Dynamical Polar Warming Amplification and a New Climate Feedback Framework

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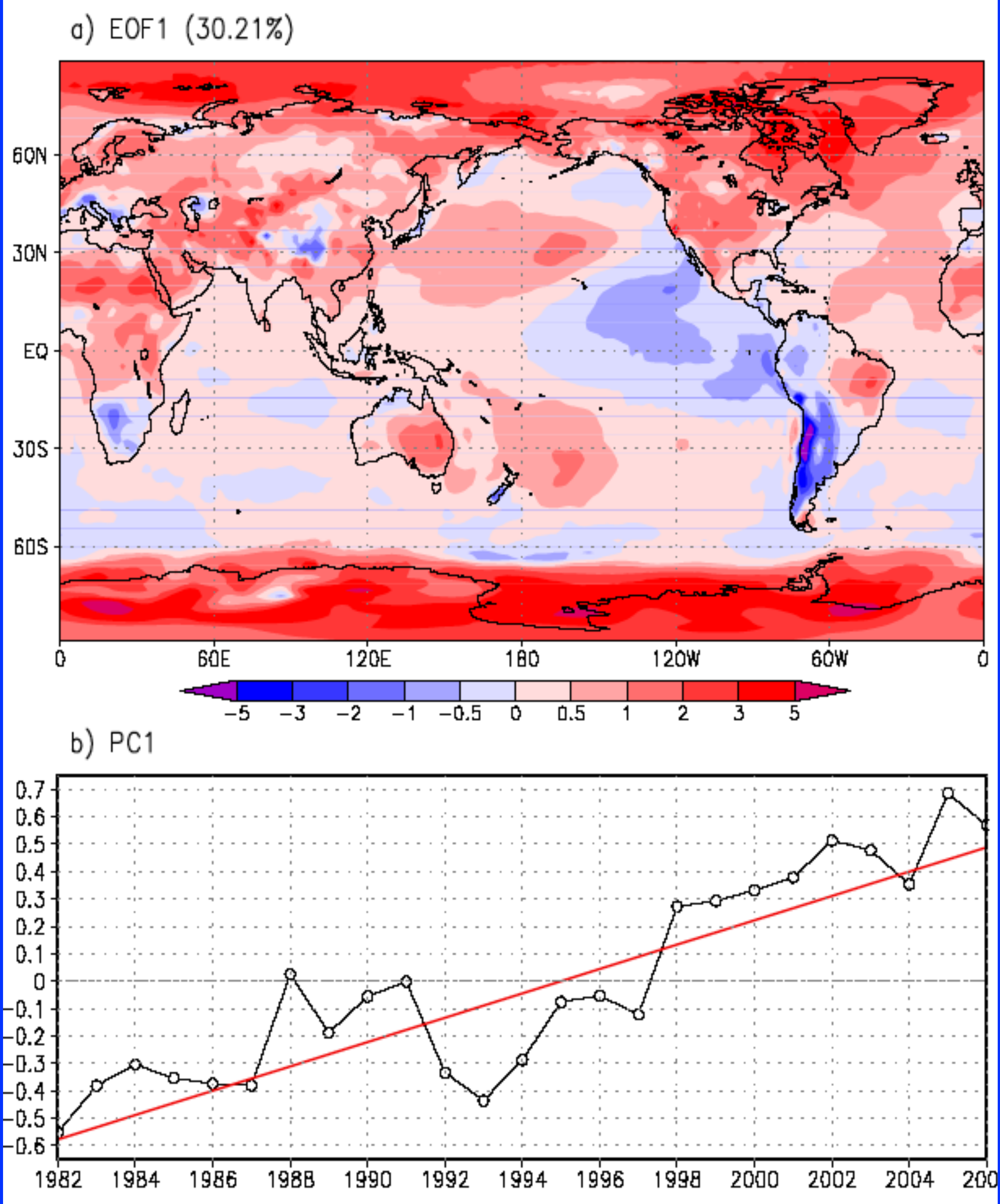
Cai (2005, 2006), Cai and Lu (2007): DPWA Theory
Lu and Cai (2009), Cai and Lu (2009): CFRAM
Lu and Cai (2010) Cai and Tung (2013): Dynamical PWA in “dry” GCM
Taylor et al. (2013): all factors to PWA in NCAR CCSM4
Sergio et al. (2013): Seasonal cycle of PWA in CCSM4

Observed global warming pattern (NCEP R2)



1st EOF (30% var.)

0.22 K/decade

Question 1

Is it POSSIBLE that a stronger surface warming in high latitudes than low latitudes in response to anthropogenic greenhouse gases can be CAUSED by the atmospheric poleward heat transport in the ABSENCE of (positive) ice-albedo feedback in high latitudes and (negative) evaporation feedback in low latitudes?

This question might sound paradoxical, given the fact the atmospheric poleward heat transport itself is driven by the poleward decreasing temperature profile.

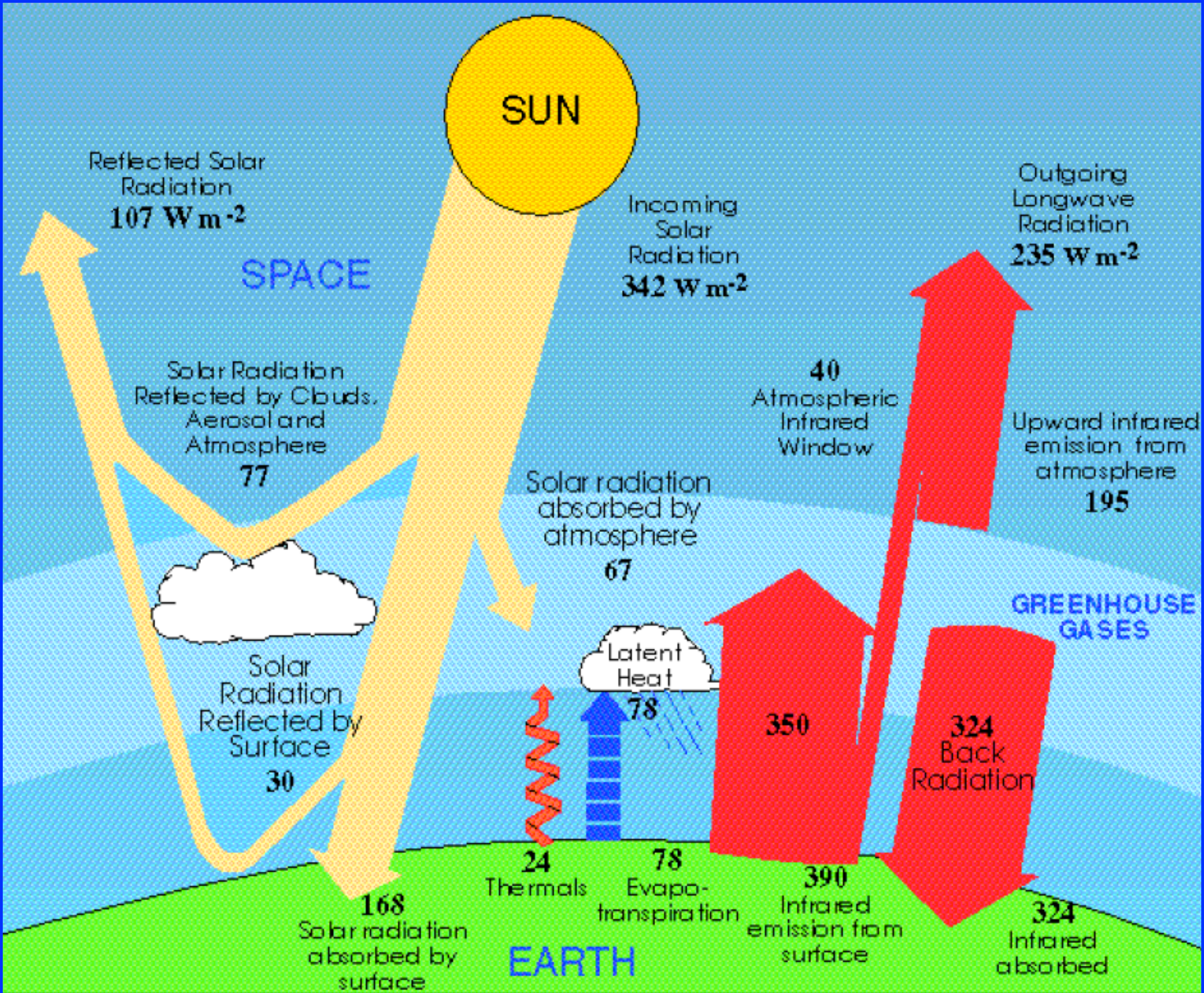
Cai (2005, 2006) answered this question: Yes

Question 2

Can polar surface warming amplification by atmospheric dynamic feedback exist without polar warming amplification in the troposphere?

Cai (2005, 2006) showed how and why an increase of air temperature gradient can still cause (i) a polar surface warming amplification (ii) a stronger surface warming than air warming in polar region.

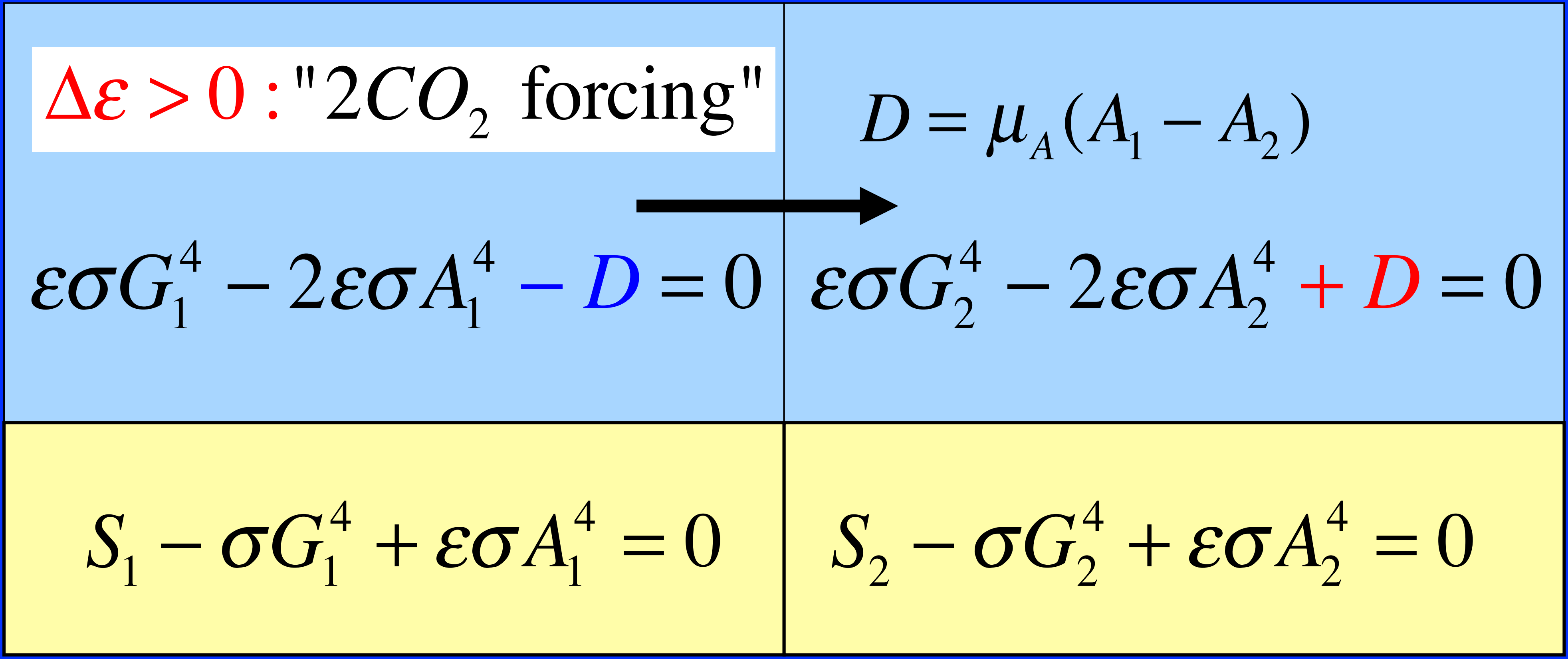
Global mean atmosphere and surface energy balance: A single column perspective



A dry radiative-transportive model

(Cai, 2005; 2006)

D=0=>TOA=0 **TOA>0** **TOA<0** **D=0=>TOA=0**



$\Delta\varepsilon > 0$: "$2CO_2$ forcing"

$$D = \mu_A (A_1 - A_2)$$

$$\varepsilon\sigma G_1^4 - 2\varepsilon\sigma A_1^4 - D = 0$$

$$\varepsilon\sigma G_2^4 - 2\varepsilon\sigma A_2^4 + D = 0$$

$$S_1 - \sigma G_1^4 + \varepsilon\sigma A_1^4 = 0$$

$$S_2 - \sigma G_2^4 + \varepsilon\sigma A_2^4 = 0$$

0° **Low-lat. Boxes** 30° **High-lat. Boxes** 90°

A: air temperature; G: surface temperature

j = 1: low latitudes; j = 2: high latitudes

Analytic Solution of the 4-box dry model

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

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

where A_{Ej} *are the ("1CO2") equilibrium air temperatures for* $\Delta\varepsilon = 0$.

