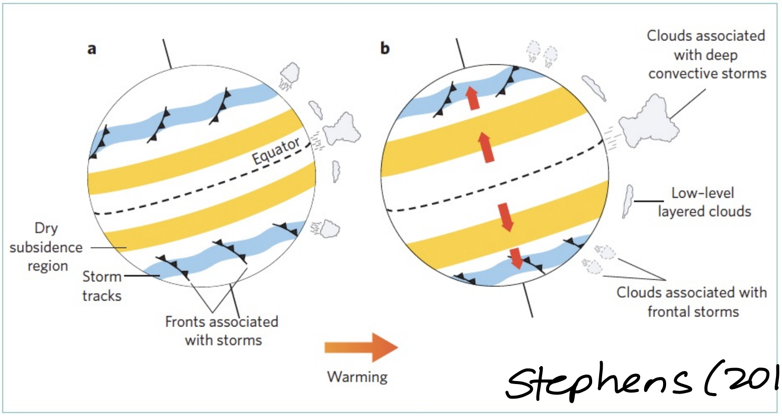
Boulder School 2022 Tiffany Shaw

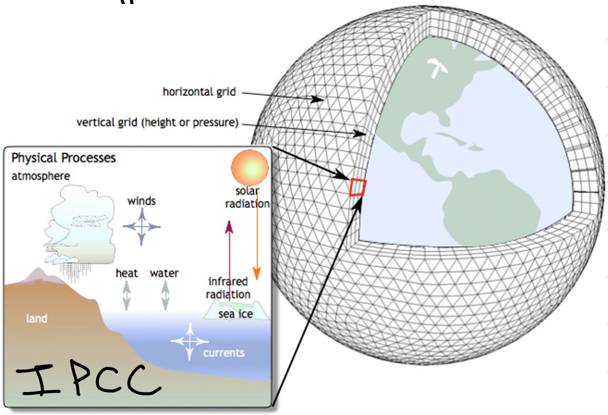
Lecture 3 Large scale circulation

Response of the large scale circulation to climate change (global warming due to increased CO_2 concentration)



Stephens (2011)

Everything we know about future climate change comes from climate models



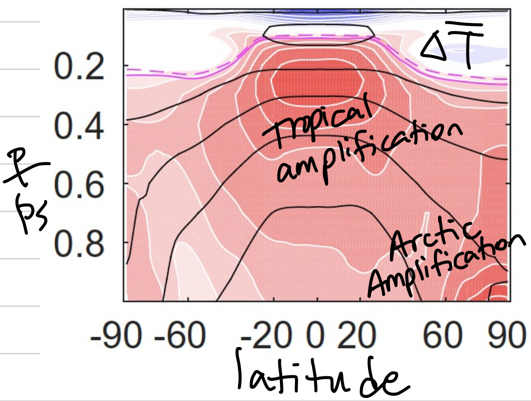
IPCC

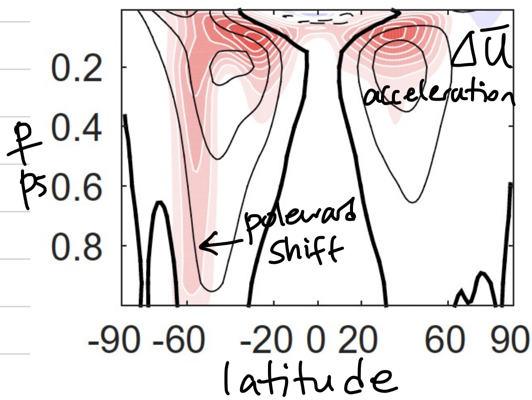
Hydrostatic primitive equations + closures for unresolved physics (parameterizations)

Force with different emission scenarios from IPCC

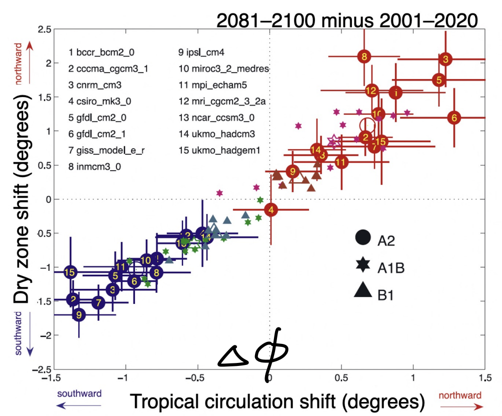
Climate model projections under high emissions scenario (business as usual)

