperature

Lots of scatter!
Shift of Hadley Cell and the Mid-latitude Westerlies

Are they correlated?

Hadley Cell expansion vs shift of the westerlies, using an overturning measure

![Graph showing correlation between Hadley Cell expansion and westerlies shift.](image)
Shift of Hadley Cell and the Mid-latitude Westerlies

Are they correlated?

Hadley Cell expansion vs shift of the westerlies, using a surface wind measure.
Surface Wind Changes
As a function of Current position

Scatter plot of latitude of surface westerlies vs shift in the future.

- Control SH Westerlies (deg.) vs Shift of SH Westerlies (deg.)
  - 1pctCO2
  - Summer (DJF); R = 0.002
  - Winter (JJA); R = −0.59

- Control NH Westerlies (deg.) vs Shift of NH Westerlies (deg.)
  - 1pctCO2
  - Summer (JJA); R = −0.12
  - Winter (DJF); R = −0.11
Dependent on season, hemisphere and the model itself!

Far more scatter in these results than in the warming itself.
Factors influencing jets

A stratospheric influence?

- Brewer-Dobson circulation
- Polar Vortex
- Eddy-driven jet
- Subtropical jet
- Tropospheric residual circulation
- Hadley Cell
- Surface westerlies
- Stratosphere
- Troposphere
Conclusions

• Thermodynamic/radiative changes to atmosphere are robust:
  • Increase in height of the tropopause, and cooling of the stratosphere, have solid physical mechanisms and are reproduced by comprehensive models.

• Dynamical or circulation changes are less well understood.
  • Hadley cell expansion is a common feature, and the poleward shift of westerlies is also common, but scatter is very large.
  • Many proposed mechanisms (acting alone, not all can be correct).
Conclusions

• Thermodynamic/radiative changes to atmosphere are robust:
  • Increase in height of the tropopause, and cooling of the stratosphere, have solid physical mechanisms and are reproduced by comprehensive models.

• Dynamical or circulation changes are less well understood.
  • Hadley cell expansion is a common feature, and the poleward shift of westerlies is also common, but scatter is very large.
  • Many proposed mechanisms (acting alone, not all can be correct).

• To what extent are dynamical changes predictable or knowable?
• Depends on interaction with subgridscale parameterizations (e.g., convection).
Conclusions

• Thermodynamic/radiative changes to atmosphere are robust:
  • Increase in height of the tropopause, and cooling of the stratosphere, have solid physical mechanisms and are reproduced by comprehensive models.

• Dynamical or circulation changes are less well understood.
  • Hadley cell expansion is a common feature, and the poleward shift of westerlies is also common, but scatter is very large.
  • Many proposed mechanisms (acting alone, not all can be correct).

• To what extent are dynamical changes are predictable or knowable?
• Depends on interaction with subgrid-scale parameterizations (e.g., convection).

• Entering a Golden Age for dynamicists! --- or at least we should be.