- **Change in the surface equator-to-pole temperature contrast**

$$\Delta(G_1 - G_2) = \frac{G_{E1} - G_{E2}}{4}\frac{\Delta\varepsilon}{(2-\varepsilon)} - 2\mu_A \frac{\Delta(A_1 - A_2)}{4\sigma G_{Ej}^3(2-\varepsilon)}$$

where G_{Ej} *are the ("1CO2")equilibrium surf. temperatures for* $\Delta\varepsilon = 0$.

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

Partial temperature changes in the dry model

(A prototype model that leads to the CFRAM)

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

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

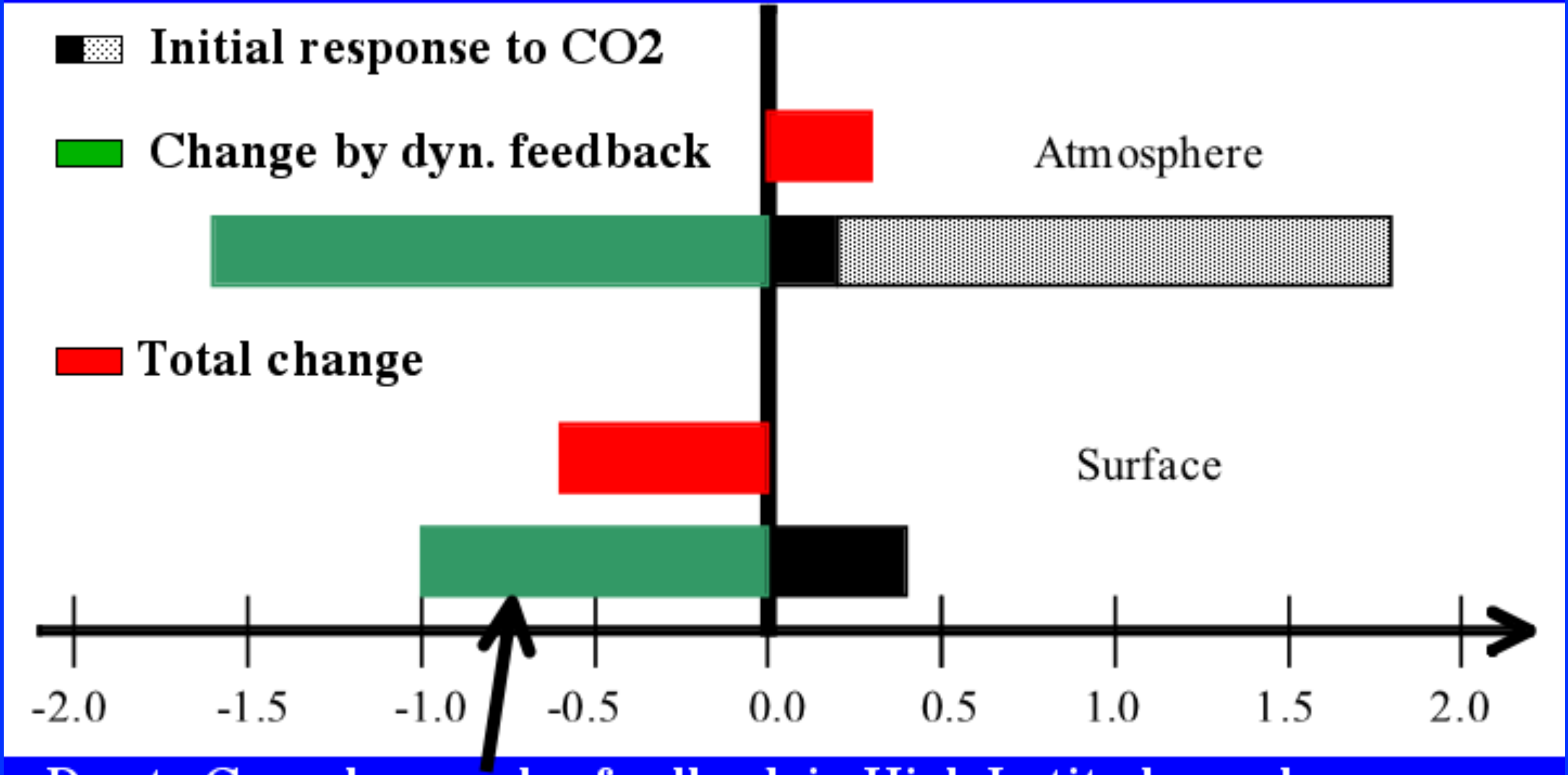
j = 1: low latitudes

j = 2: high latitudes

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

The reduction of SURFACE warming in low latitudes is due to the less “BACK-RADIATION” from a colder atmosphere in low latitudes 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



Due to Greenhouse-plus feedback in High Latitudes and Greenhouse-minus feedback in Low Latitudes

Figure 9 in Cai (2006)

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

(designed for a globally uniform SURFACE warming)

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"
    - Positive feedback: enhance the original energy input.
    - Negative feedback: reduce/oppose the original energy input.

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**

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

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

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

$\lambda_{tot} < 0$: (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

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

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

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

$$=\lambda_p\{1-(\lambda_{H_2O}+\lambda_{albedo}+\lambda_{cloud}+\lambda_{lapse\_rate})/(-\lambda_p)\}$$

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

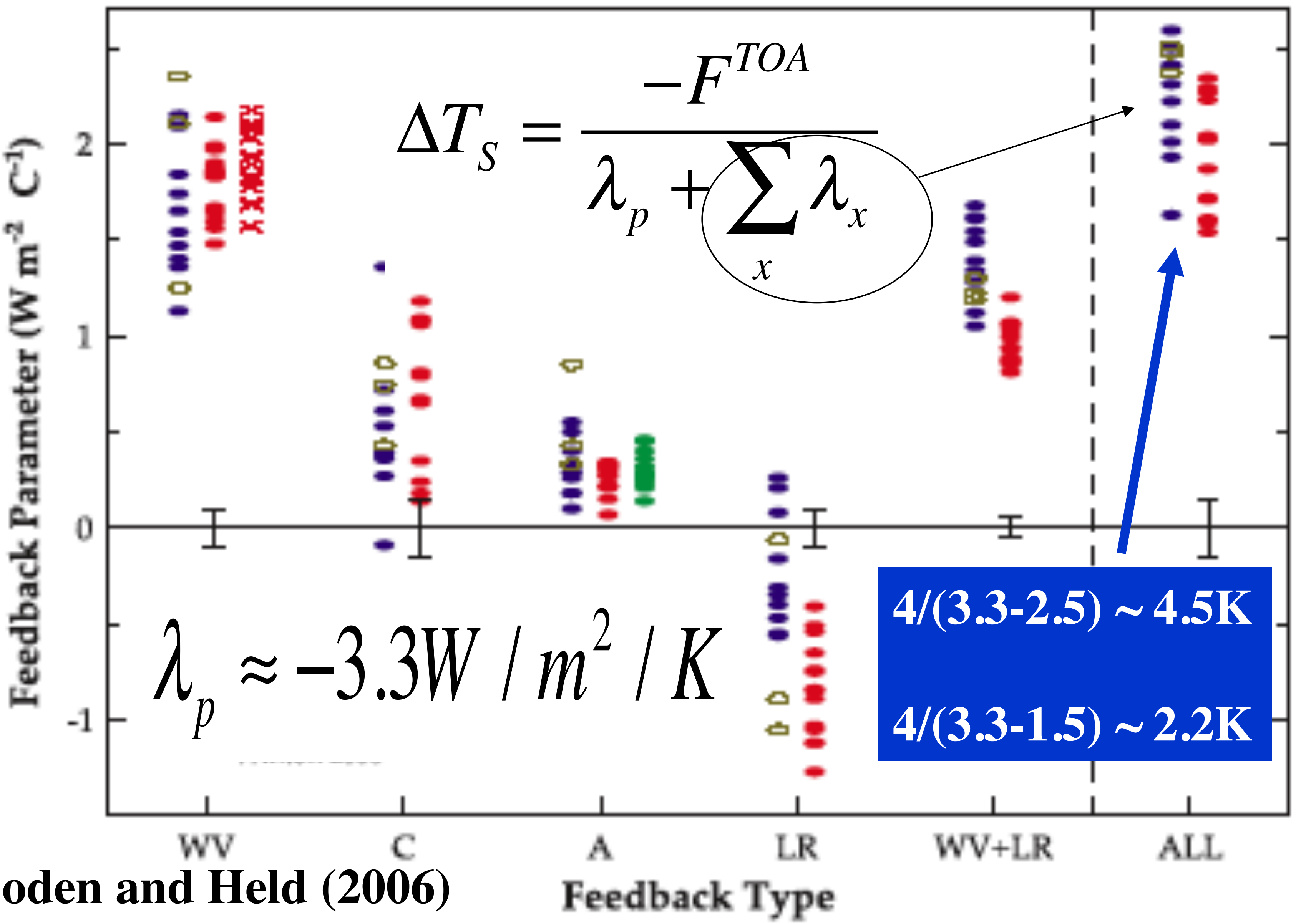
$$\Delta T_S=\frac{\Delta F^{TOA}}{-(\lambda_P+\sum_x\lambda_x)}=\frac{G_0\Delta F^{TOA}}{1-\sum_x g_x}$$

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

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

Feedbacks are additive, but their effects are not!!

Climate feedbacks and climate projection uncertainties (IPCC AR4)



$$\Delta T_S = \frac{-F^{TOA}}{\lambda_p + \sum_x \lambda_x}$$

$$\lambda_p \approx -3.3 W / m^2 / K$$

4/(3.3-2.5) ~ 4.5K

4/(3.3-1.5) ~ 2.2K

Soden and Held (2006)

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 a change in the atmospheric circulation alone explain a larger warming in high latitudes?**

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

Why do we need to incorporate the dynamics into feedback analysis?

- **Atmospheric motions play a role in the climate response to the external forcing.**

- **Even for a global uniform SURFACE warming, local convection and surface evaporation and sensible heat fluxes would act to reduce surface warming while enhancing the atmospheric warming**

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

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

(Lu & Cai 2009; Cai & Lu 2009)

- **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.**

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

Lu and Cai (2008) and Cai and Lu (2008)

Unperturbed climate state

$$\underbrace{(\vec{\mathbf{S}}-\vec{\mathbf{R}})}_{\textit{net}\ \text{rad. cooling/heating}} + \underbrace{\vec{\mathbf{Q}}}_{\textit{non-radiative}\ \text{dyn. heating/cooling}} = \frac{\partial \vec{E}}{\partial t}$$

Perturbation in response an external forcing

$$\underbrace{\Delta(\vec{\mathbf{S}}-\vec{\mathbf{R}})}_{\textit{change}\ \text{in}\ \textit{net}\ \text{rad. cooling/heating}\ (\mathrm{F}^{2CO_2}\ \text{included})}$$

$$+ \underbrace{\Delta\vec{\mathbf{Q}}}_{\textit{change}\ \text{in}\ \textit{non-radiative}\ \text{dyn. heating/cooling}} = \underbrace{\Delta\vec{E}}_{\textit{Heat Storage}}$$

Mathematical formulation of CFRAM

$$\left(\frac{\partial \vec{\overline{\mathbf{R}}}}{\partial \vec{\overline{\mathbf{T}}}}\right)\Delta\vec{\overline{\mathbf{T}}}^{tot} = \{\vec{\overline{\mathbf{F}}}^{ext} + \underbrace{\Delta^{(\alpha)}\vec{\overline{\mathbf{S}}} + \Delta^{(c)}(\vec{\overline{\mathbf{S}}} - \vec{\overline{\mathbf{R}}}) + \Delta^{(w)}(\vec{\overline{\mathbf{S}}} - \vec{\overline{\mathbf{R}}})}_{non\_temp\_induced\_radiative\_energy}$$

$$+ \underbrace{\Delta\vec{\overline{\mathbf{Q}}}}_{non-radiative\_energy} - \underbrace{\Delta\vec{E}}_{Heat\ \ Storage}\ \}$$

The radiation flux change only due to a change in the atmosphere-surface column temperature **=** **Radiative energy input due to the external forcing** **+**

Radiative and non-radiative energy flux perturbations that are not due to the radiation change associated with temperature changes and external forcing

$$\left(\frac{\partial \vec{\overline{\mathbf{R}}}}{\partial \vec{\overline{\mathbf{T}}}}\right)$$ **Planck feedback matrix**

Mathematical formulation of CFRAM

$$\Delta\overrightarrow{\mathbf{T}}^{tot} = \left(\frac{\partial\overrightarrow{\mathbf{R}}}{\partial\overrightarrow{\mathbf{T}}}\right)^{-1}\{\overrightarrow{\mathbf{F}}^{2CO_2} + \Delta^{(\alpha)}\overrightarrow{\mathbf{S}} + \Delta^{(c)}(\overrightarrow{\mathbf{S}}-\overrightarrow{\mathbf{R}}) + \Delta^{(w)}(\overrightarrow{\mathbf{S}}-\overrightarrow{\mathbf{R}})$$

$$+\Delta\overrightarrow{\mathbf{Q}} - \Delta\vec{E}\}$$

$$-\Delta^{total}(\vec{S}-\vec{R}) = \Delta\overrightarrow{\mathbf{Q}} - \Delta\vec{E} = \Delta\overrightarrow{\mathbf{Q}}^{evaporation} + \Delta\overrightarrow{\mathbf{Q}}^{surface\_sensibl\_heat+flux}$$

$$+(\Delta\overrightarrow{\mathbf{Q}}^{convection} + \Delta\overrightarrow{\mathbf{Q}}^{ATM\_\lg\_dyn}) + \Delta\overrightarrow{\mathbf{Q}}^{OCN\_dyn+storage}$$

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

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

Both feedbacks and their effects are additive!