Δ = future − present

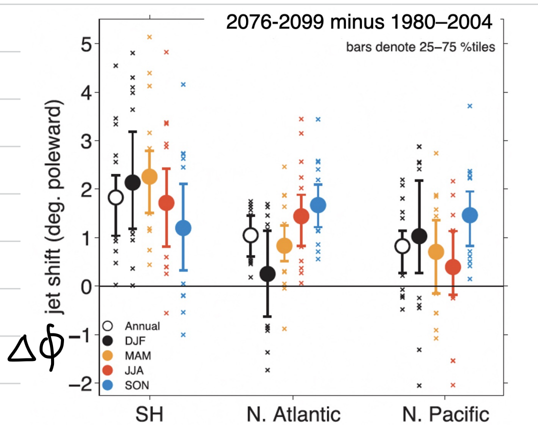




Shaw (2019)



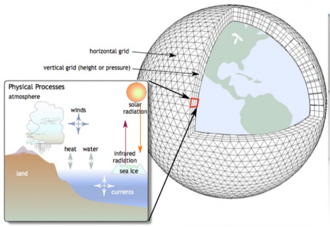
poleward shift Hadley cell edge
Lu et al. (2007)

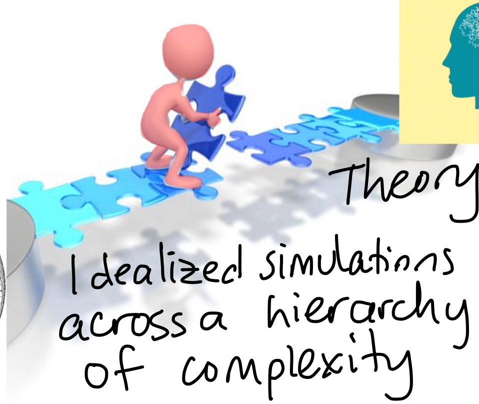


poleward shift of surface eastward flow

Barnes & Polvani (2013)

Closing the gap between simulating and understanding climate change

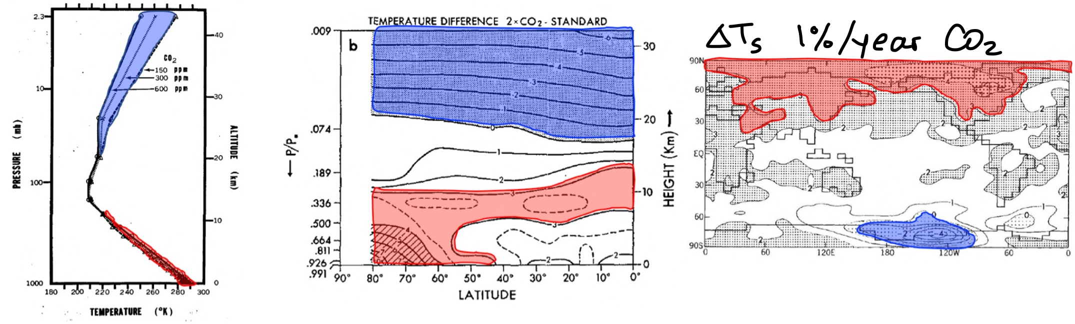




Theory

Idealized simulations across a hierarchy of complexity

Manabe's work illustrates how a climate model hierarchy can be used to simulate and understand climate change



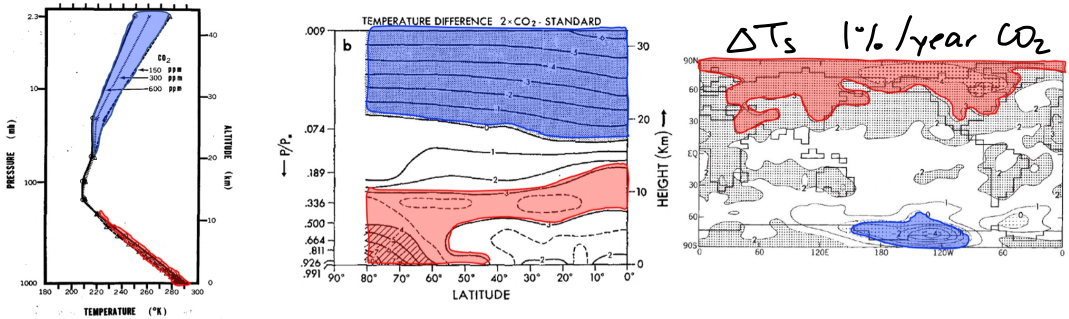
Manabe & Weatherald (1967) | Manabe & Weatherald (1975) | Stouffer, Manabe & Weatherald (1989)

