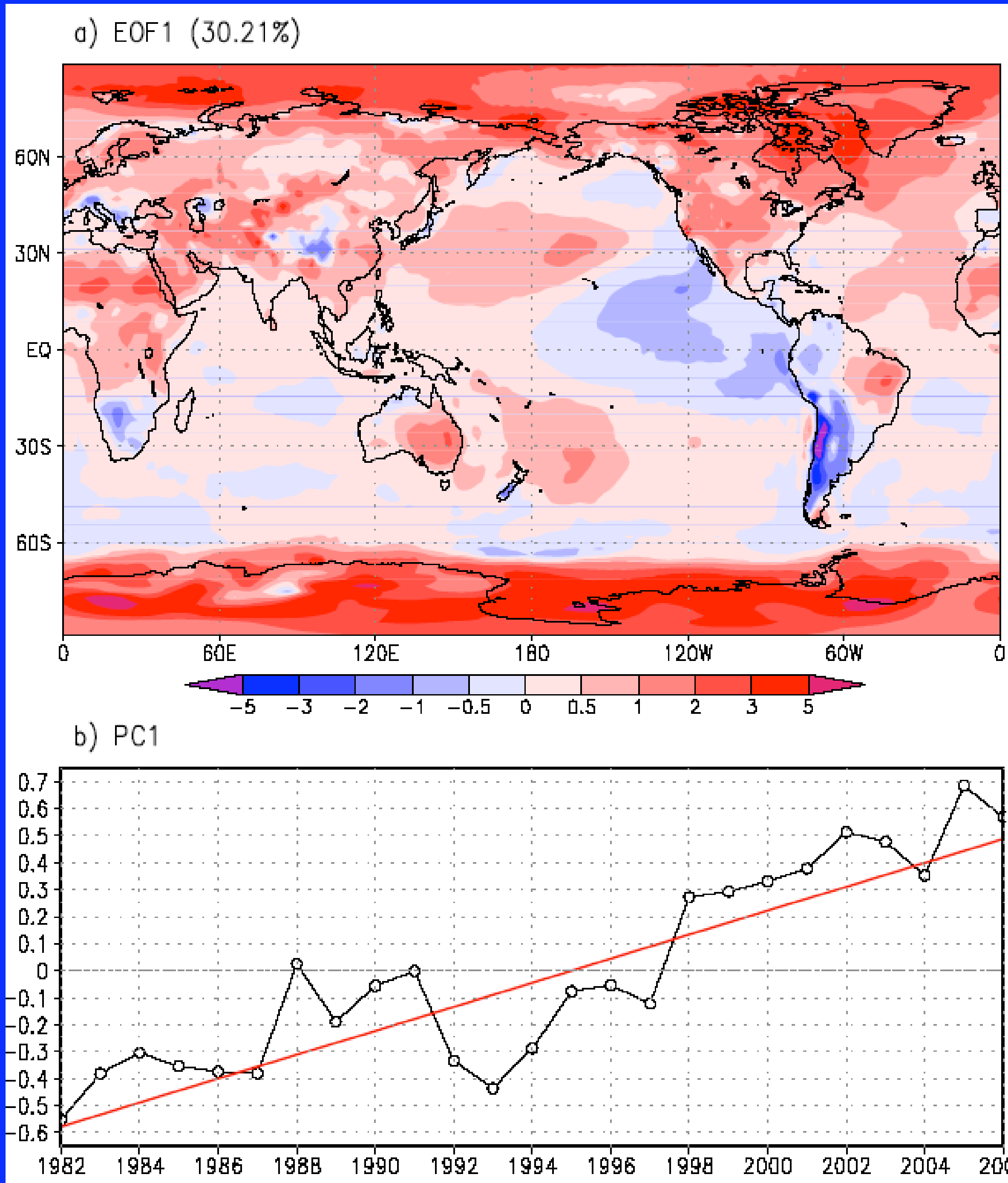
Dynamical Polar Warming Amplification and a New Climate Feedback Analysis Framework

Ming Cai
Florida State University
Tallahassee, FL 32306
cai@met.fsu.edu

Cai (2005, GRL); Cai (2006, Climate Dynamics)
Cai and Lu (2007, Climate Dynamics)
Lu and Cai, and Cai and Lu (2008, Climate Dynamics, in revision)

Observed global warming pattern (NCEP R2)



1st EOF (30% var.)

0.22 K/decade

Questions

- **Science: What are the roles of atmospheric motions (turbulences, convections, large-scale motions) for the spatial (vertical and horizontal) variations of the warming pattern? Specifically, can the atmospheric dynamics alone explain a larger warming in high latitudes?**

- **Technique: How do we incorporate atmospheric dynamics in the climate feedback analysis?**

Outline

- **Brief review on the TOA-based feedback analysis method (PRP method).**
- **Prototype approach in a theoretical model.**
- **Formulation of a new framework (CFRAM)**
- **Demonstration of the CFRAM and comparison with the PRP method.**
- **Application of CFRAM to understand the polar warming amplification in a GCM without hydrological cycle.**
- **Summary**

General definition of feedback

- **Forcing: an energy input to the system**
- **Response: an output of the system**
- **A feedback: an "induced input from the output"**

A brief overview of the Partial Radiative Perturbation (PRP) method

(designed for a globally uniform SURFACE warming)

Partial Radiative Perturbation Method

- **Forcing: a radiative flux perturbation at the TOA**
- **Response: surface temperature (or system temperature)**
- **Feedback: additional radiative flux perturbations at the TOA in response to surface temperature**

$$\Delta F^{ext} = -(\Delta S_{TOA} - \Delta OLR_{TOA}) = -\frac{d(S_{TOA} - OLR_{TOA})}{dT_S}\Delta T_S$$

$$\lambda_{tot} = \frac{d(S_{TOA} - OLR_{TOA})}{dT_S}$$

$$\Delta T_S = \frac{-F^{TOA}}{\lambda_{tot}} = G_{tot} F^{TOA}$$

$\lambda_{tot} < 0$: (Total) Feedback parameter

The warmer surface temperature is, the more energy outputs from the climate system

$G_{tot} = (-\lambda_{tot})^{-1}$: (Total) Gain of the climate system

Partial Radiative Perturbation Method

$$\lambda_{tot} = -\frac{\partial R_{TOA}}{\partial T_S} + \frac{\partial (S_{TOA} - R_{TOA})}{\partial H_2O}\frac{d(H_2O)}{dT_S} + \frac{\partial (S_{TOA} - R_{TOA})}{\partial \alpha}\frac{d\alpha}{dT_S}$$

$$+ \frac{\partial (S_{TOA} - R_{TOA})}{\partial cloud}\frac{d(cloud)}{dT_S} + \frac{\partial (S_{TOA} - R_{TOA})}{\partial T_{air}}\frac{dT_{air}}{dT_S}$$

$$= \lambda_p + \lambda_{H_2O} + \lambda_{albedo} + \lambda_{cloud} + \lambda_{lapse\_rate}$$

$$= -\lambda_p (1 - g_{H_2O} - g_{albedo} - g_{cloud} - g_{lapse\_rate})$$

$$\Delta T_S = \frac{G_0 F^{TOA}}{1 - \sum_x g_x}$$

$G_0 = -1/\lambda_P$: initial gain

$G_{tot} = G_o / (1 - \sum_x g_x)$: total gain

Feedbacks are additive, but their effects are not!!

How do we incorporate the dynamics into feedback analysis?

- **Does atmospheric motion play a role in the climate response to the external forcing?**

- **Even for a global uniform SURFACE warming, what are the roles of evaporation and surface sensible heat flux?**