Demonstration of the Dynamical PWA mechanism in a GCM without hydrological cycle

(Lu and Cai 2010; Cai and Tung 2012)

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 and evaporation feedbacks, and poleward latent heat transport in a GCM model?

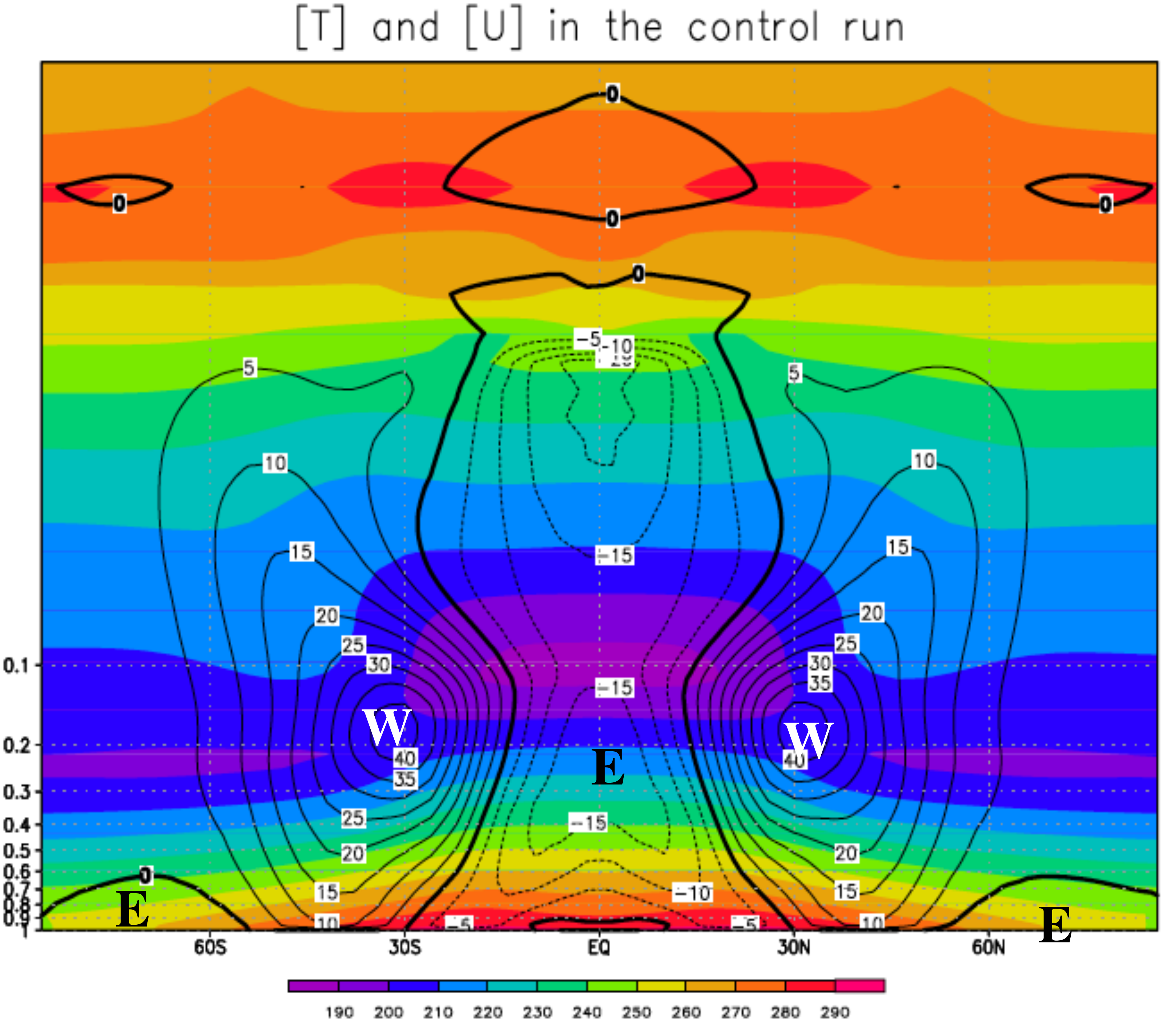
The key features of the coupled GCM model

Dynamical core: Suarez and Held (1992)

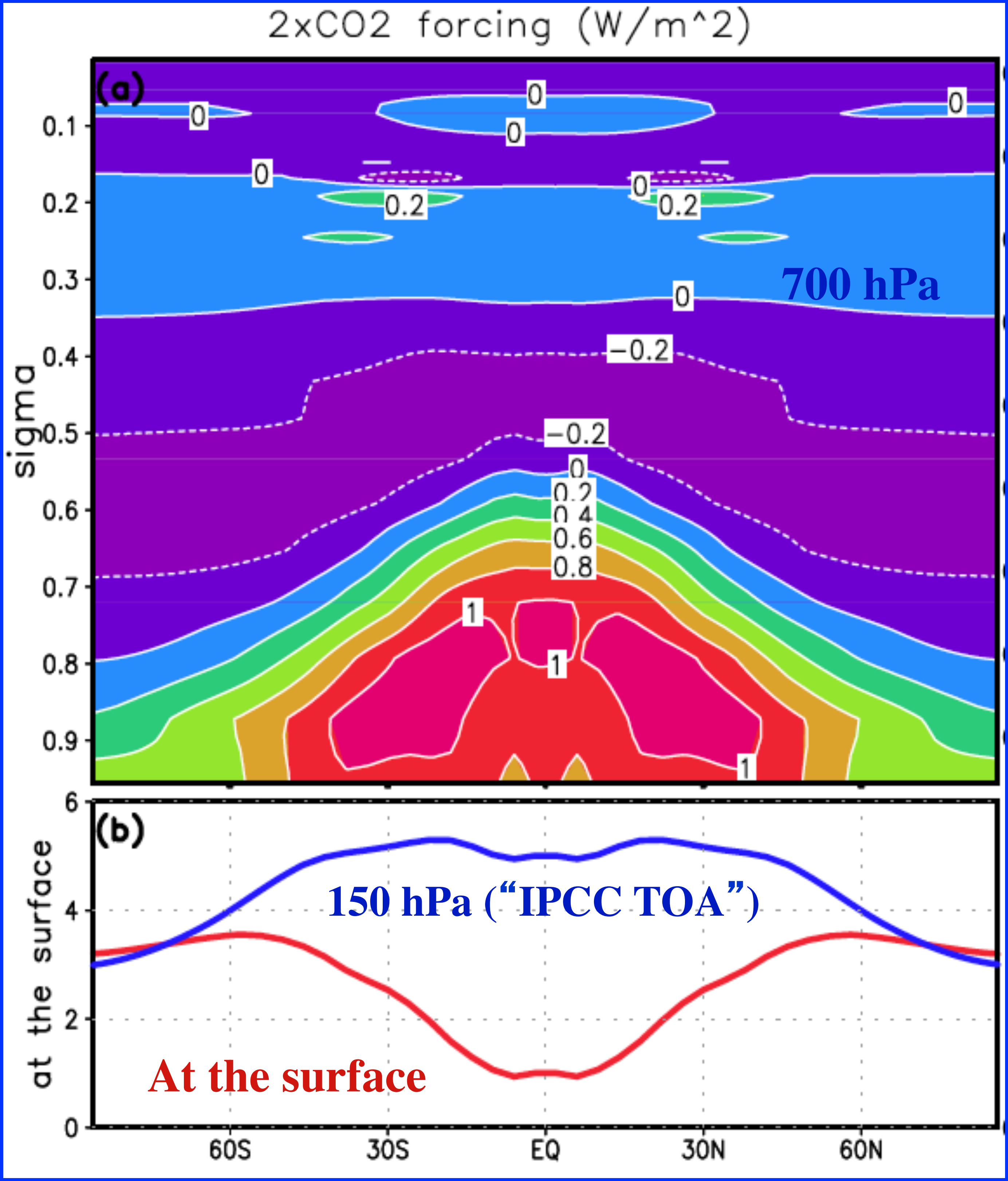
Physics:

- **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 prescribed 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)