→ Physical Complexity

1967: Radiative convective equilibrium model (convection parametrized) predicts cooling in the stratosphere, warming in the troposphere, surface warming depends on H_2O and clouds

Change of CO_2 content (ppm)	Fixed relative humidity		Fixed absolute humidity	
	Average cloudiness	Clear	Average cloudiness	Clear
300 → 600	+2.36	2.92	+1.33	+1.36

1975: First general circulation model simulates solution to equations of motion to $2xCO_2$ with land and ocean surface, includes snow and ice (T dependent surface albedo). Still get cooling but new signal emerges: amplified polar warming



Manabe & Weatherald (1967) Manabe & Weatherald (1975) Stouffer, Manabe & Weatherald (1989)

Physical Complexity

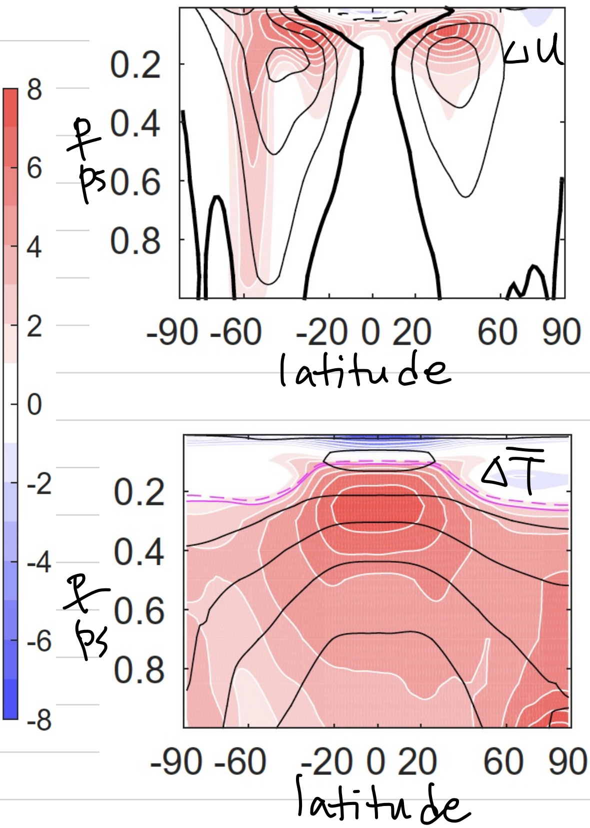
1989: Coupled (atmosphere + ocean) GCM still get amplified Ts over Arctic. Ocean circulation leads to southern ocean cooling (delayed warming due to passive advection of T)

$$\Delta\left(v \frac{\partial T}{\partial y}\right) \approx \Delta v \frac{\partial T}{\partial y} + \boxed{v \frac{\partial \Delta T}{\partial y}}$$

All of these predictions of the temperature response to climate change have emerged in the observations!!

Examples of simulating and understanding large scale circulation response to climate change

Example 1 : Acceleration of subtropical jet



Climate model projection under high emissions scenario

$$\Delta u \approx +6 \ \frac{m}{s}$$

Related to the temperature response

Recall thermal wind balance from Lecture 1

$$f \frac{\partial u}{\partial p} = \frac{R}{p} \frac{1}{a} \frac{\partial T}{\partial \phi}$$

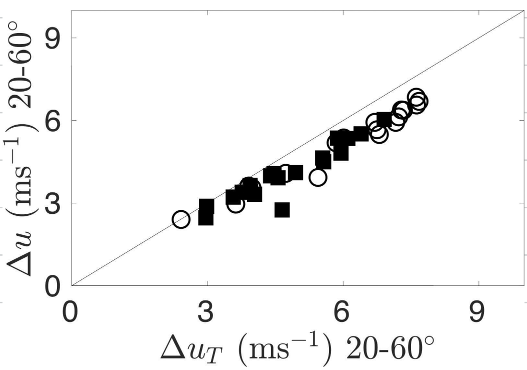
Let Δ = future − present
use differential calculus rules

$$\Rightarrow f \frac{\partial \Delta u}{\partial p} = \frac{R}{p} \frac{1}{a} \frac{\partial \Delta T}{\partial \phi}$$

Explain $\Delta T \xrightarrow{\text{thermal wind}}$ explain Δu

$$\Delta u_T \approx \int_1^{0.2} \frac{R}{fp} \frac{1}{a} \frac{\partial \Delta T}{\partial \phi} \, dp > 0$$

Response of Δu vs Δu_T in state of the art climate models



■ = Northern Hemisphere

○ = Southern Hemisphere

Why is there more warming aloft?

