Water Vapor and the Dynamics of Climate Changes
Water vapor dynamics in warming climate

**Facts**

- Saturation vapor pressure increases with temperature at \( \sim 7\%/K \)
- Relative humidity *near the surface* stays roughly constant
- Precipitable water \( q \) increases at \( \sim 7\%/K \) with sfc. temperature
- Precipitation \( P \) increases more slowly, at \( \sim 2–3 \%/K \)
- Water vapor cycling rate \( P/q \) decreases

**Common conjectures**

- Tropical circulations (particularly Walker circulation) slow down
- Hadley circulation widens
- Extratropical storms become more energetic
- Precipitation extremes increase more rapidly than mean
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Water vapor dynamics in warming climate

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- Relative humidity *near the surface* stays roughly constant
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Simulations with idealized moist GCM

- Aquaplanet: uniform, water-covered surface; no ocean dynamics
- Built on GFDL FMS [similar to Frierson et al. (2006)]
- Only vapor-liquid phase transition considered (no ice)
- Radiative transfer of semi-gray atmosphere
- Climate varied by varying “greenhouse gas concentrations”: scaling of optical thickness of longwave absorber (by factor 12)

Allows very large climate variations: Global-mean surface temperatures between 259 K and 316 K (!)
A wide range of climates...

(O’Gorman & Schneider 2008)
Saturation vapor pressure

Clausius-Clapeyron relation between temperature $T$ and saturation vapor pressure $e_s$

$$\frac{\delta e_s}{e_s} = \frac{L}{RT^2} \delta T$$

For warming Earth, this implies increase in saturation vapor pressure of 7%/K, or 21% for 3K warming.

(Boer 1993; Wentz & Schabel 2000; Trenberth 2003)
Near-surface relative humidity

Evaporation (over ocean) is proportional to

\[ E \sim e_s - e = e_s(1 - RH) \]

so \( \delta E \sim \delta e_s(1 - RH) - e_s \delta (RH) \) or

\[ \delta (RH) = (1 - RH) \left( \frac{\delta e_s}{e_s} - \frac{\delta E}{E} \right) \]

RH~85% near surface, \( E \approx 80 \text{ W/m}^2 \), climate sensitivity \(~0.8 \text{ K/(W m}^2\text{)}\), so \( \delta E/E \sim O(1%/\text{K}) \Rightarrow \delta (RH) \sim O(1%/\text{K}) \)

Near-surface RH changes strongly energetically constrained

(Boer 1993; Held & Soden 2000, Schneider et al. 2010)
Near-surface relative humidity

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Near-surface RH changes strongly energetically constrained

(Boer 1993; Held & Soden 2000)
Precipitable water in idealized GCM

Earth-like (6.2%/K)

Most water vapor near surface
(Free-tropospheric RH does change)

RH=0.67

(O’Gorman & Schneider 2008)
Precipitable water in comprehensive GCMs

Global–Means (year 61–80 @ 1% CO₂ per year)

Coupled Model Intercomparison Project

CMIP II Models

(Held & Soden 2006; figure courtesy I. Held)
Precipitation in idealized GCM

\[ \langle P \rangle = \langle (1 - \alpha) S_{sfc} \rangle / L \]

Earth-like (2.5%/K)

Asymptotes to energetic bound

(O’Gorman & Schneider 2008)
Precipitation in comprehensive GCMs

Global-Means (year 61–80 @ 1% CO₂ per year)
Coupled Model Intercomparison Project

Decrease in cycling rate

(Held & Soden 2006; figure courtesy I. Held)
Water vapor cycling rate

Generally decreases (except in cold climates)

(O’Gorman & Schneider 2008; Schneider et al. 2010)
Tropical convective mass flux

Moisture (or thermodynamic) balance in saturated updrafts

\[-\omega^{\uparrow} \partial_p q^* \approx c,\]

where

\[\omega^{\uparrow} = \begin{cases} 
\omega & \text{if } \omega < 0 \\
0 & \text{if } \omega \geq 0 
\end{cases}\]

Mass-weighted vertical integral \{ \cdot \}

\[-\{\omega^{\uparrow} \partial_p q^*\} \approx P,\]

(avg’d over convective system)

(cf. Iribarne & Godson 1981; Schneider et al. 2010)
Scaling of convective mass flux

General scaling behavior

\[-\frac{\omega^\dagger}{g} \sim \frac{P}{\Delta q^*}, \quad \Delta q^* = q_s^* - q^*\]

**Case A**

\[\Delta q^* \sim \partial_p q^*|_{\theta_e^*} \Delta p \sim S^* \Delta p\]

*Mass flux scales with inverse static stability* (Betts & Harshvardhan 1987)

**Case B**

\[\Delta q^* \sim q_s^* - q^* \sim q_s^*\]

*Mass flux scales with cycling rate* (Betts 1998; Held & Soden 2006)
Scaling estimates are very different...