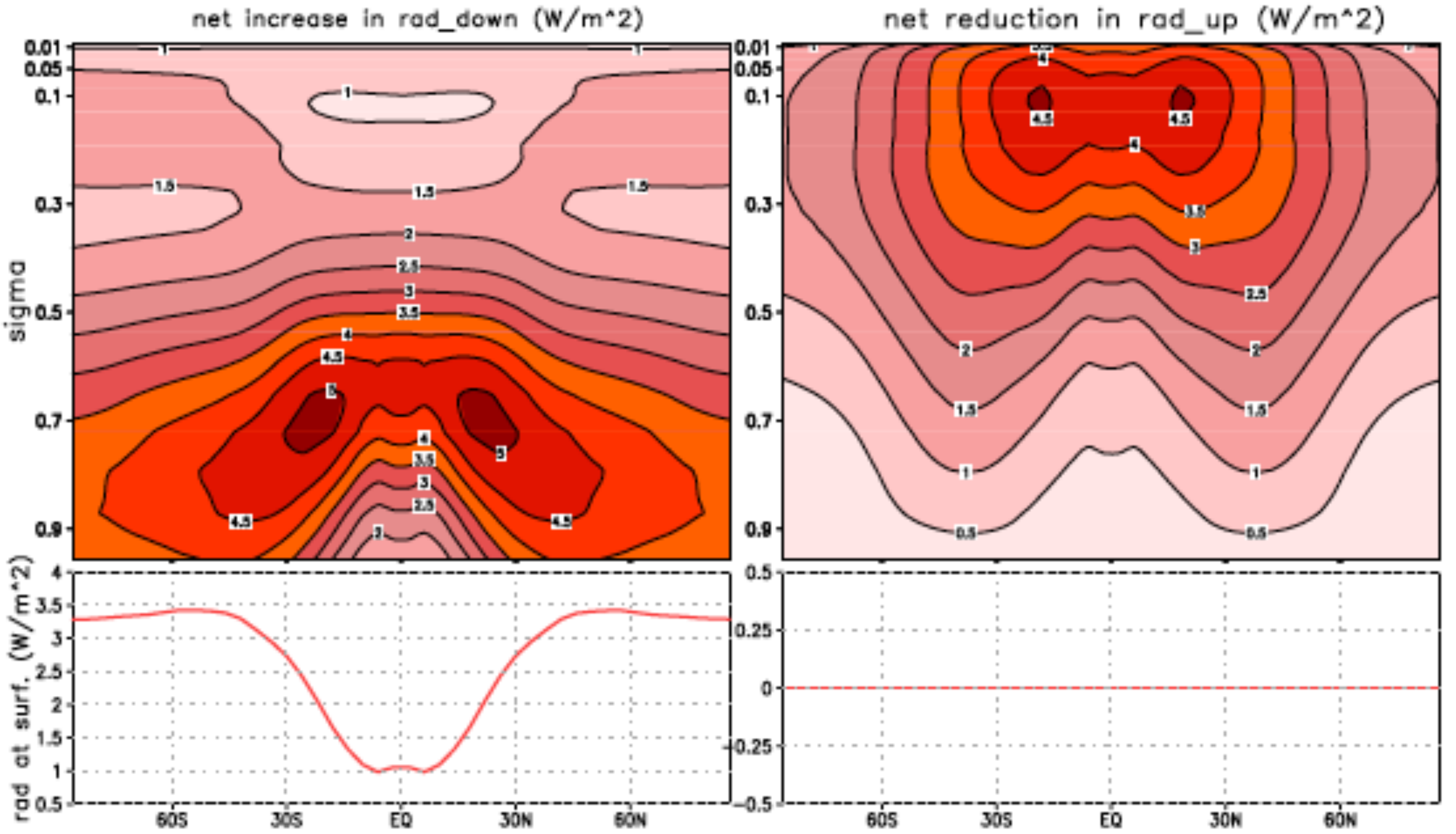


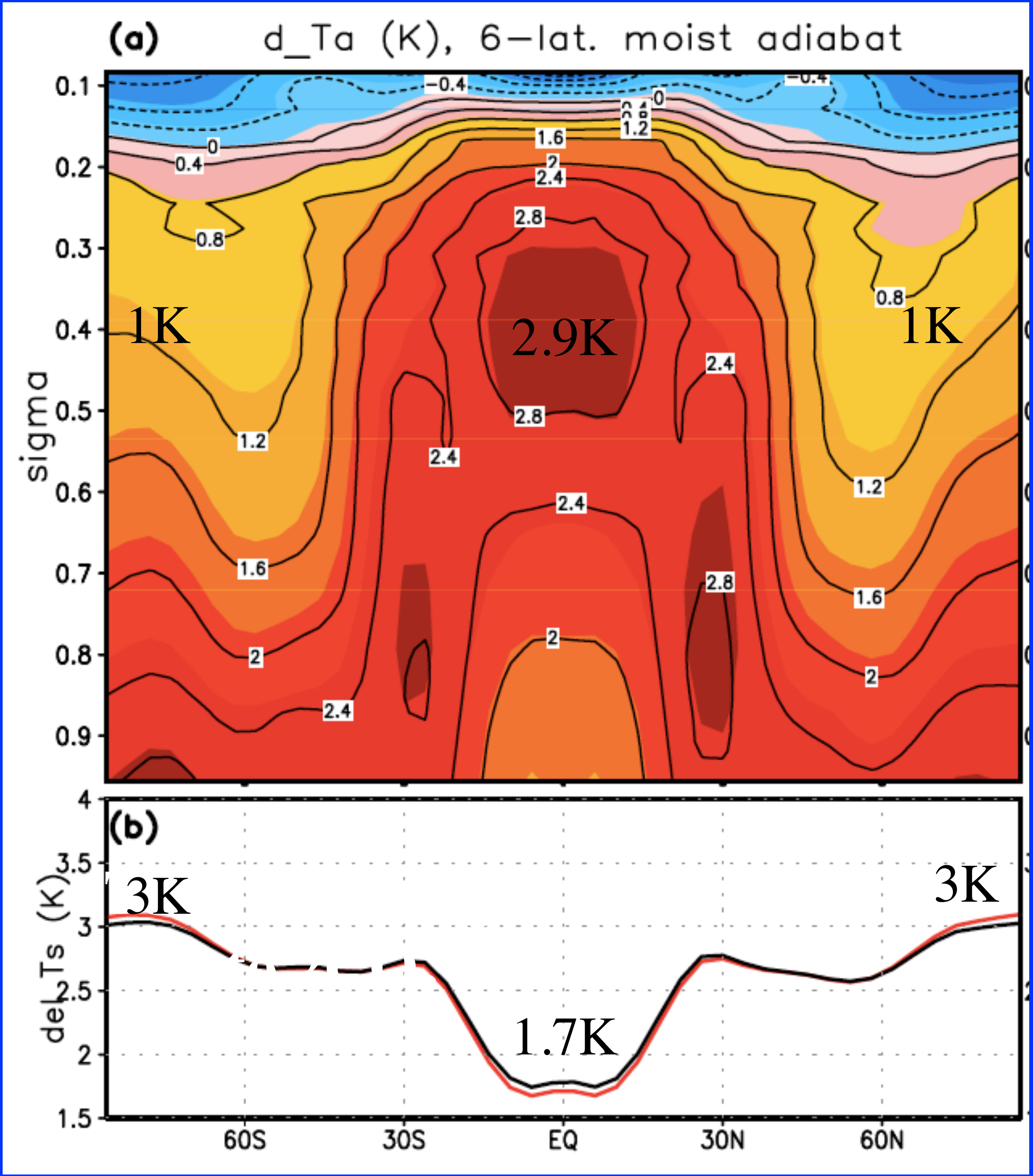
700 hPa

150 hPa ("IPCC TOA")

At the surface

Changes in upward and downward LW radiative energy flux due 2xCO2

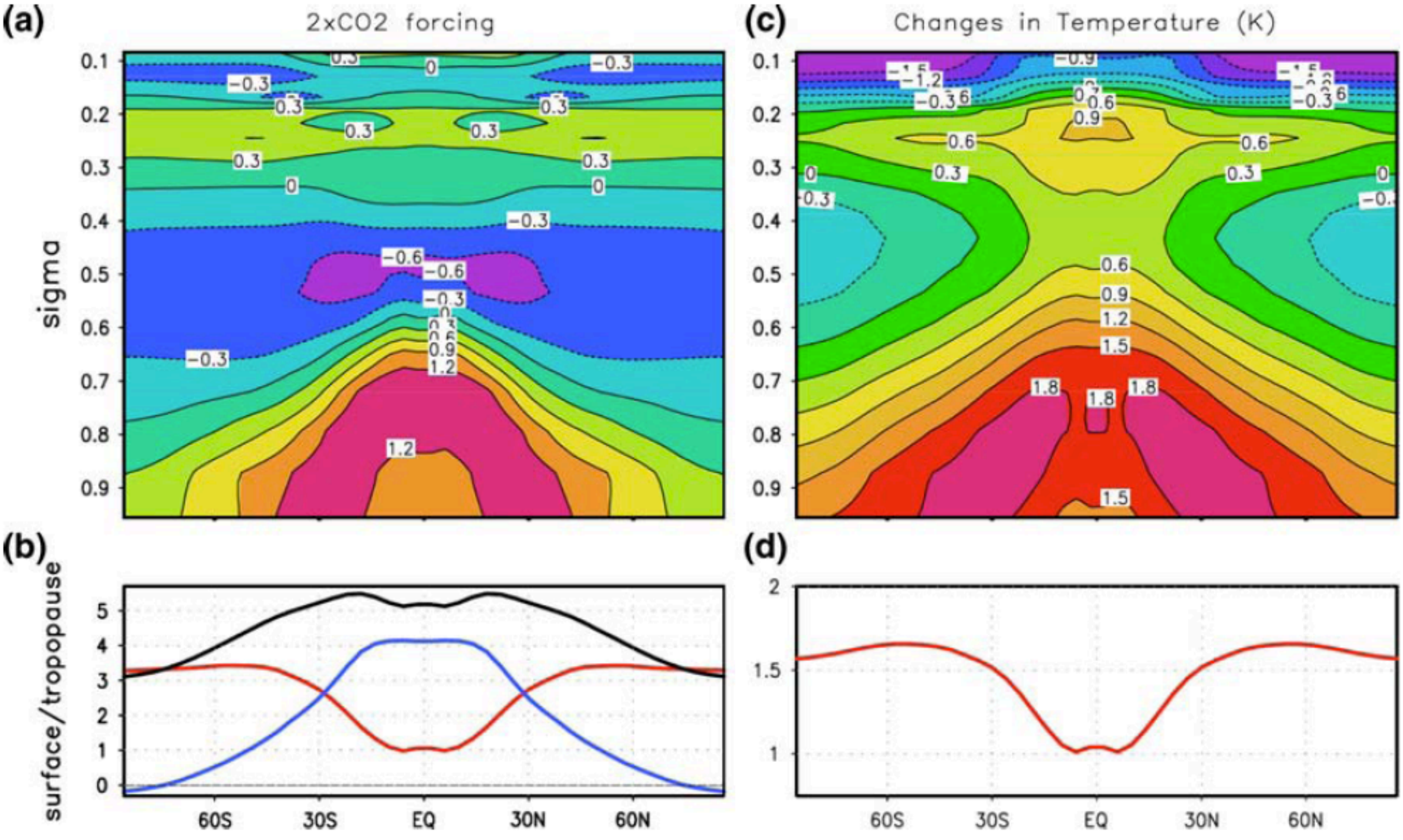




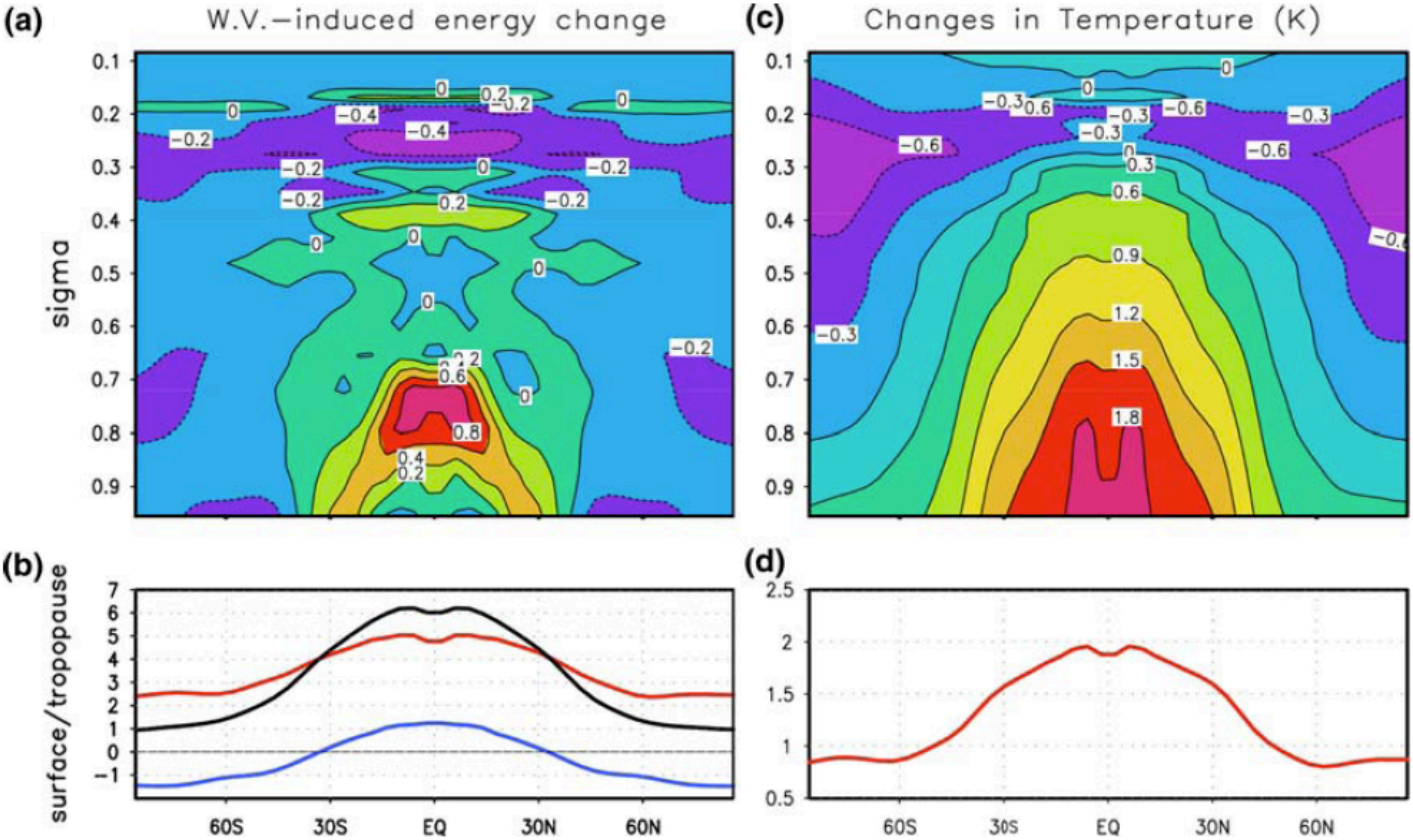
Total Warming Pattern

1. In the atmosphere: ΔT(trop) > ΔT(polar)
2. At the surface: ΔT(trop) > ΔT(polar)
3. In the tropics: ΔT_surf < ΔT_air)
4. In polar region: ΔT_surf > ΔT_air)