Temperature response consistent with moist adiabatic adjustment

$$\Gamma_m = \Gamma_d \left[\frac{1 + \frac{L q^*(T,p)}{R_d T}}{1 + \frac{L^2 q^*(T,p)}{c_p R_v T^2}} \right]$$

$$\Delta \Gamma_m = \Gamma_d \left[- \frac{\frac{L^2 \Delta q^*}{c_p R_v T^2} \left(1 + \frac{L q^*}{R_d T}\right)}{\left(1 + \frac{L^2 q^*}{c_p R_v T^2}\right)^2} + \dots \right]$$

$$\Delta q^* = \alpha \Delta T q^* > 0$$

$$\Rightarrow \Delta \Gamma_m < 0$$

Explains $\Delta\Gamma$ see Miyawaki et al (2020)

Amplified warming aloft in low latitudes due to increase latent heat released aloft following moist adiabatic adjustment

Opposed by Arctic amplification of surface warming

Example 2: Poleward shift of Hadley cell extent

Recall Lecture 2 $\quad 2\Omega^2 a^2 \phi^3 \geq \frac{\partial s^*}{\partial \phi}(T_T - T_s)$

$$\Rightarrow \quad 6\Omega^2 a^2 \phi^2 \Delta\phi \geq \frac{\partial \Delta s^*}{\partial \phi}(T_T - T_s) + \frac{\partial s^*}{\partial \phi}\Delta(T_T - T_s)$$

$$\Rightarrow \quad \Delta\phi \geq \frac{\frac{\partial \Delta s^*}{\partial \phi}(T_T - T_s) + \frac{\partial s^*}{\partial \phi}\Delta(T_T - T_s)}{6\Omega^2 a^2 \phi^2}$$

$$\frac{\Delta\phi}{\phi} \geq \frac{\frac{\partial \Delta s^*}{\partial \phi}(T_T - T_s) + \frac{\partial s^*}{\partial \phi}\Delta(T_T - T_s)}{6\Omega^2 a^2 \phi^3}$$

$$\frac{\Delta\phi}{\phi} \geq \frac{1}{3}\,\frac{\frac{\partial \Delta s^*}{\partial \phi}(T_T - T_s) + \frac{\partial s^*}{\partial \phi}\Delta(T_T - T_s)}{\partial s^*/\partial \phi\,(T_T - T_s)}$$

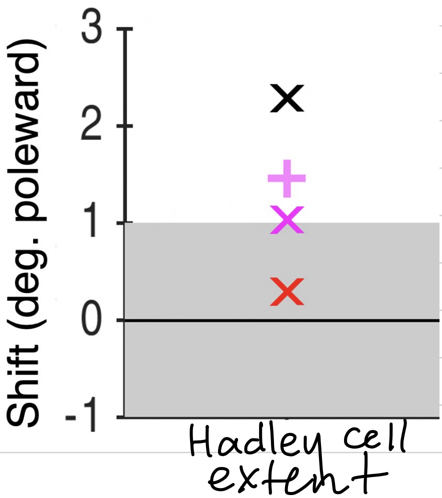
$$\frac{\Delta\phi}{\phi} \gtrsim \frac{1}{3}\left[\underbrace{\frac{\partial \Delta s^*/\partial\phi}{\partial s^*/\partial\phi}}_{\substack{\text{Shaw \& Voigt (2016)} \\ \approx \alpha \Delta T_s \\ \approx +28\%}} + \underbrace{\frac{\Delta(T_T - T_s)}{T_T - T_s}}_{\substack{\frac{+10K}{-100K} \\ \approx -10\%}}\right]$$

opposing effects

$$\frac{\Delta\phi}{\phi} \gtrsim 18\%$$ poleward shift

Tug of war between impacts of moisture response on vertical temperature structure and latitudinal structure
s^*

Test moisture mechanism in idealized model (aquaplanet) that captures the signal and where we can control moisture through surface boundary condition (turbulent flux bulk aerodynamic formula parameterization)