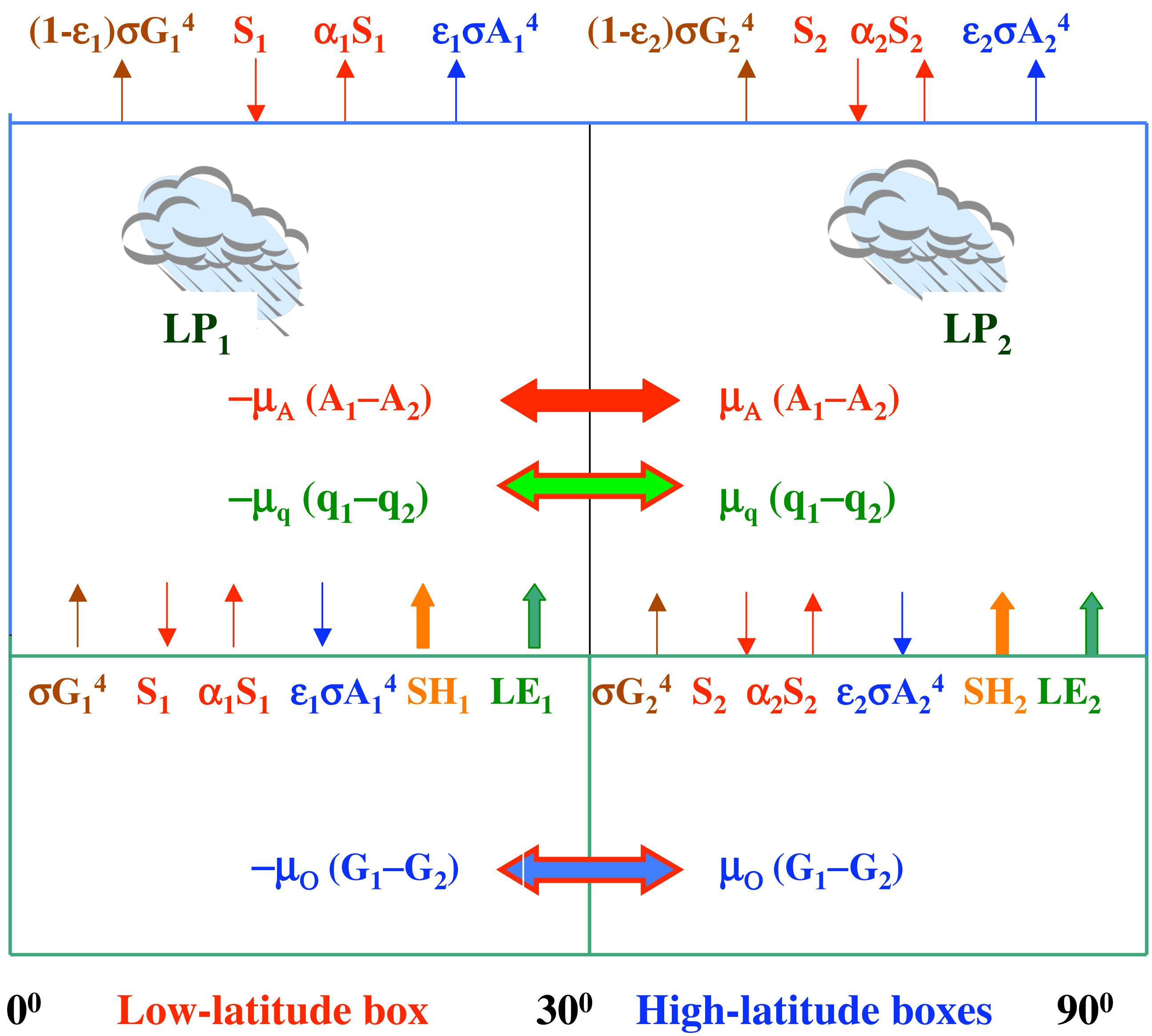
It turns out they are hidden in the lapse rate feedback!!!

Illustration of the new feedback analysis in a simple climate model

The science question:

Can the surface warming in response to anthropogenic greenhouse gases be still stronger in high latitudes than in low latitudes in the absence of ice-albedo feedback?

A coupled atmosphere-land/ocean moist radiative-transportive climate model



Coupled Response to external and feedbacks

(A prototype model that leads to the CFRAM, Cai and Lu 2007)

$$\Delta G_j = \frac{1}{4\sigma G_{Ej}^3}\{\underbrace{\frac{\sigma G_{Ej}^4 \Delta\varepsilon_{ext}}{(2-\varepsilon_{Ej})}} + \frac{\sigma G_{Ej}^4 \Delta\varepsilon_j}{(2-\varepsilon_{Ej})} + \frac{2\Delta S_j}{2-\varepsilon_{Ej}} - \frac{\Delta F_j}{2-\varepsilon_{Ej}} + \frac{(-1)^j \Delta D}{2-\varepsilon_{Ej}}\}$$

$$\Delta A_j = \frac{1}{4\sigma A_{Ej}^3}\{\underbrace{(\sigma A_{Ej}^4 - \frac{Q_{Ej}}{\varepsilon_{Ej}^2})\frac{\Delta\varepsilon_{ext}}{(2-\varepsilon_{Ej})}} + (\sigma A_{Ej}^4 - \frac{Q_{Ej}}{\varepsilon_{Ej}^2})\frac{\Delta\varepsilon_j}{(2-\varepsilon_{Ej})}$$

$$+ \frac{\Delta S_j}{2-\varepsilon_{Ej}} + \frac{(1-\varepsilon_{Ej})\Delta F_j}{(2-\varepsilon_{Ej})\varepsilon_{Ej}} + \frac{(-1)^j \Delta D}{(2-\varepsilon_{Ej})\varepsilon_{Ej}}\}$$

Partial temperature changes due to (1) external forcing alone, (2) water vapor, (3) ice-albedo, (4) surface turbulent energy flux (5) (non-local) dynamical feedbacks.

Dry Model Solution

(Cai, 2005; Cai, 2006)

1. emissivity = constant;

2. Only partial temperature changes due to the external forcing alone and due to a change in the atmos. poleward sensible heat transport (non-local dynamical feedback)

Dry Model Solution

- Change in atmospheric equator-to-pole temperature contrast:

$$\Delta(A_1 - A_2) = \frac{\sigma A_{E1}^3 A_{E2}^3 + \mu_A \dfrac{A_{E1}^3 + A_{E2}^3}{\varepsilon^2}}{\left(4\sigma A_{E1}^3 A_{E2}^3 + \mu_A \dfrac{A_{E1}^3 + A_{E2}^3}{(2-\varepsilon)\varepsilon}\right)} (A_{E1} - A_{E2}) \frac{\Delta\varepsilon}{(2-\varepsilon)} > 0$$

where A_{Ej} *are the equilibrium air temperatures for* $\Delta\varepsilon = 0$.

- Change in the surface *temperatures*:

$$\Delta G_j = \frac{G_{Ej}}{4} \frac{\Delta\varepsilon}{(2-\varepsilon)} + (-1)^j \mu_A \frac{\Delta(A_1 - A_2)}{4\sigma G_{Ej}^3 (2-\varepsilon)}$$

j = 1: low latitudes

j = 2: high latitudes

where G_{Ej} *are the equilibrium surface temperatures for* $\Delta\varepsilon = 0$.

How can it be possible that an increase of air temperature gradient can cause a reduction of the surface temperature gradient?

Partial temperature changes in the dry model

$$\Delta A_j = \frac{1}{4\sigma A_{Ej}^3}\{\underbrace{(\sigma A_{Ej}^4 - \frac{Q_{Ej}}{\varepsilon_{Ej}^2})\frac{\Delta\varepsilon_{ext}}{(2-\varepsilon_{Ej})}} + \frac{(-1)^j \Delta D}{(2-\varepsilon_{Ej})\varepsilon_{Ej}}\}$$

$$\Delta G_j = \frac{1}{4\sigma G_{Ej}^3}\{\underbrace{\frac{\sigma G_{Ej}^4 \Delta\varepsilon_{ext}}{(2-\varepsilon_{Ej})}} + \frac{(-1)^j \Delta D}{2-\varepsilon_{Ej}}\}$$