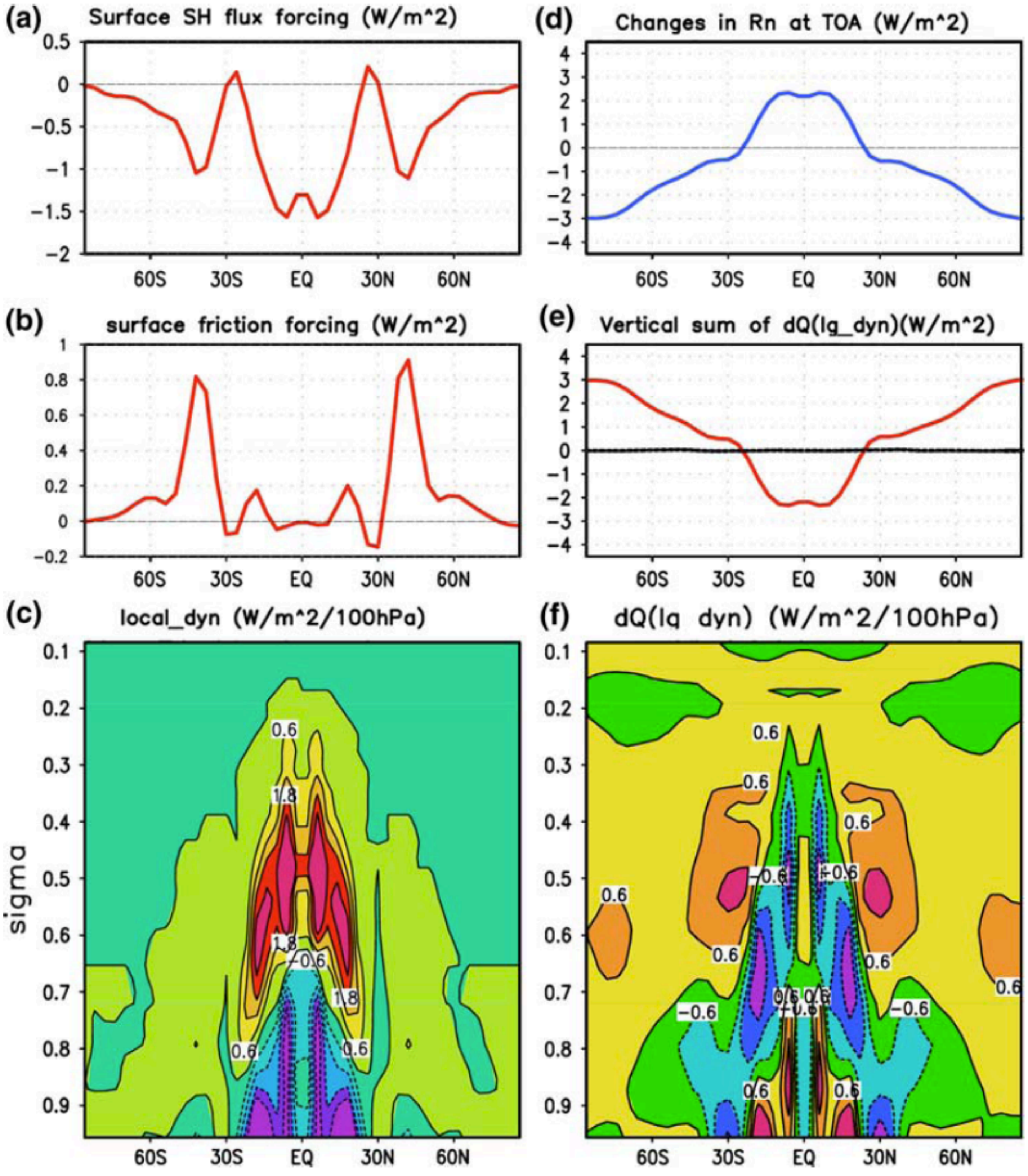
Temp. Changes due to 2xCO2 Alone



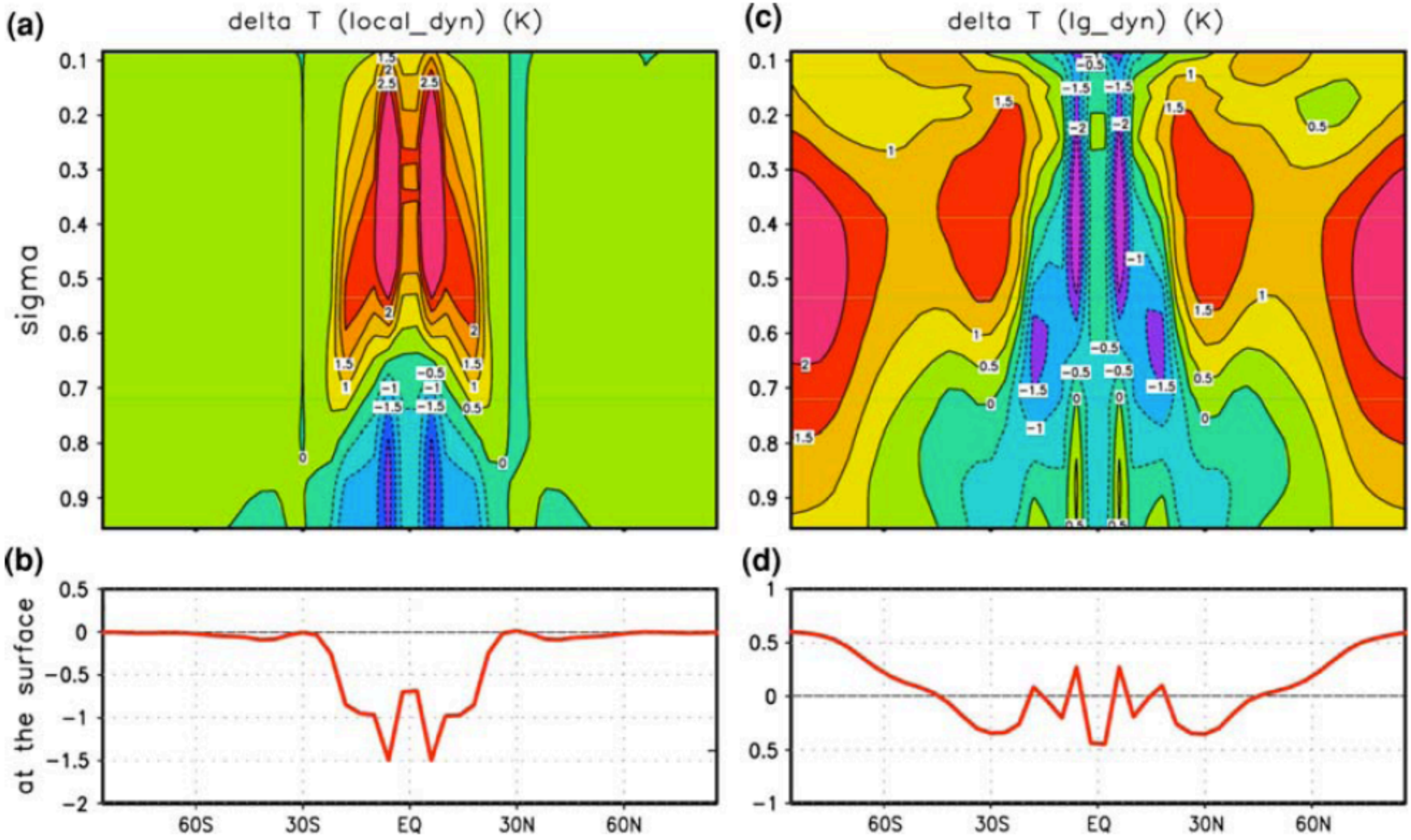
Temp. Changes due to WV feedback Alone



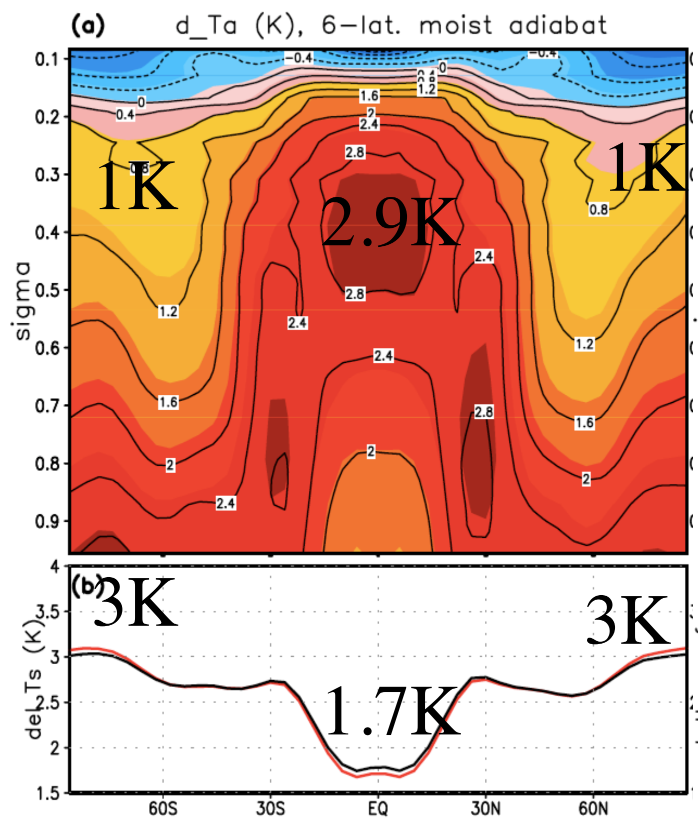
Changes in non-radiative energy fluxes



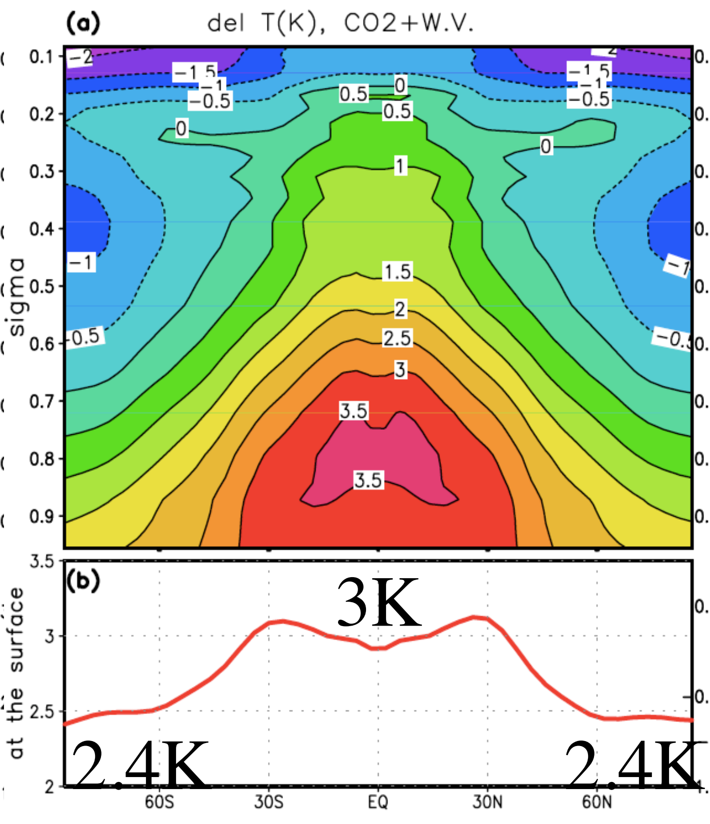
Temp. Changes due to non-radiative conv. & large-scale dyn. feedbacks



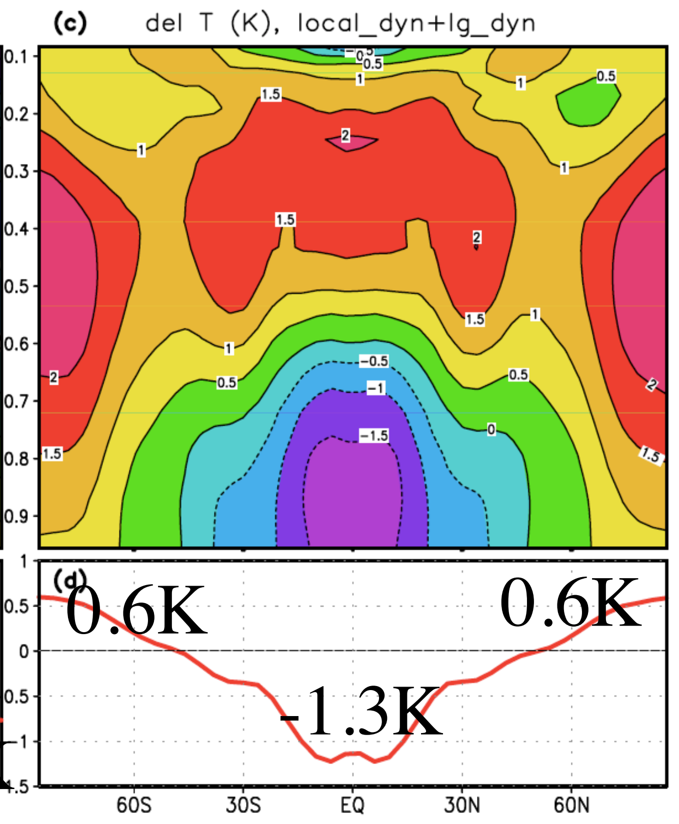
Total warming and Sum of partial ΔTs



Total warming due to $2CO_2$



2CO2 + H2O) feedbacks



Convective + Poleward energy tranport feedbacks

Summary of "dry" GCM results

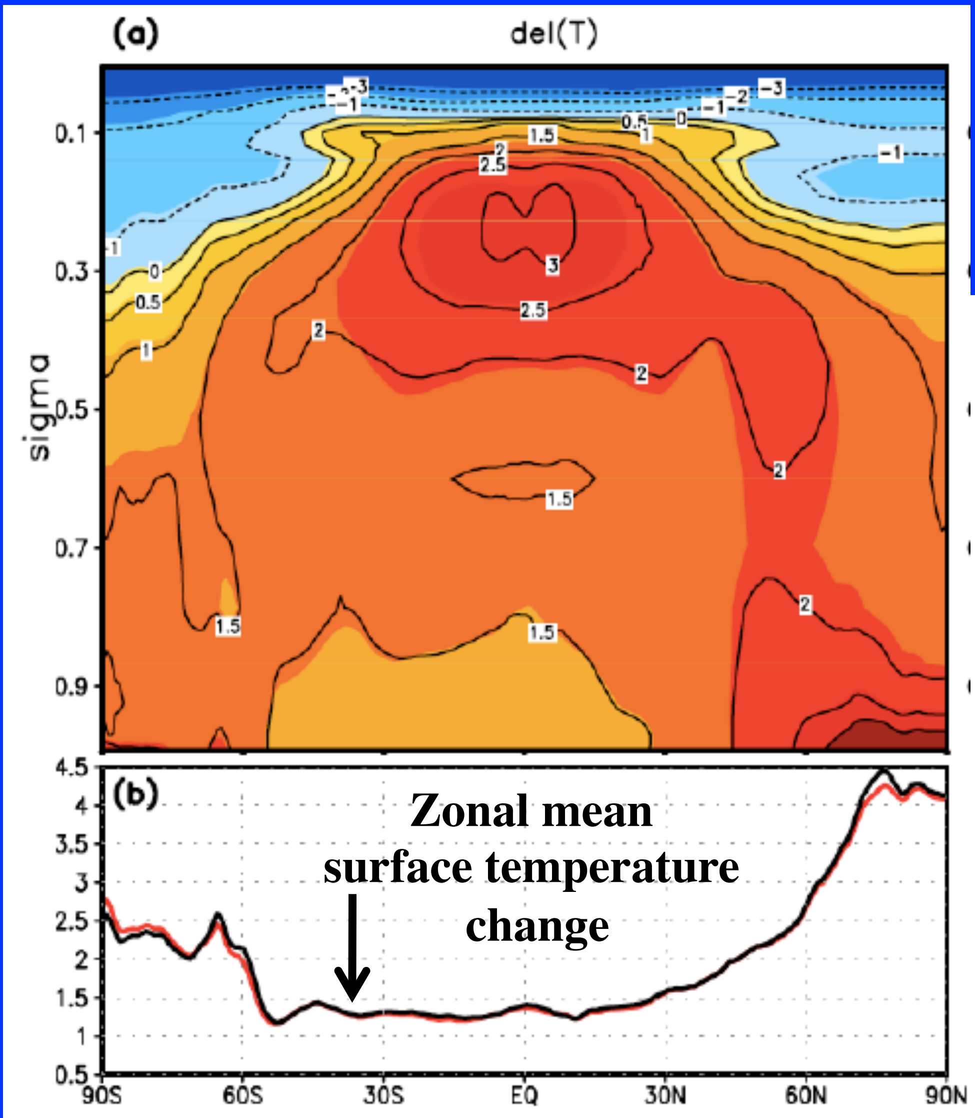
- **2CO2 forcings exhibits a poleward and vertical decreasing profile in the atmosphere.**
- **The water vapor feedback strengthens the upward decreasing radiative heating profile in the tropics and the poleward decreasing radiative heating profile in the lower troposphere.**
- **The convective feedback plays an important role only in the tropics where they act to reduce the warming at the surface and lower troposphere in favor of upper troposphere warming.**
- **The large-scale dynamical poleward energy transport is enhanced in both cases, contributing to a polar amplification of warming aloft and a warming reduction in the tropics.**
- **The dynamical amplification of polar atmospheric warming also contributes additional warming to the surface below via downward thermal radiation.**