At 290 K, $\Delta q$ increases at 2.0%/K, $q$ at 6.4%/K

With $\delta P/P \sim 2.5%/K$, mass flux increases under A, decreases under B!

(Schneider et al. 2010)
Convective mass flux scales inversely with static stability, not with cycling rate; non-monotonic function of surface temperature
Changes in comprehensive GCMs

Fig. 1. Time series of the global-mean response of (a) surface air temperatures, (b) total column water vapor, (c) precipitation, and (d) upward component of the monthly mean vertical velocity at 500 hPa. The latter three quantities are expressed as the fractional change relative to the first 10 yr of the A1B integration (2000–10). All fields are computed by subtracting the mean seasonal cycle from 2000–10, global averaging, and then differencing the anomalies relative to their average computed over the first 10 yr of the simulation (2000–10).

(Vecchi and Soden 2007)
Changes in convective mass flux

Weakening of Walker circulation

(Vecchi and Soden 2007)
Weakening of Walker circulation

Fig. 7. The multimodel (a) ensemble-mean 500hPa $\omega$ (hPa/day) and (b) ensemble-mean change in 500hPa $\omega$ per degree global warming; positive values are downward. The change is computed by differencing the decadal-mean 500hPa $\omega$ from the first and last 10 yr of the twenty-first century, normalizing this difference by the change in global-mean temperature for each model, and then averaging the result across all models. Units: (top) hPa day$^{-1}$ and (bottom) hPa day$^{-1}$K$^{-1}$. (Vecchi and Soden 2007)
Precipitation extremes scale similarly...

Based on IPCC 21st-century global warming simulations

(O’Gorman and Schneider 2009)
So the convective (gross) upward mass flux (zonally asymmetric) is thermally driven and depends on moist-adiabatic static stability. It may increase or decrease as the climate warms.

What does that imply about the net upward mass flux (Hadley circulation)?
Hadley cell strength in idealized GCM

(Levine and Schneider 2011)
Non-monotonic function of surface temperature

(Schneider et al. 2010)
Eddies mediate Hadley circulation response

Eddy momentum flux scales with EKE, which is maximal near reference climate and scales with MAPE

(O’Gorman & Schneider 2008; Schneider & Walker 2008)
Eddies mediate Hadley circulation response

Non-monotonic function of surface temperature (e.g., LGM less stormy?)

(O’Gorman & Schneider 2008; Schneider & Walker 2008)
Why is eddy scaling non-monotonic?

- MAPE
- Temperature gradient at surface
- Static stability
- Tropopause height

(O’Gorman & Schneider 2008)
Poleward energy flux

(Schneider et al. 2010)
Why is eddy scaling non-monotonic?

(O’Gorman & Schneider 2008)
Hadley cell width

Increases with surface temperature; eddies control it; static stability in subtropics influences it
Storm tracks (near-surface EKE)

Storm tracks at local MAPE maximum (Mbengue’s thesis 2015)
Conclusions

• Precipitable water increases rapidly with temperature, precipitation less rapidly, water vapor cycling rate generally decreases, but...

• Gross upward mass flux in tropics may depend on static stability (increases slowly with temperature).

• Hadley cell during equinox, summer, and in annual mean controlled by eddy fluxes.

• Eddy scaling (non-monotonic) imprinted on Hadley cell response to climate change.

• Hadley cell and extratropical storms weaker in warmer and in (much) colder climates.

• Changes in Hadley cell width likewise eddy-controlled, with implications for subtropical dryness.
The multimodel (a) ensemble-mean and (b) ensemble-mean change in 500 hPa mass flux per degree global warming; positive values are downward. The change is computed by differencing the decadal-mean 500 hPa mass flux from the first and last 10 yr of the twenty-first century, normalizing this difference by the change in global-mean temperature for each model, and then averaging the result across all models. Units: (top) hPa day$^{-1}$ and (bottom) hPa day$^{-1}$ K$^{-1}$.