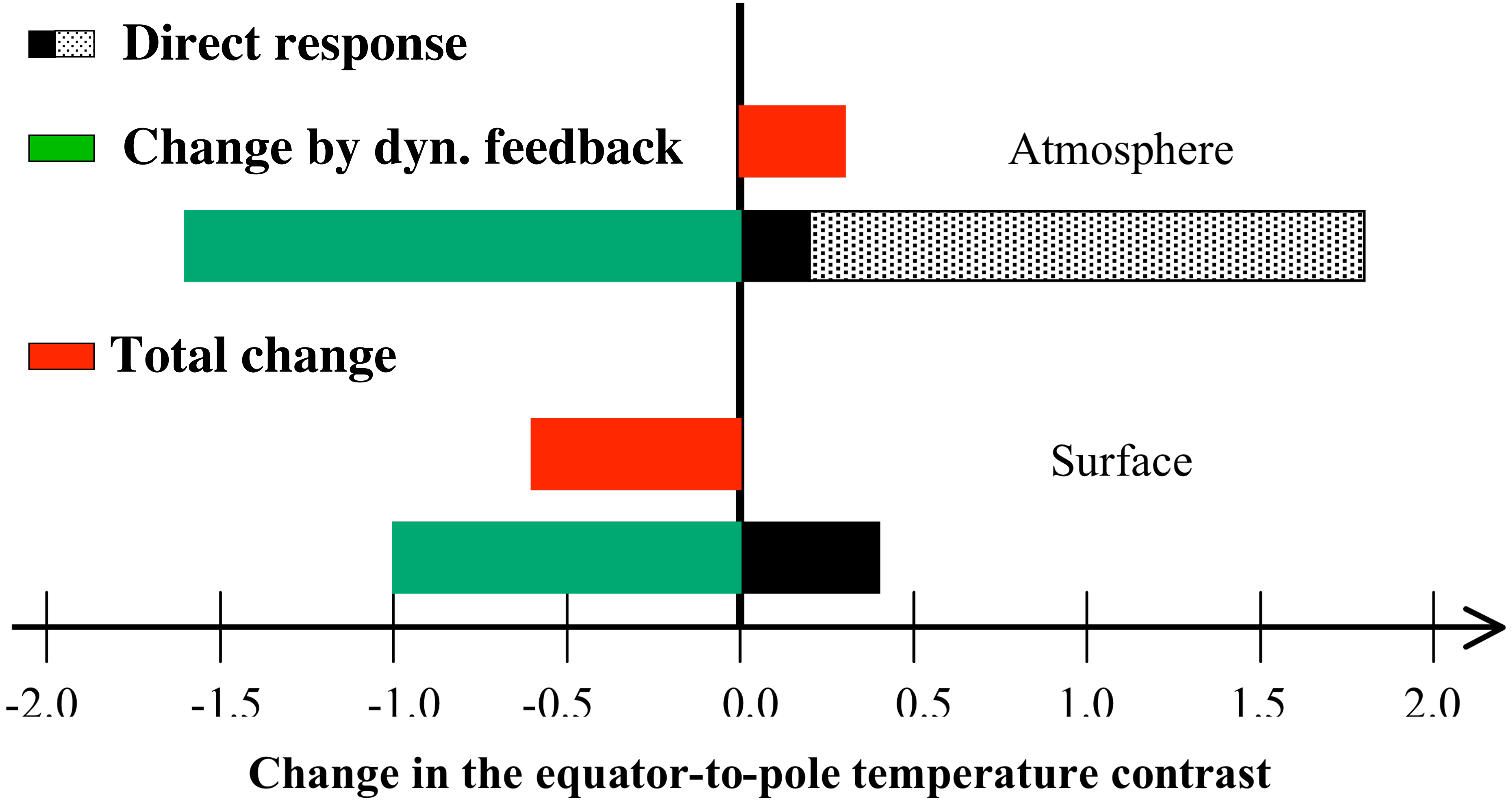
j = 1: low latitudes

j = 2: high latitudes

The additional SURFACE warming in high latitudes is due to the more "BACK-RADIATION" resulting from the increase poleward heat transport ($\Delta D > 0$) => "greenhouse-plus" feedback

The reduction of SURFACE warming in high latitudes is due to the less "BACK-RADIATION" resulting from $\Delta D > 0$ => "greenhouse-minus" feedback in low latitudes.

Change of meridional temperature gradient due to external forcing alone versus that due to dynamic feedback in the dry model



Coupled Atmosphere-Surface Climate Feedback-Response Analysis Method (CFRAM) for CGCM feedback analysis

(Lu & Cai 2008; Cai & Lu 2008)

- **Forcing: an external perturbation profile in the atmosphere-surface column at each grid point**

- **Response: a vertically varying atmosphere-surface temperature profile at each grid point.**

- **Feedback: any energy flux perturbations that are not caused by the the longwave radiation change due to temperature changes.**

Mathematical formulation

$$\left(\frac{\partial \overline{\mathbf{R}}}{\partial \overline{\mathbf{T}}}\right)\Delta\overline{\mathbf{T}}^{tot} = \{\Delta\overline{\mathbf{F}}^{ext} + \Delta^{(\alpha)}\vec{\mathbf{S}} + \Delta^{(c)}(\vec{\mathbf{S}} - \overline{\mathbf{R}}) + \Delta^{(w)}(\vec{\mathbf{S}} - \overline{\mathbf{R}})$$

$$+\Delta\overline{\mathbf{Q}}^{conv} + \Delta\overline{\mathbf{Q}}^{turb} - \Delta\overline{\mathbf{D}}^{v} - \Delta\overline{\mathbf{D}}^{h} + \Delta\overline{\mathbf{W}}^{fric}\}$$

The radiation flux change only due to a change in the atmosphere-surface column temperature

=

Radiative energy input due to the external forcing

+

Energy flux perturbations that are not due to the radiation change associated with temperature changes

Mathematical formulation

$$\Delta\vec{\overline{\mathbf{T}}}^{tot} = \left(\frac{\partial\vec{\overline{\mathbf{R}}}}{\partial\vec{\overline{\mathbf{T}}}}\right)^{-1} \{\Delta\vec{\overline{\mathbf{F}}}^{ext} + \Delta^{(\alpha)}\vec{\overline{\mathbf{S}}} + \Delta^{(c)}(\vec{\overline{\mathbf{S}}} - \vec{\overline{\mathbf{R}}}) + \Delta^{(w)}(\vec{\overline{\mathbf{S}}} - \vec{\overline{\mathbf{R}}})$$

$$+\Delta\vec{\overline{\mathbf{Q}}}^{conv} + \Delta\vec{\overline{\mathbf{Q}}}^{turb} - \Delta\vec{\overline{\mathbf{D}}}^{v} - \Delta\vec{\overline{\mathbf{D}}}^{h} + \Delta\vec{\overline{\mathbf{W}}}^{fric}\}$$

RHS: external forcing plus energy flux perturbations due to each of (thermodynamic, local, and non-local dyn. feedbacks

$$\Delta\vec{\overline{\mathbf{T}}}^{(n)} = \left(\frac{\partial\vec{\overline{\mathbf{R}}}}{\partial\vec{\overline{\mathbf{T}}}}\right)^{-1} \Delta\vec{\overline{\mathbf{F}}}^{(n)}$$

$$\Delta\vec{\overline{\mathbf{T}}}^{tot} = \sum_{n} \Delta\vec{\overline{\mathbf{T}}}^{(n)}$$

Feedback Gain Matrices in CFRAM

$$\Delta\vec{\mathbf{T}}^{tot} = \mathbf{G}\Delta\vec{\mathbf{F}}^{ext} = \mathbf{G}_0(\mathbf{I} + \sum_{n>0}\mathbf{g}^{(n)})\Delta\vec{\mathbf{F}}^{ext}$$

$$\mathbf{G}_0 = \left(\frac{\partial\vec{\mathbf{R}}}{\partial\vec{\mathbf{T}}}\right)^{-1} = \begin{pmatrix} r_{1,1} & \cdots & r_{1,M+1} \\ \vdots & \ddots & \vdots \\ r_{M+1,1} & \cdots & r_{M+1,M+1} \end{pmatrix}$$

Initial gain matrix= inverse of the Planck feedback matrix

Both feedbacks and their effects are additive!

$$\mathbf{g}^{(n)} = \begin{pmatrix} g_{1,1}^{(n)} & 0 & \cdots & 0 \\ 0 & g_{2,2}^{(n)} & 0 & \vdots \\ \vdots & 0 & \ddots & 0 \\ 0 & \cdots & 0 & g_{M+1,M+1}^{(n)} \end{pmatrix}$$

$$g_{i,i}^{(n)} = \frac{\sum_{m=1}^{M+1} r_{i,m}\Delta F_m^{(n)}}{\sum_{m=1}^{M+1} r_{i,m}\Delta F_m^{ext}}$$

Feedback gain matrix

What happens to the lapse rate?

Vertical summation from the TOA to surface

$$(-\sum_{j=1}^{M+1}\frac{\partial R^{toa}}{\partial T_j})\Delta T_s + \sum_{j=1}^{M}(-\frac{\partial R^{toa}}{\partial T_j})(\Delta T_j - \Delta T_s)$$
$$+\Delta^{(\alpha)}S^{toa} + \Delta^{(c)}(S^{toa} - R^{toa}) + \Delta^{(w)}(S^{toa} - R^{toa}) - \Delta D = -\Delta F^{toa}$$

Feedback parameters in PRP

$$\lambda_{tot} = \frac{-\Delta F^{toa}}{\Delta T_s} = \lambda_P + \lambda_\Gamma + \lambda_\alpha + \lambda_c + \lambda_w + \lambda_D$$

Lapse rate feedback

$$\lambda_\Gamma = \sum_{j=1}^{M+1}(-\frac{\partial R^{toa}}{\partial T_j})\frac{\Delta T_j - \Delta T_s}{\Delta T_s}$$