NCAR NCAR CCSM4 Climate Simulations

(Taylor et al. 2013, J. of Climate)

Temperature Response to 2xCO_2 Forcing in NCAR CCSM4 Climate Simulations

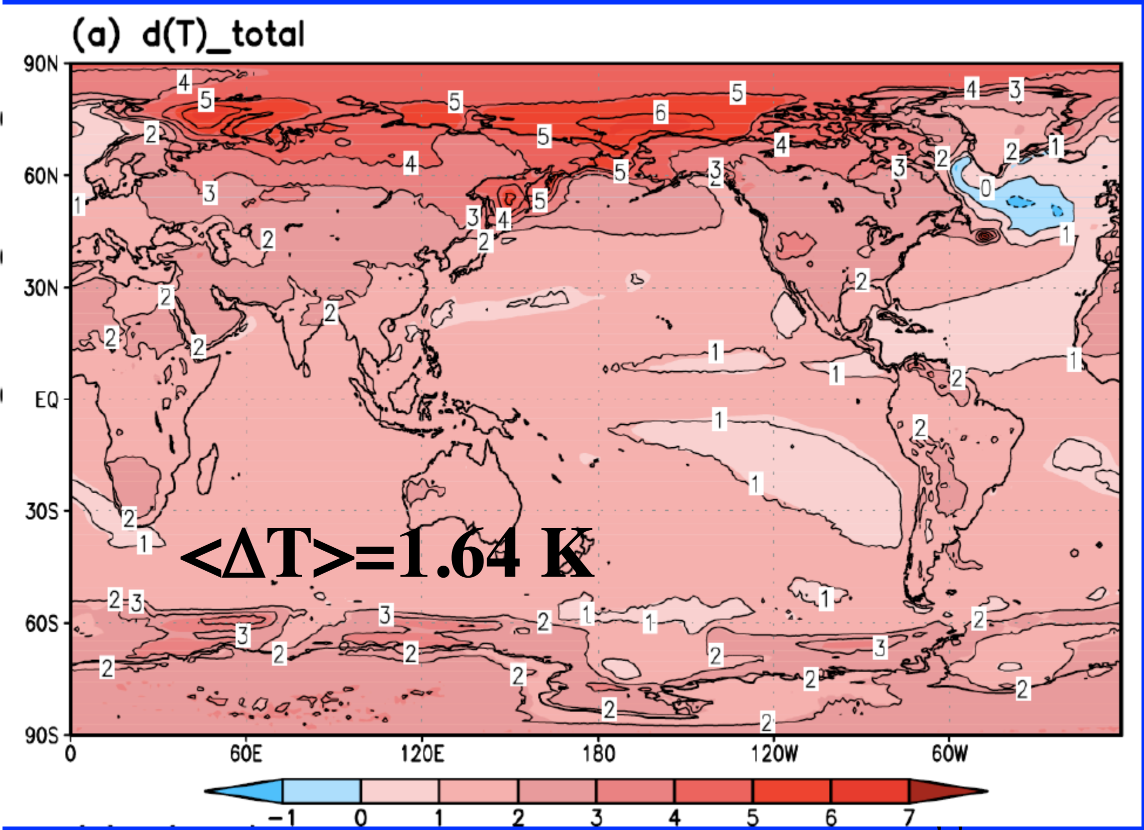
(at the time of doubling of CO_2 from its pre-industry level 284.7 PPM)



Zonal mean surface temperature change

Zonal mean air temperature change

surface



<ΔT>=1.64 K

Questions

- **How much warming is just due to the doubling of CO_2 alone?**

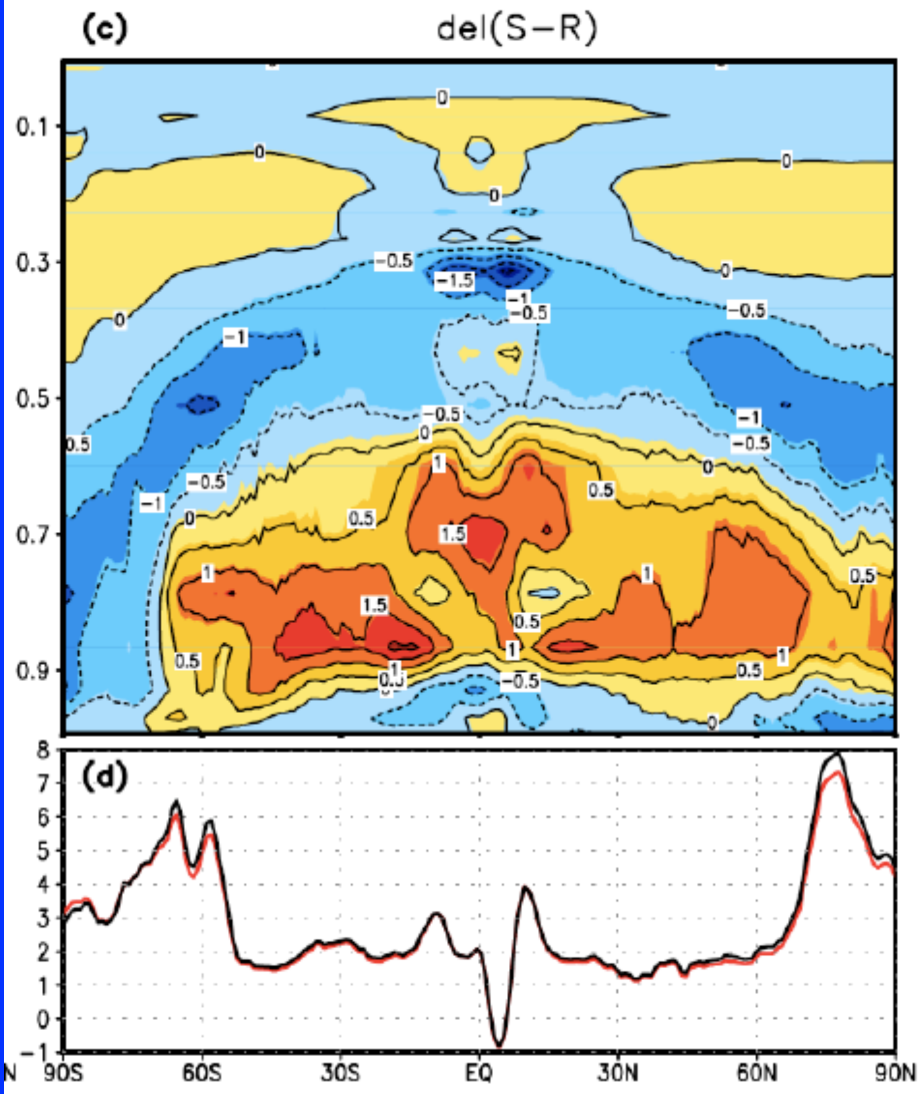
- **What are the additional temperature changes due to various radiative and non-radiative feedback processes?**

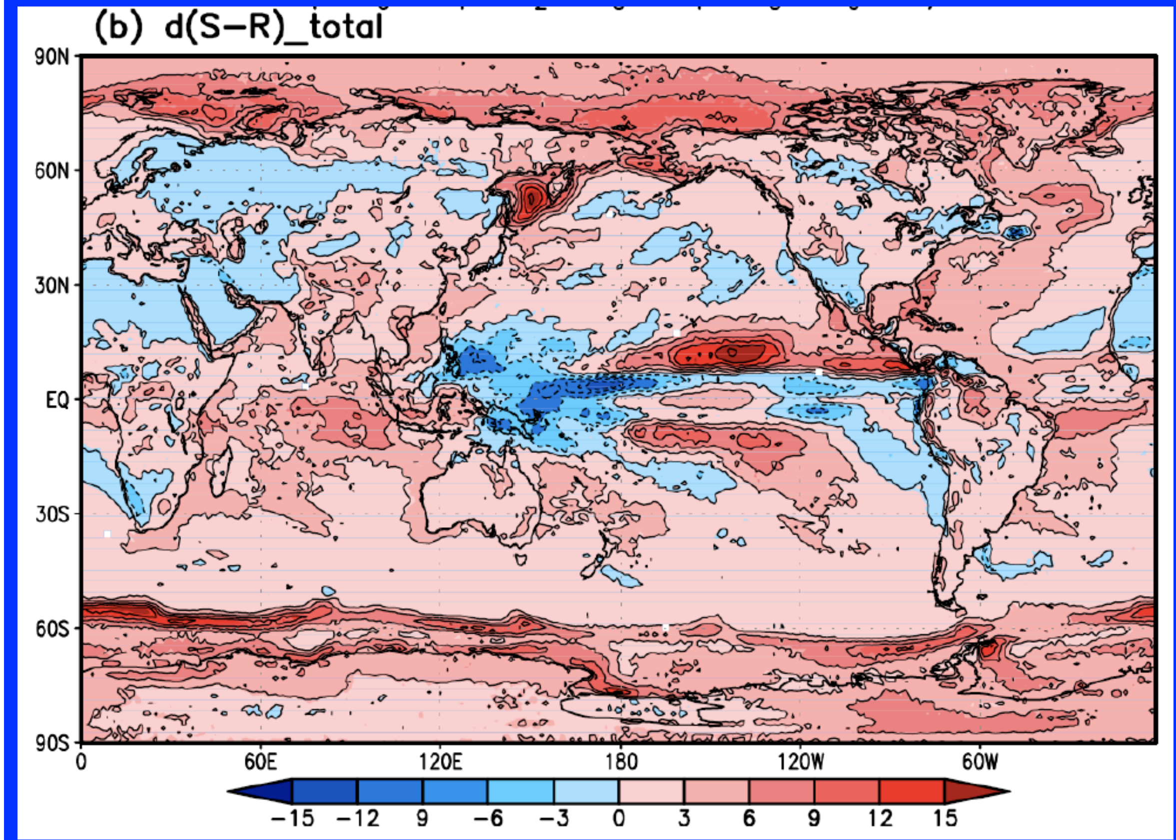
- **What are their contributions to the final warming pattern?**