Full response to $4xCO_2$

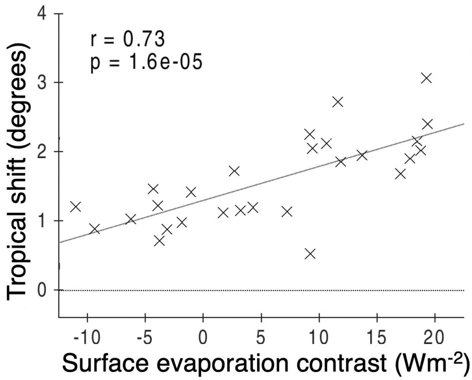
Dry atmosphere response

Surface humidity induced evaporation $\frac{\partial \Delta s^*}{\partial y} < 0$

surface humidity induced by global mean ΔT

Tan and Shaw (2020)

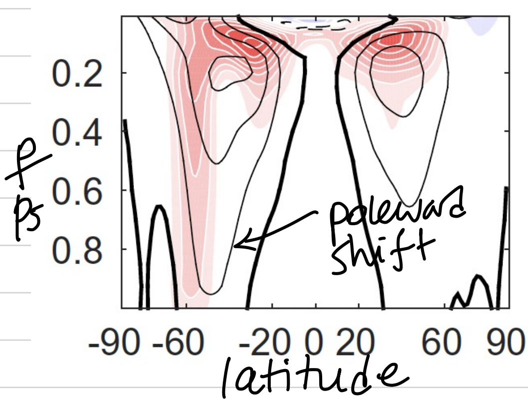
Significant connection also exists across State of the art IPCC models



Each x is a different model

Tan & Shaw (2020)

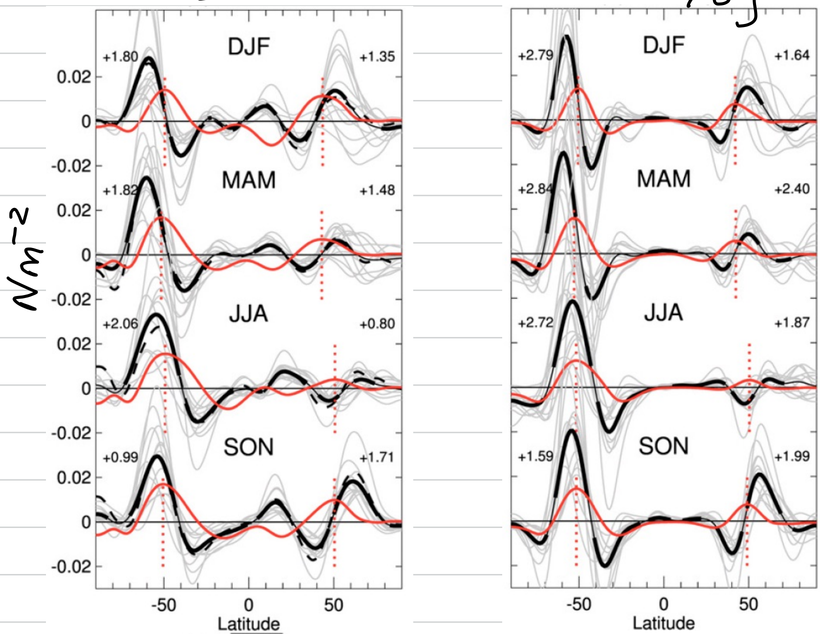
Example 3: Poleward shift of surface eastward flow



Annual mean

Simpson, Shaw, Seager (2014) showed poleward shift dominated by eddy momentum flux

$\Delta\tau$ $\qquad\qquad$ $-\partial\Delta\overline{u'v'}/\partial y$



Recall extratropical regime Lecture 2:

$$r\,\bar{u}_{surf} = -\frac{\partial}{\partial y}\left(\overline{u'v'}\right)$$

$$r\,\Delta\bar{u}_{surf} = -\frac{\partial}{\partial y}\left(\Delta\overline{u'v'}\right)$$

Recall from Lecture 2

$$\overline{u'v'} = -\frac{1}{2} A^2 k l$$

$$\Delta\left(\overline{u'v'}\right) \approx -Akl\Delta A - \frac{1}{2} A^2 (k\Delta l + l\Delta k)$$