Lapse rate feedback decomposition

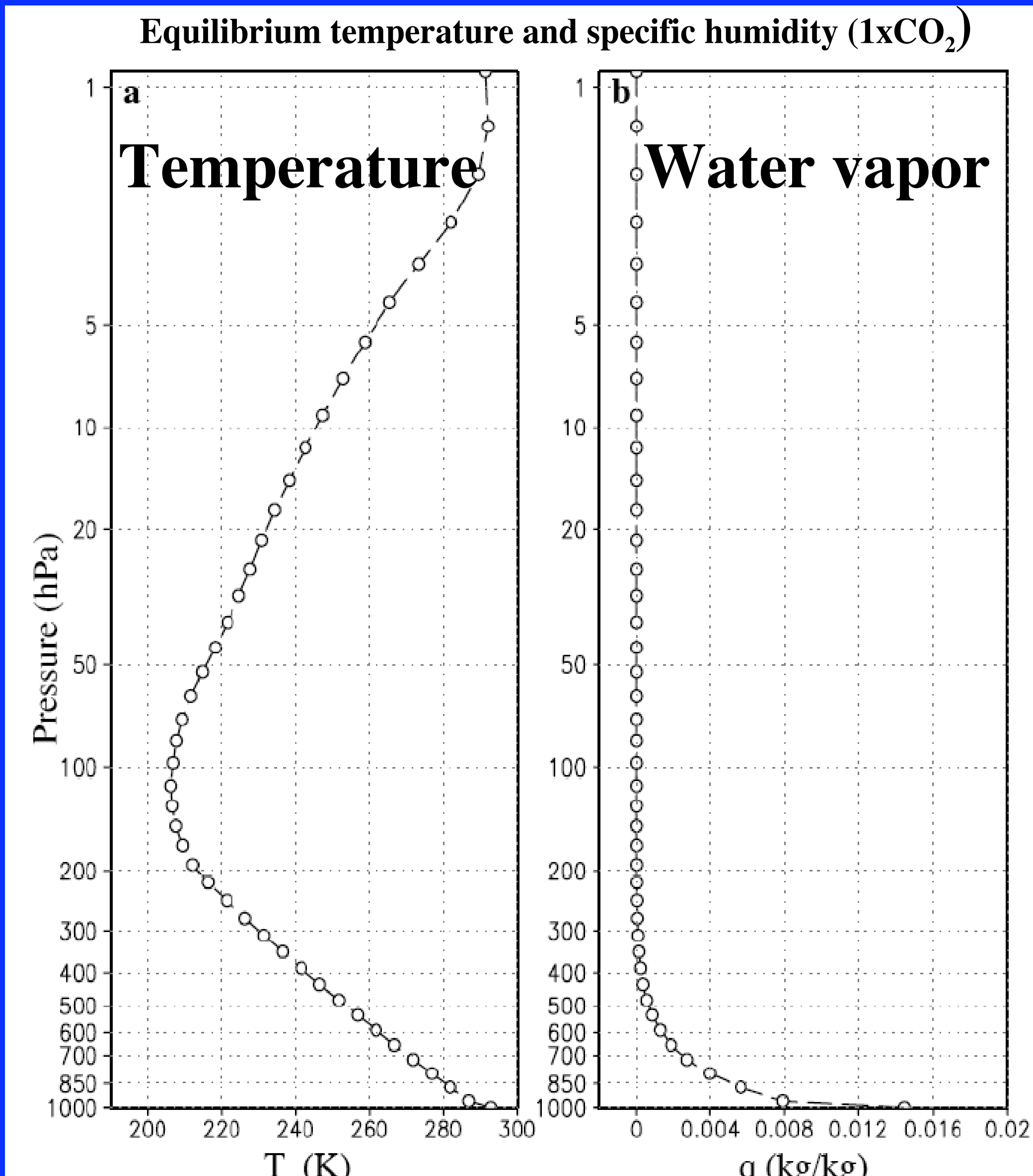
$$\sum_{j=1}^{M+1}(-\frac{\partial R^{toa}}{\partial T_j})(\frac{\Delta T_j - \Delta T_s}{\Delta T_s}) = \sum_{n=0}^{N}\{\sum_{j=1}^{M+1}(-\frac{\partial R^{toa}}{\partial T_j})(\frac{\Delta T_j^{(n)} - \Delta T_{M+1}^{(n)}}{\Delta T_s})\}$$
$$= \sum_{n=0}^{N}\lambda_\Gamma^{(n)} = \lambda_\Gamma$$

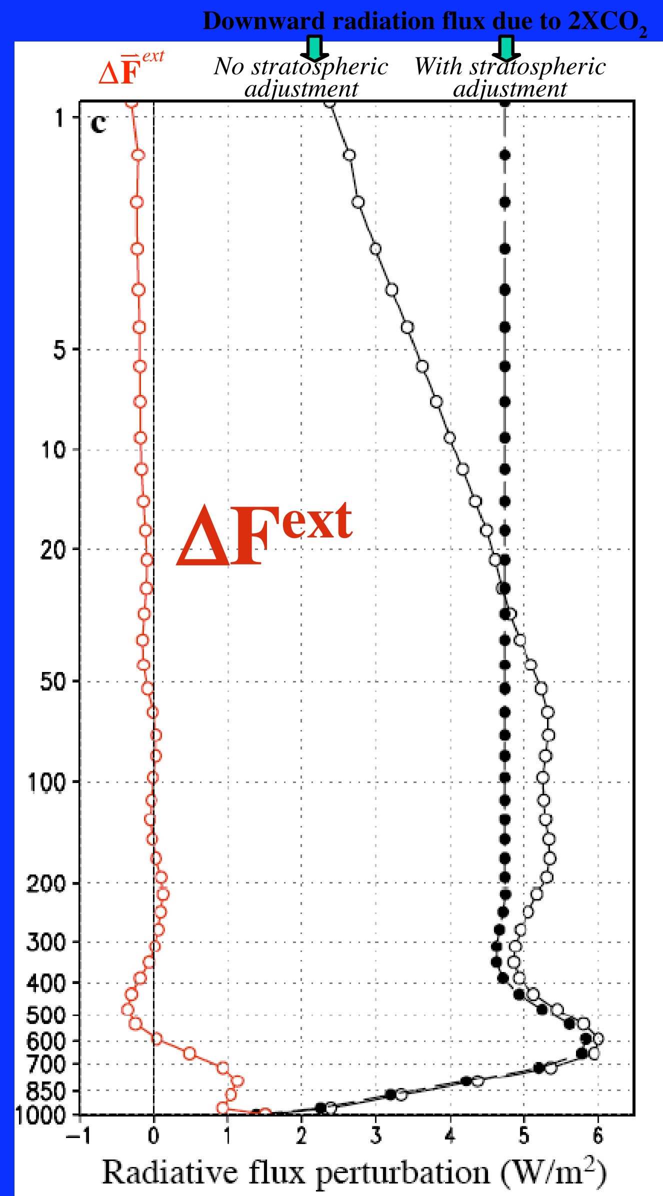
Demonstration of the CFRAM in the context of a <u>**single-column radiative-convective model**</u>

- **Climate perturbation simulations by doubling CO2 in the model.**

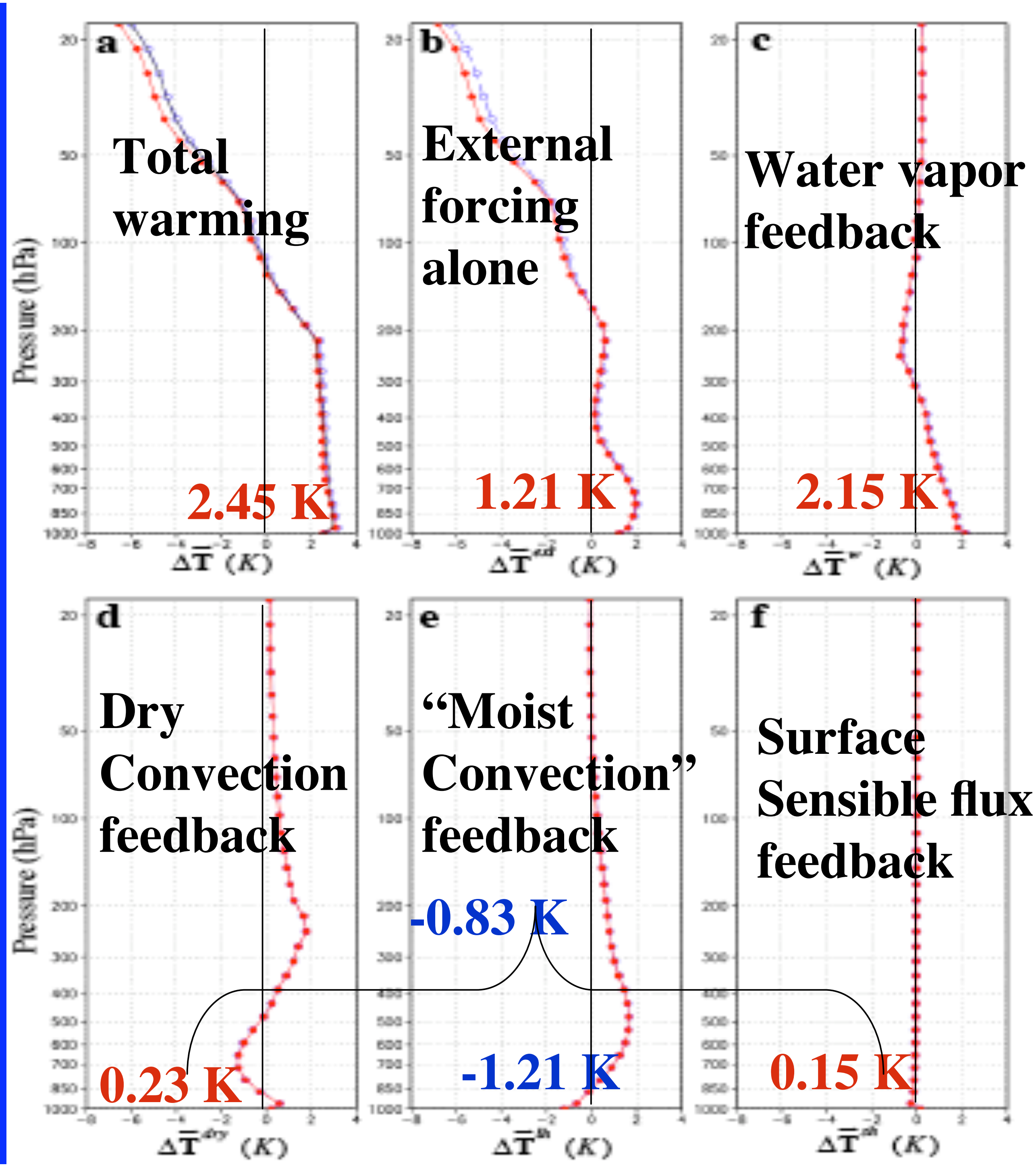
- **Feedbacks: water vapor, surface sensible and dry convection, evaporation and "moist convection" feedbacks**