- **What are the main processes contributing the polar warming amplification?**

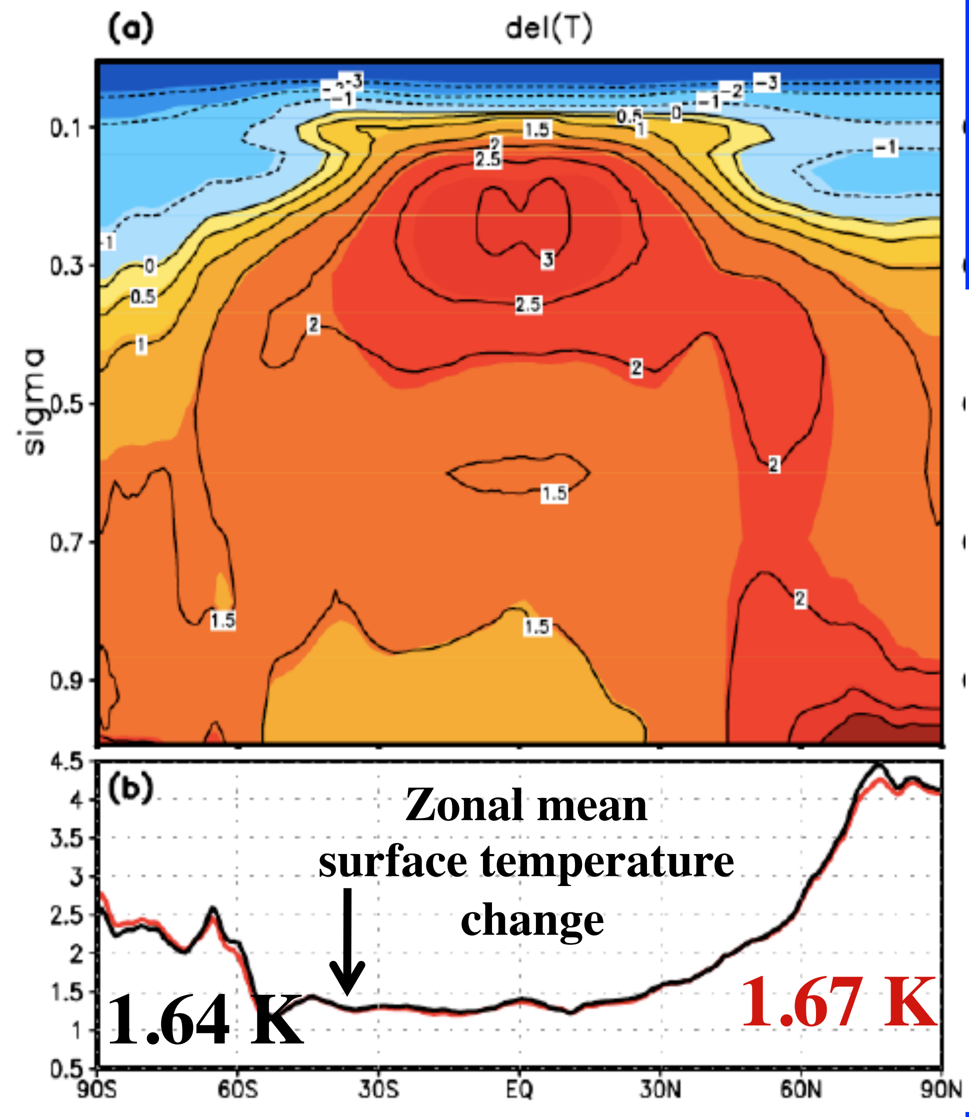
Validation of Linearization

$$\underbrace{\Delta^{(tot)}(\vec{\mathbf{S}}-\vec{\mathbf{R}})}_{Shadings} \simeq \underbrace{\vec{\mathbf{F}}^{2CO_2} + \Delta^{(\alpha)}\vec{\mathbf{S}} + \Delta^{(c)}(\vec{\mathbf{S}}-\vec{\mathbf{R}}) + \Delta^{(w)}(\vec{\mathbf{S}}-\vec{\mathbf{R}}) - \left(\frac{\partial \vec{\mathbf{R}}}{\partial \vec{\mathbf{T}}}\right)\Delta\vec{\mathbf{T}}^{tot}}_{contours}$$





Validation of CFRAM: $\Delta T^{tot} = \sum_{n} \Delta T^{(n)}$



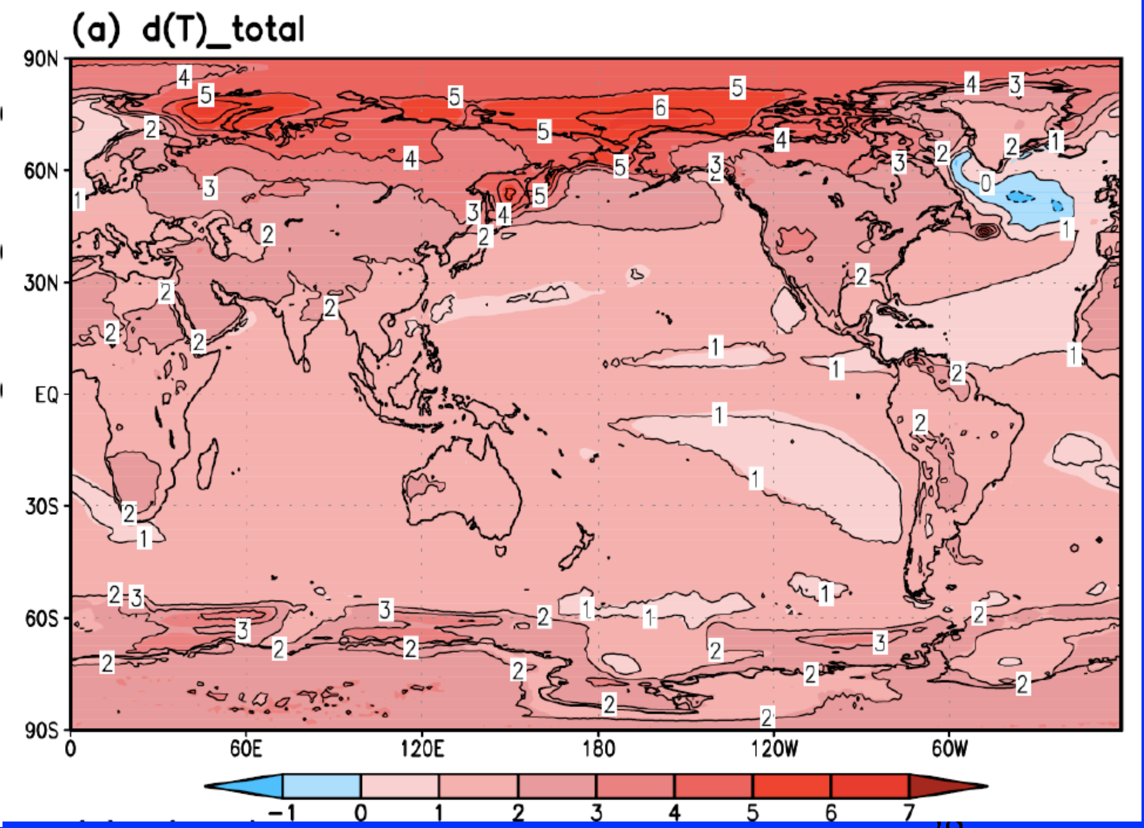
Zonal mean surface temperature change

1.64 K

1.67 K

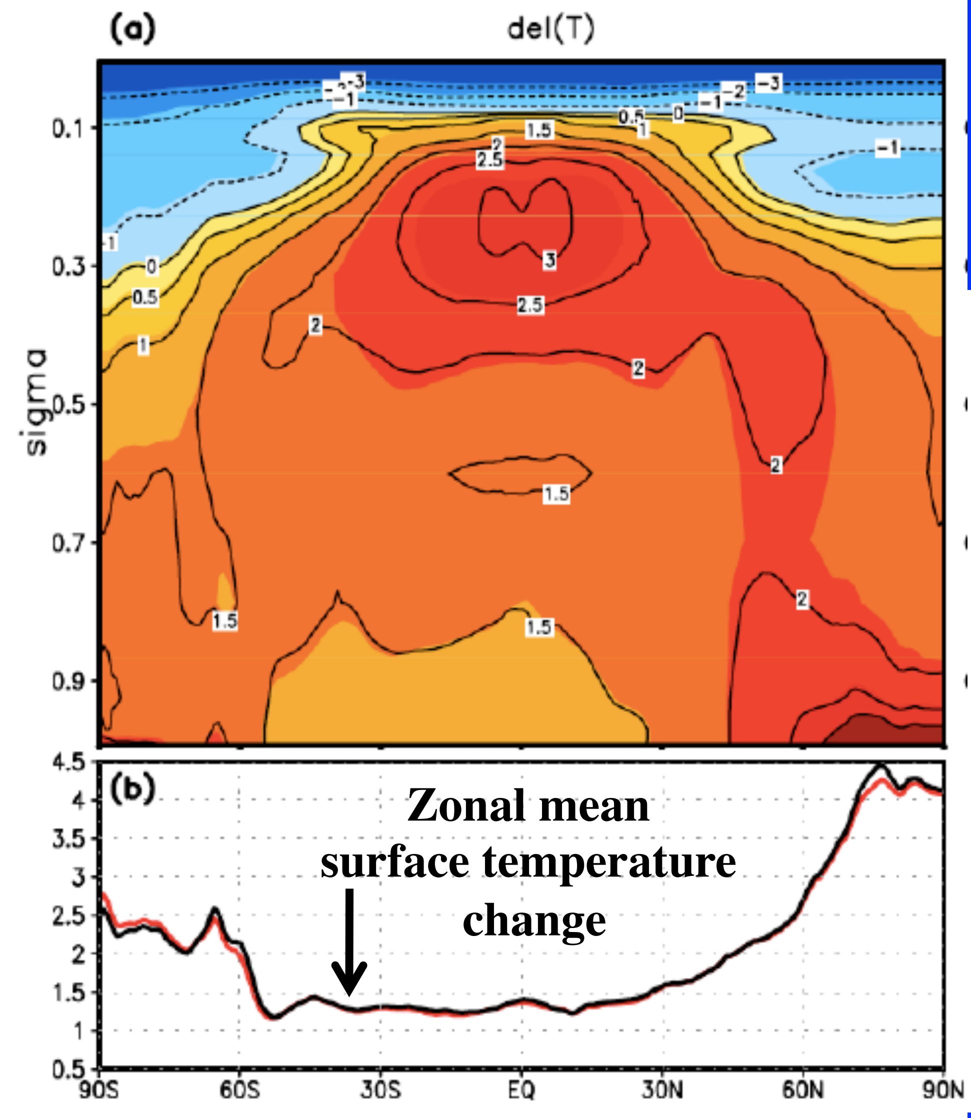
← Zonal mean air temperature change

surface ↓



Temperature Response to 2xCO_2 Forcing in NCAR CCSM4 Climate Simulations

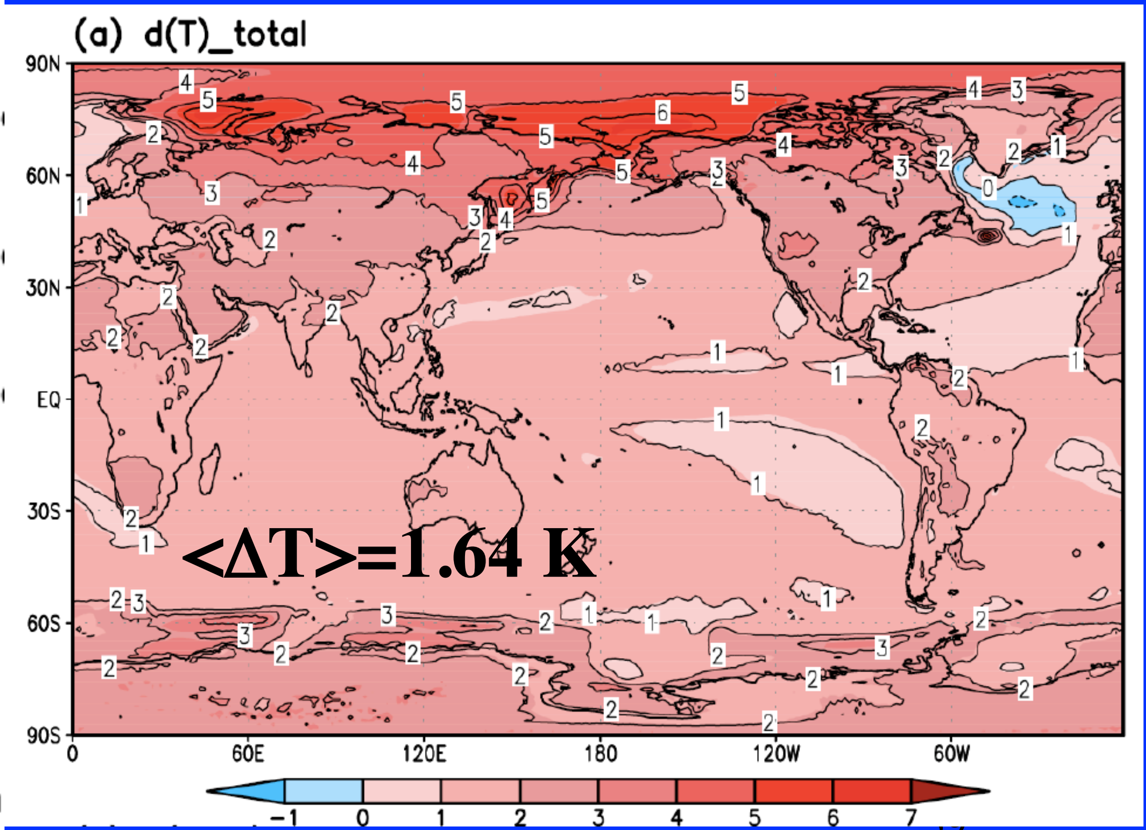
(at the time of doubling of CO_2 from its pre-industry level 284.7 PPM)