Many different possibilities

Option 1: Δk

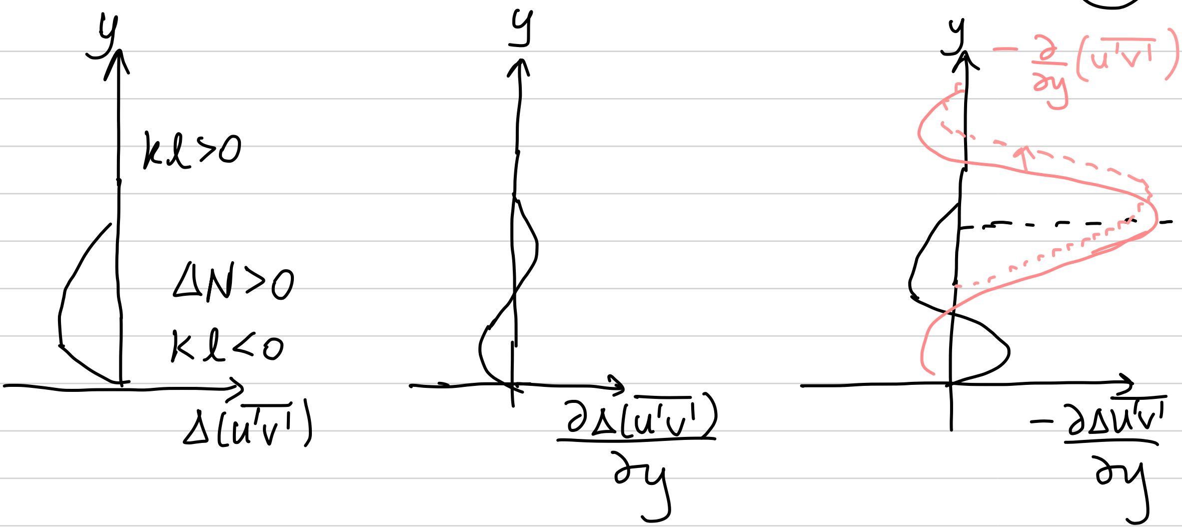
$$k = \frac{2\pi}{L} \qquad L = L_d = \frac{NH}{f}$$

$$\Delta L = \frac{H\Delta N}{f} + \frac{N\Delta H}{f} > 0$$

$$\Rightarrow \Delta k = -\frac{2\pi}{L^2} \Delta L < 0$$

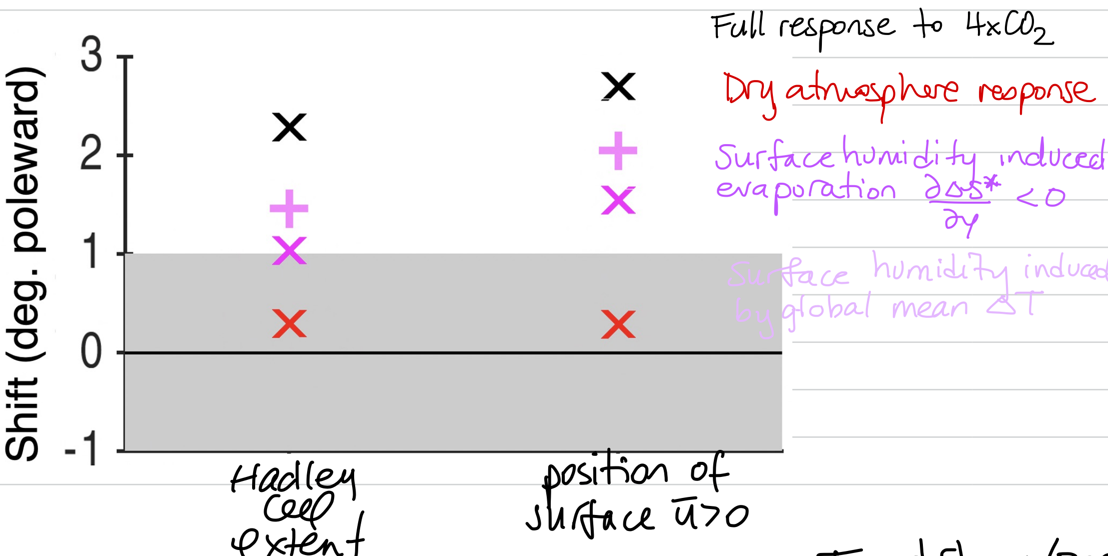
$$\Rightarrow \Delta\left(\overline{u'v'}\right) \approx -\frac{1}{2} A^2 l \Delta k$$

$$< 0$$



Evidence from aquaplanet simulations

$\Delta N > 0$ due to moist adiabatic adjustment



Tan & Shaw (2020)

Exercise using

$$\Delta\left(\overline{u'v'}\right) \approx -Ak\ell\Delta A - \frac{1}{2}A^2\left(k\Delta\ell + \ell\Delta k\right)$$

explore other possibilities including how $\Delta\bar{u}$ impacts $\Delta\ell$.

See Shaw (2019)