Model Climate and Climate forcing





Partial versus total temperature changes



PRP method and Lapse Rate Feedback Decomposition

Uniform	Lapse Rate	Total (sum of the left)
$g_w = 0.586$ (sum of the above)	$g_\Gamma = -0.067$ (sum of the above)	Total feedback gain: $g^{tot} = 0.519$
Initial Gain		$G_0 = 1/(-\lambda_p) = 0.257$ K/(Wm^{-2})
Total Gain		$G = G_0(1-g^{tot})^{-1} = 0.534K/(Wm^{-2})$

Demonstration of the CFRAM in the context of a GCM without hydrological cycle

(manuscript in preparation)

The science question:

Can the surface warming in response to anthropogenic greenhouse gases be still stronger in high latitudes than in low latitudes in the absence of ice-albedo feedback in a GCM model?

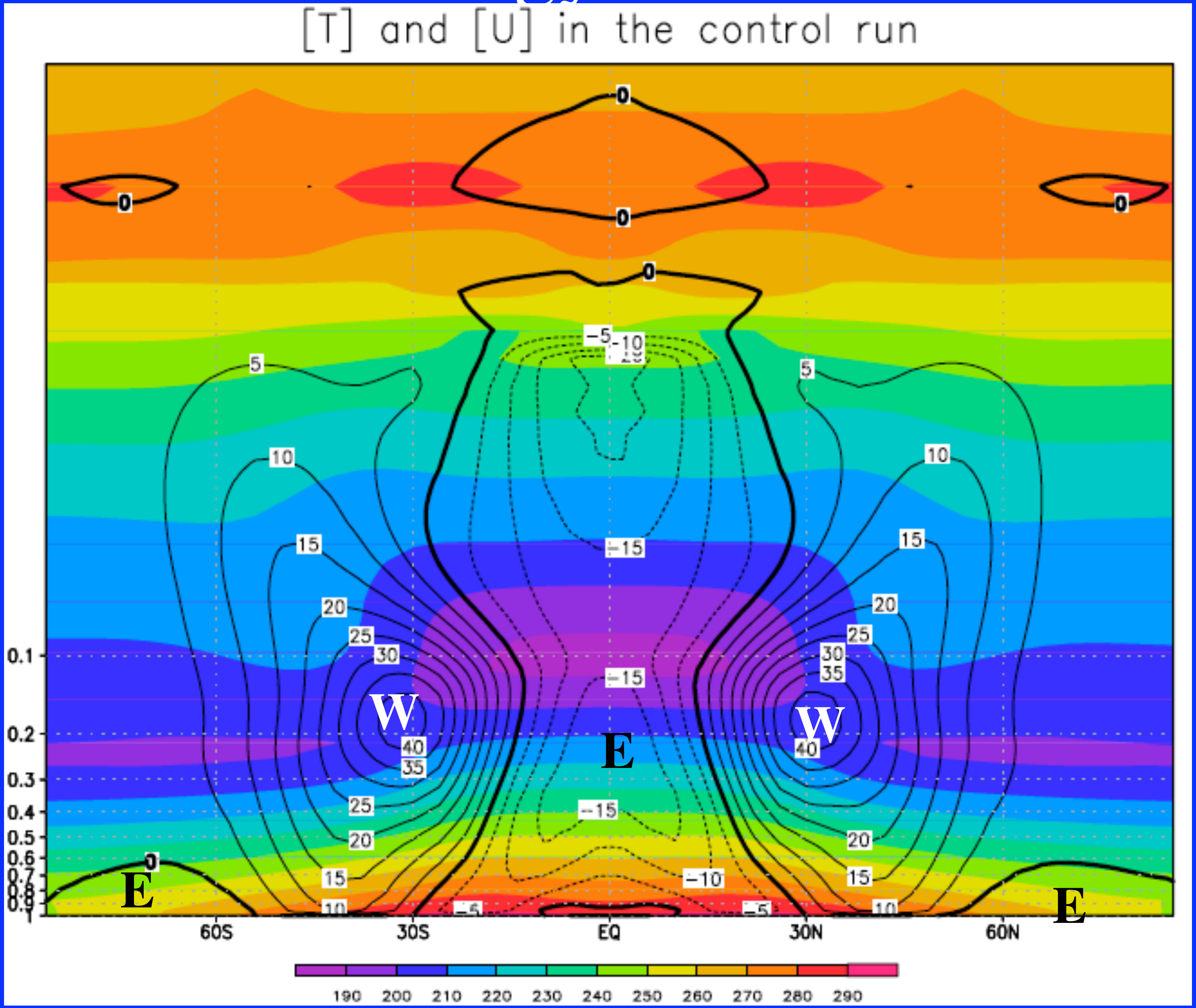
The key features of the GCM model

Dynamical core: Suarez and Held (1992)

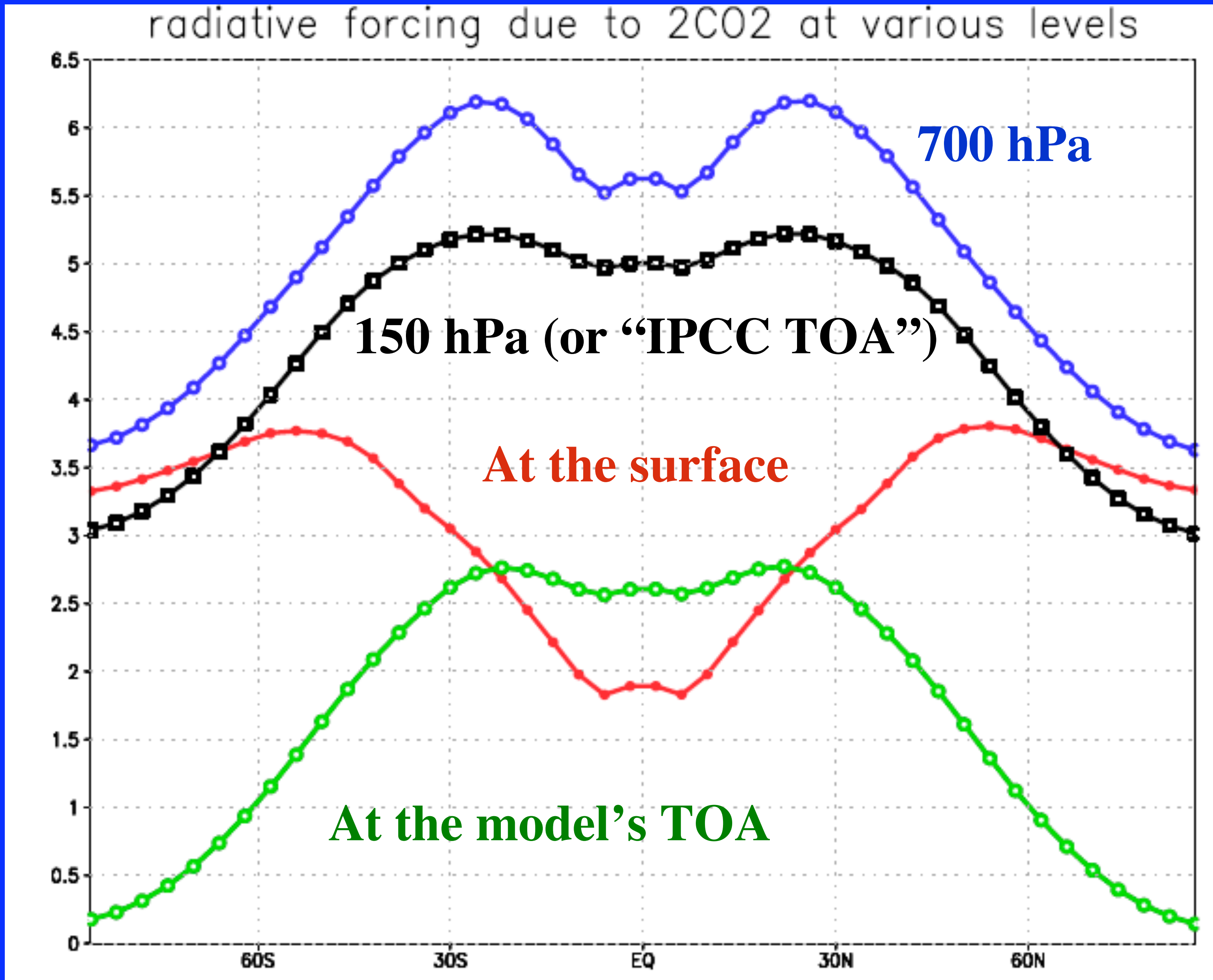
Physics:

- **Fu et al. (1992)'s radiation model.**
- **Dry convection adjustment so that maximum lapse rate cannot exceed a preset meridional profile (6.5K/1km in tropics and 9.8K/1km outside).**
- **Atmospheric relative humidity is kept at a preset vertical and meridional profile.**
- **The surface energy balance model that exchanges sensible heat flux, emits long wave radiation out, and absorbs downward radiation at the surface.**
- **The annual mean solar forcing.**
- **1CO2 versus 2CO2 climate simulations**

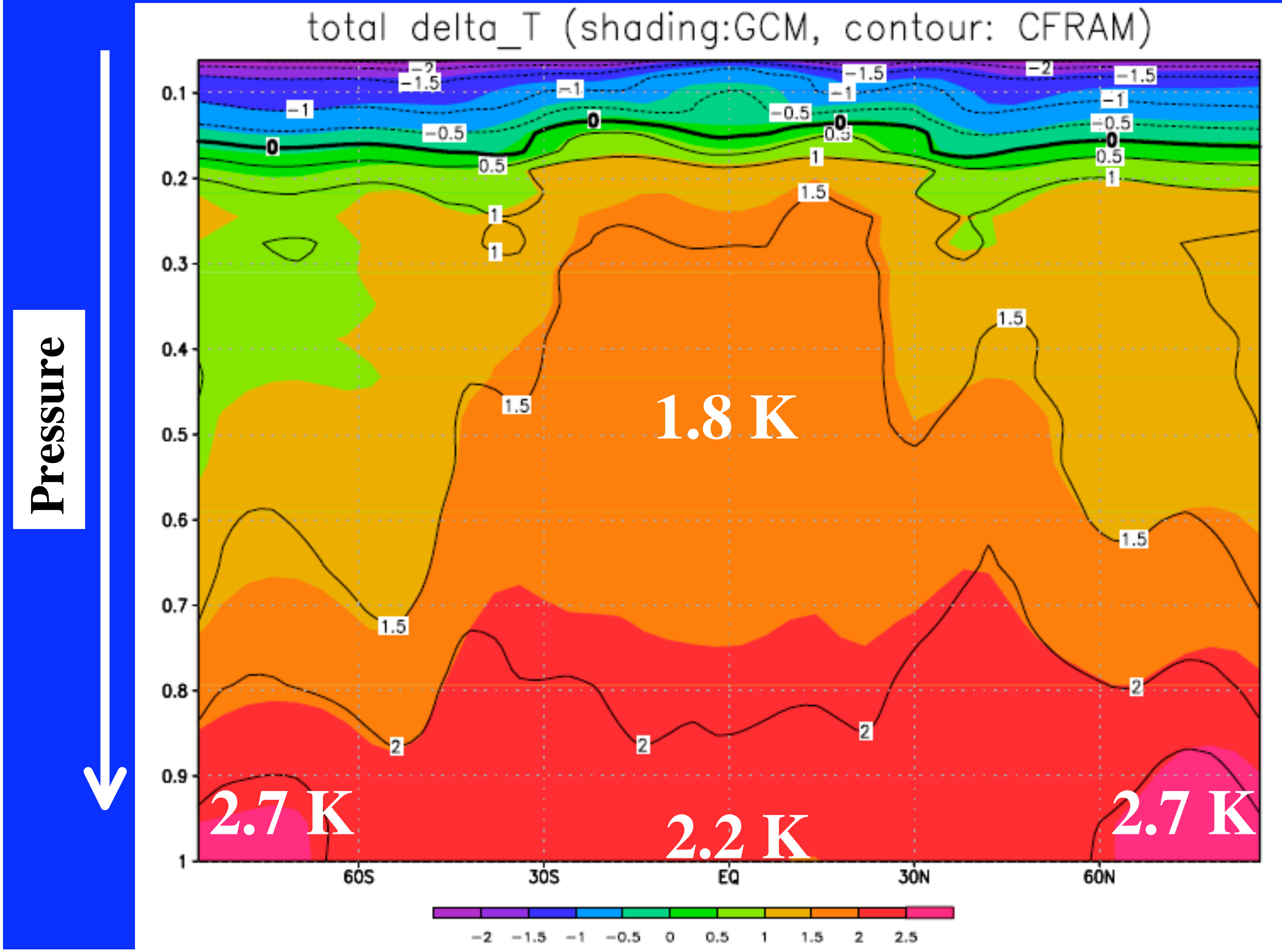
Climatology of the GCM



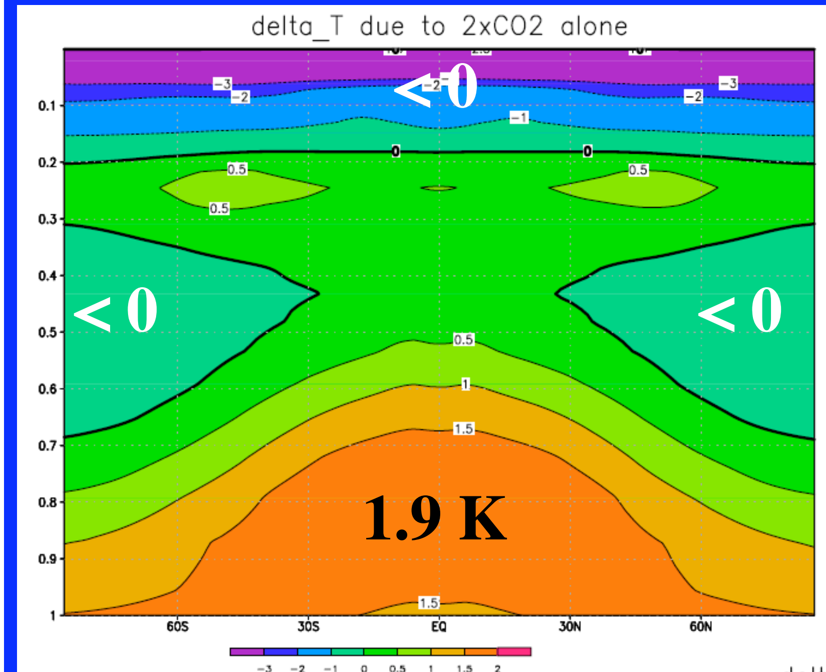
2CO2 Climate forcing (w/m2) in the GCM

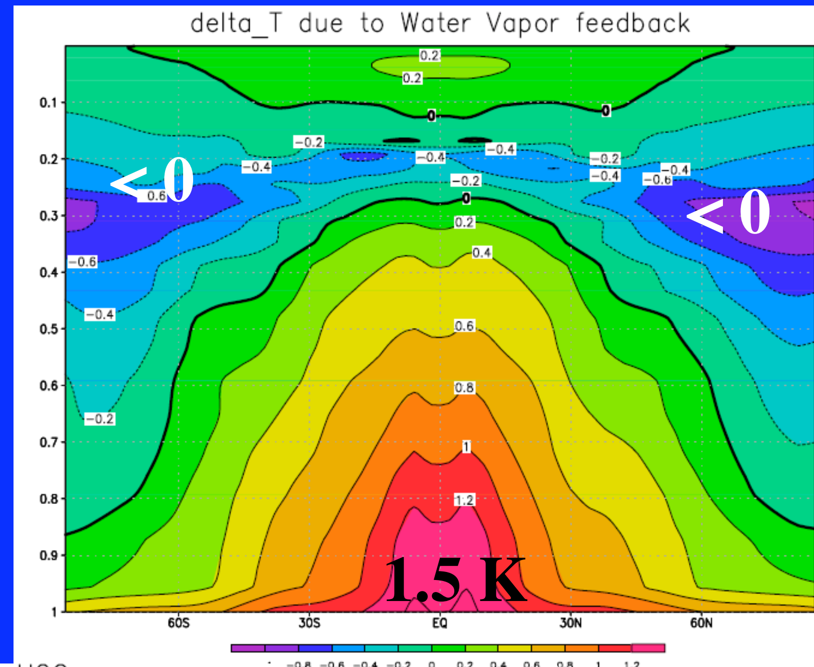


(Total) Warming pattern



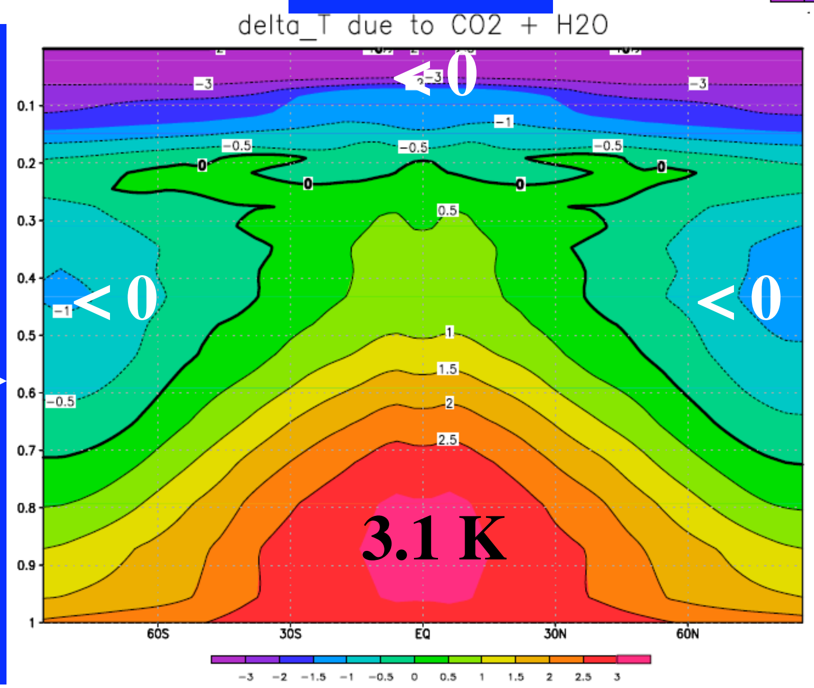
Partial ΔT due to 2CO2 & H2O





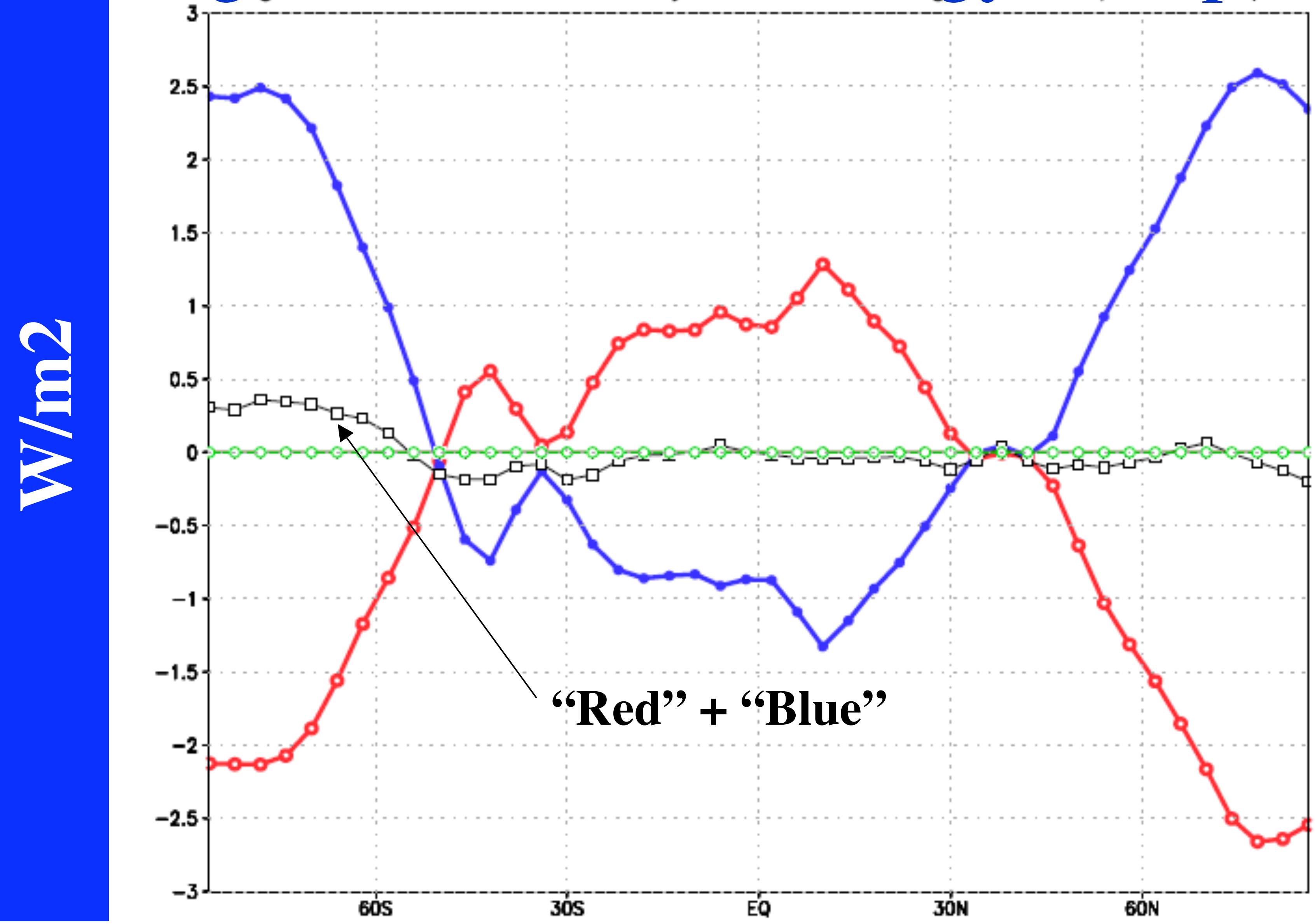
2CO2 only

H2O change only

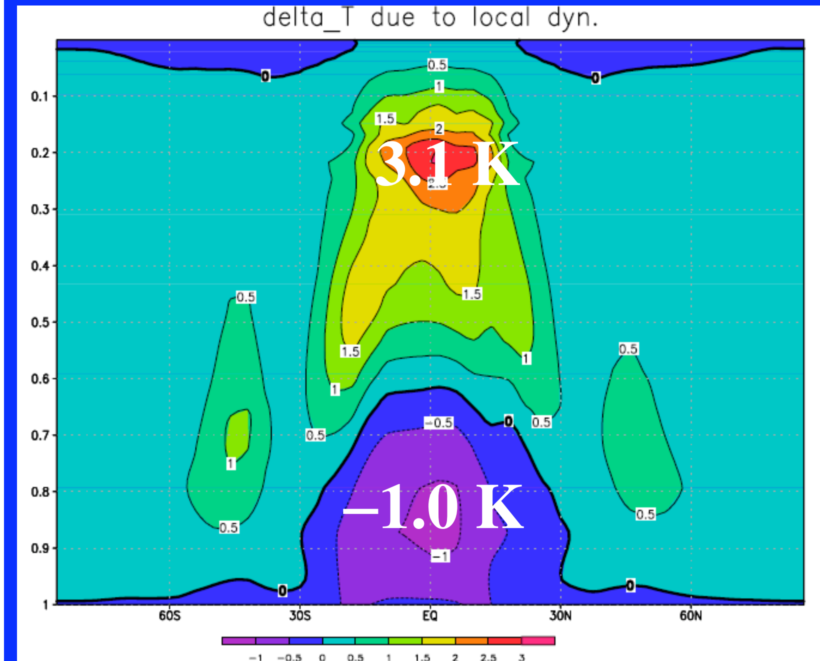


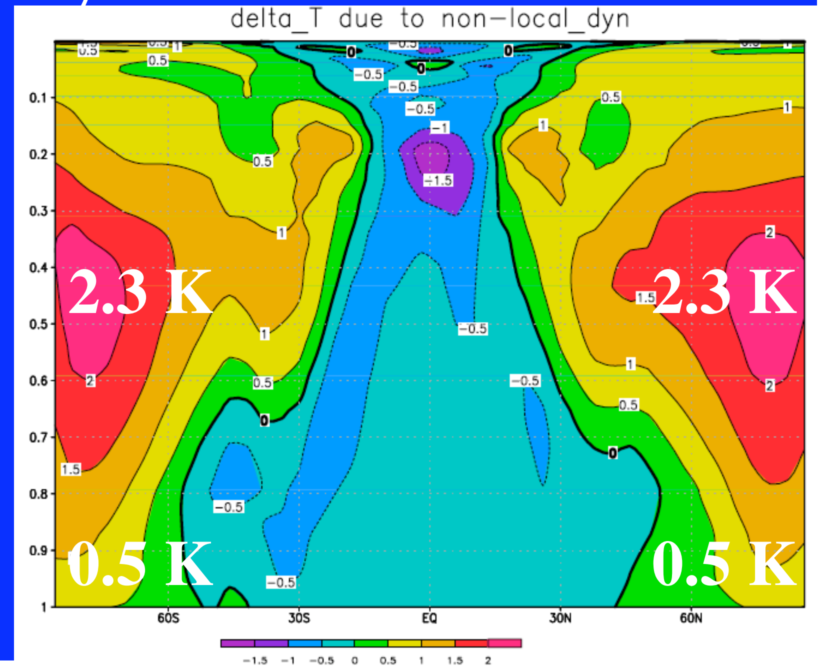
2CO2 + Δ(H2O)

Change in TOA (S – R) and vertically integrated horizontal energy transport



Partial ΔT due to dynamics

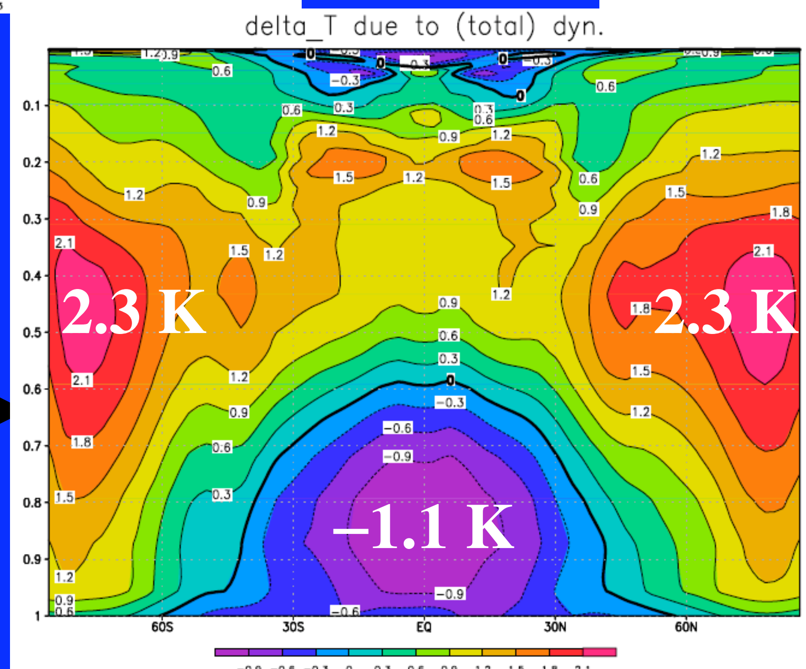




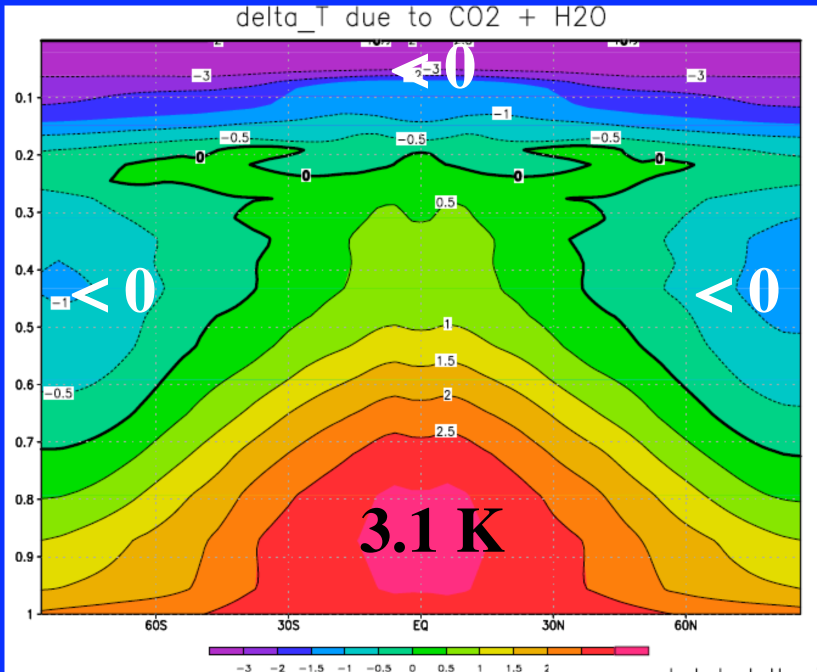
Vertical transport

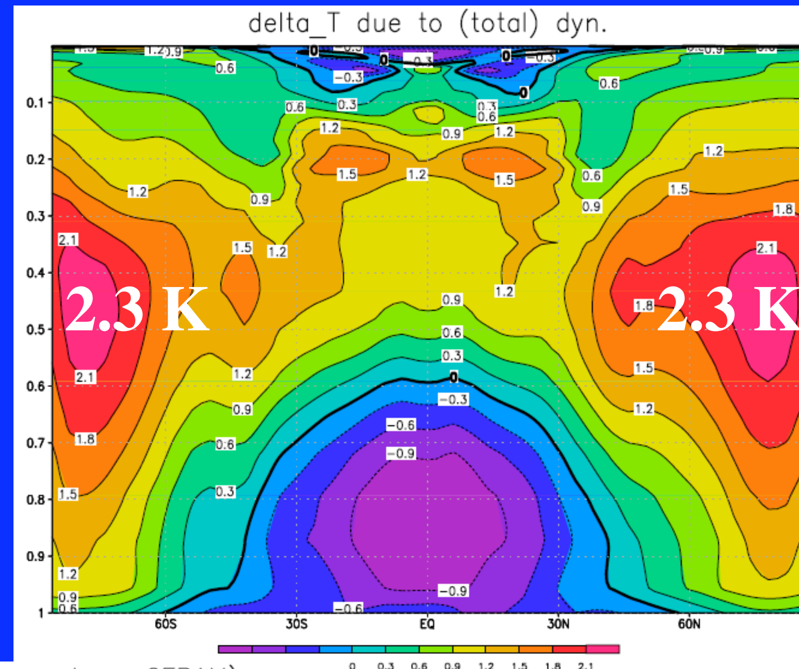
Horizontal transport

Total dynamical feedback



Sum of partial ΔTs

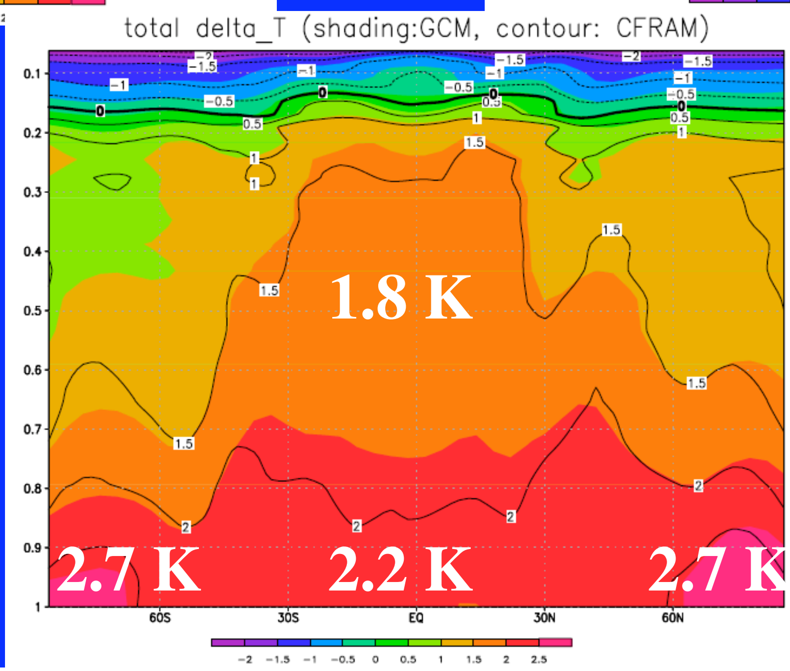




2CO2 + Δ(H2O)

Total dynamical feedback

Total change



Summary

- Radiative forcing of greenhouse gases (including water vapor) tend to cause a stronger warming in low latitudes and weaker warming in high latitudes in atmos. and surf..
- Vertical convection reduces the surface warming in tropics and an enhanced poleward heat transport results in a "greenhouse-plus" feedback (more back radiation from the air to the surface) => a large SURFACE warming in high latitudes even without ice-albedo feedbacks!
- Part of the total effects of individual thermodynamic and non-local dynamical feedbacks and the total effects of all local dynamical feedbacks are lumped into the lapse rate feedback in a TOA-based framework.
- The CFRAM allows us to explicitly examine the roles of both thermodynamic and dynamical feedback processes in giving rise to the observed warming pattern.

In a realistic model with water cycle:

We expect:

- A much stronger reduction of the surface warming in tropics/subtropics due to the evaporation feedback (about 1-2 K more warming reduction).
- Stronger moist convection in the deep tropics brings energy to further up => stronger poleward heat transport (including the latent heat transport) => a larger dynamical warming amplification.
- Ice albedo feedback => further strengthens the polar warming amplification.
- Role of clouds? But the CFRAM can help to answer that question!