Zonal mean surface temperature change

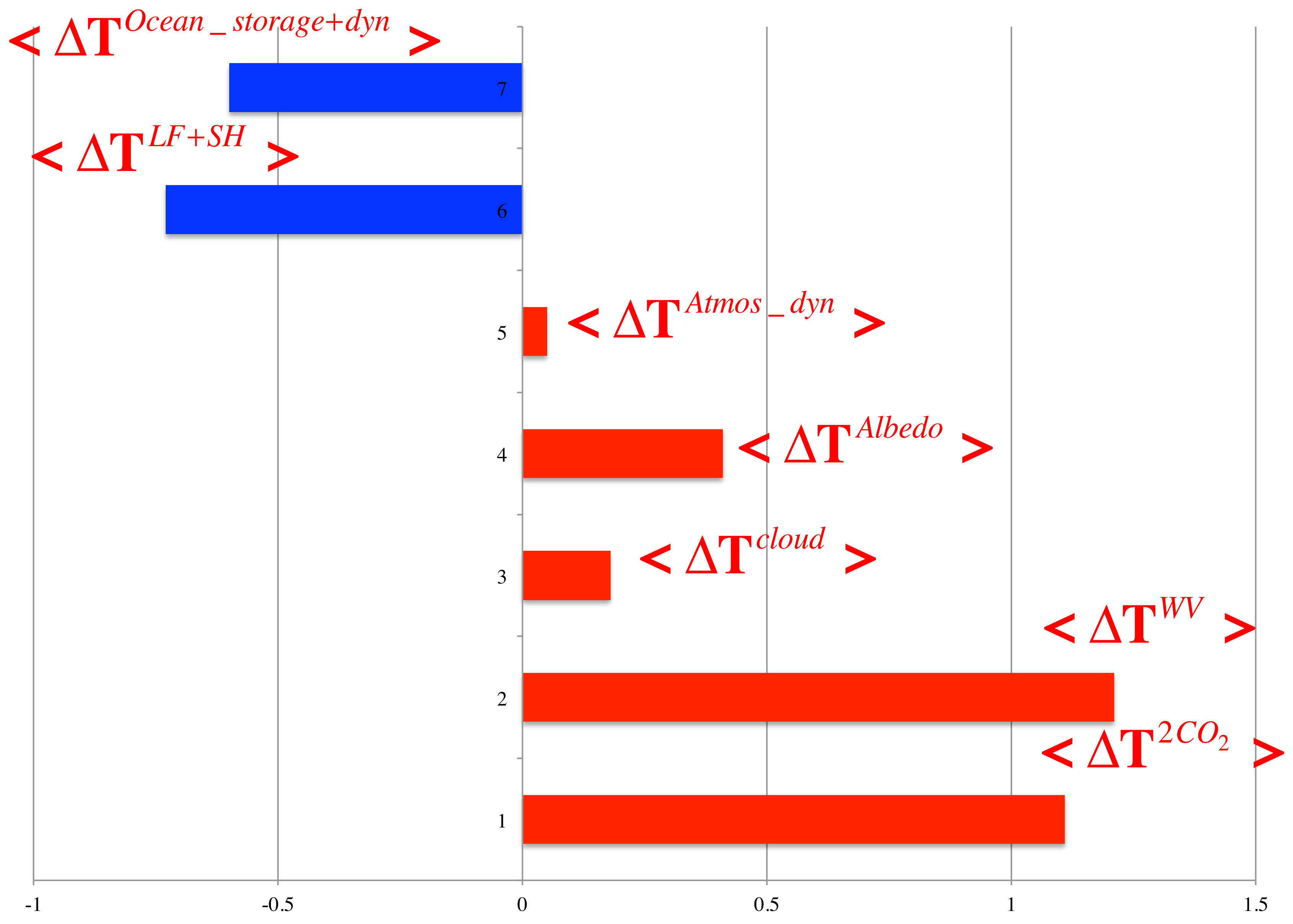
Zonal mean air temperature change

surface

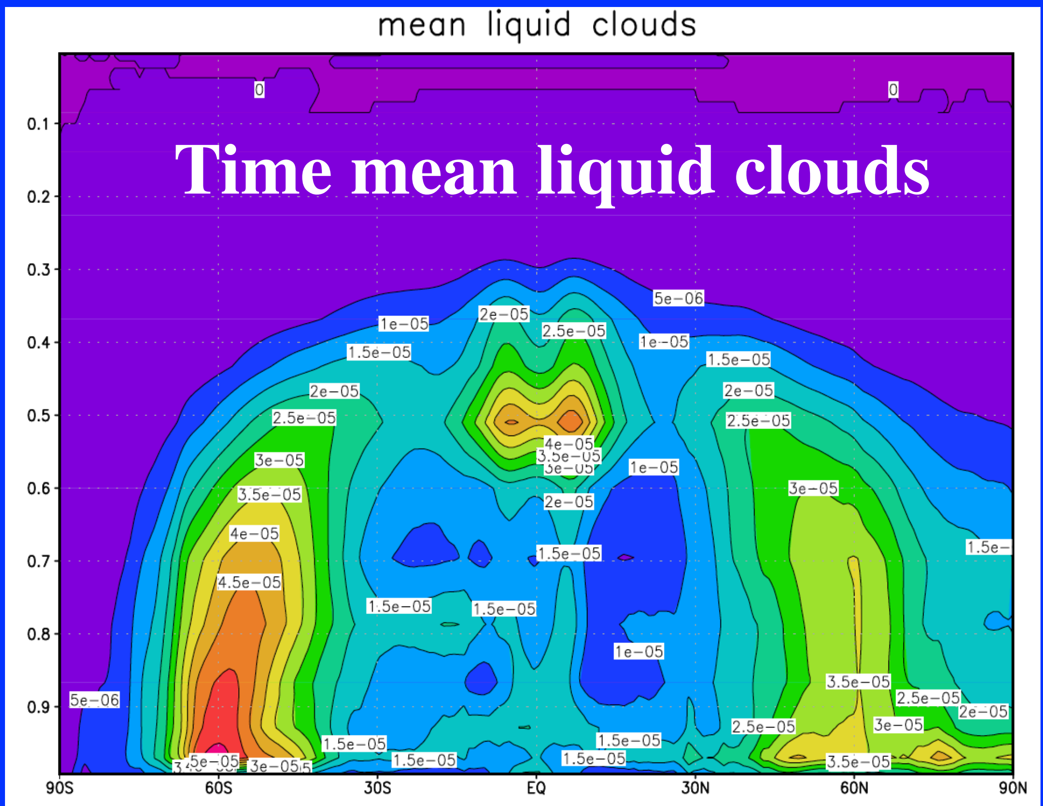


<ΔT>=1.64 K

Global Mean Surf. Temp. Changes

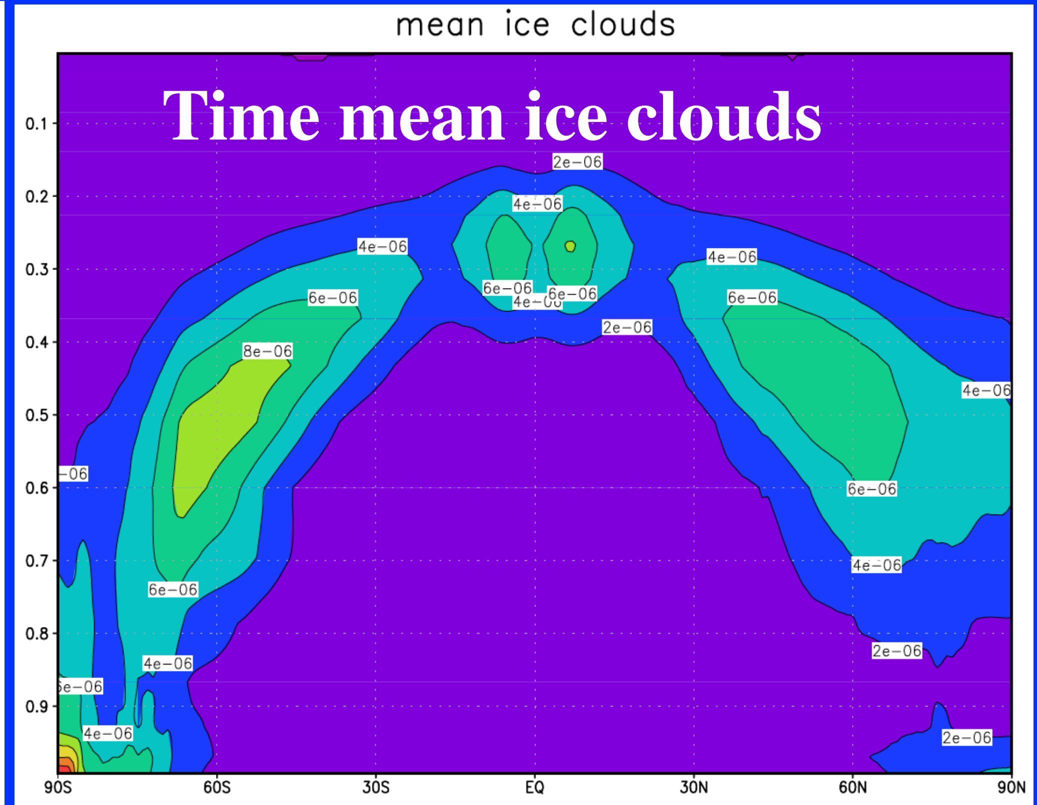


Global mean equilibrium response should be somewhat larger than 2.2K



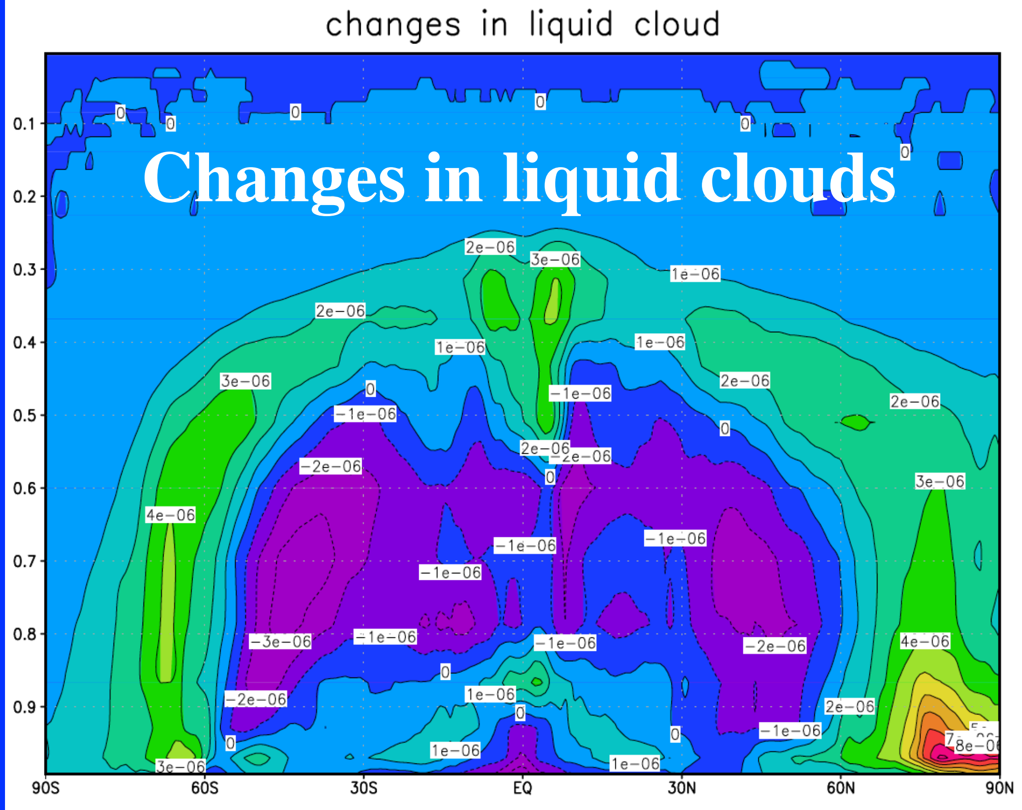
mean liquid clouds

Time mean liquid clouds



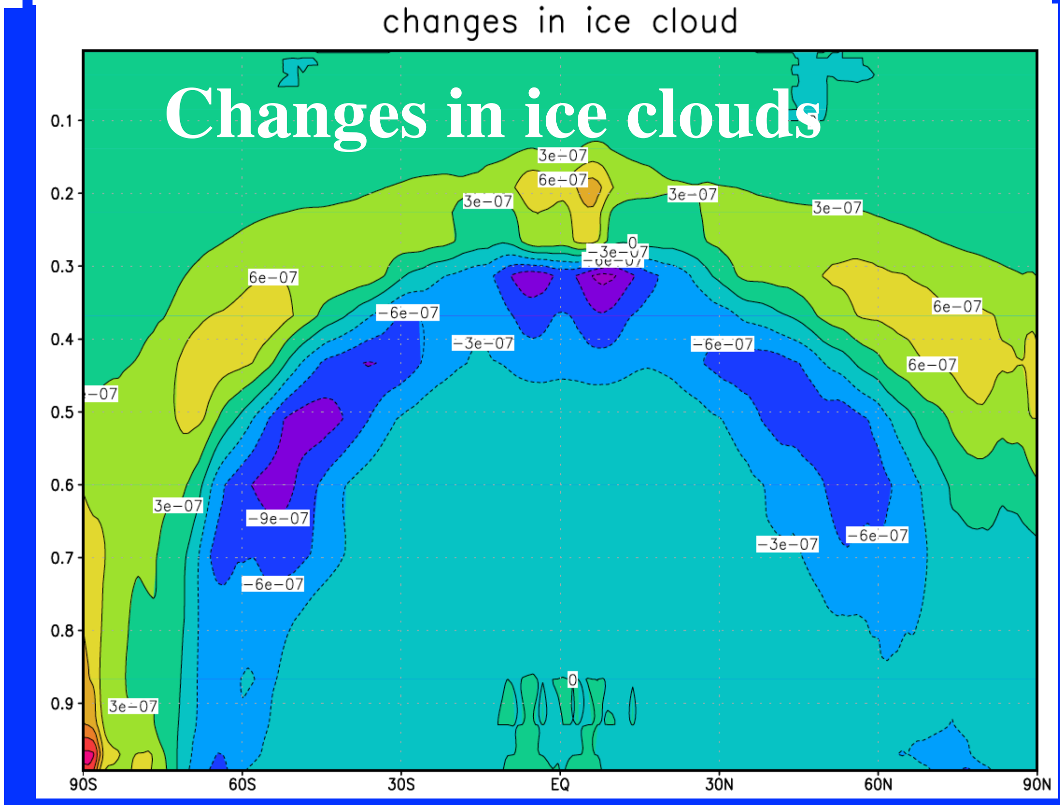
mean ice clouds

Time mean ice clouds



changes in liquid cloud

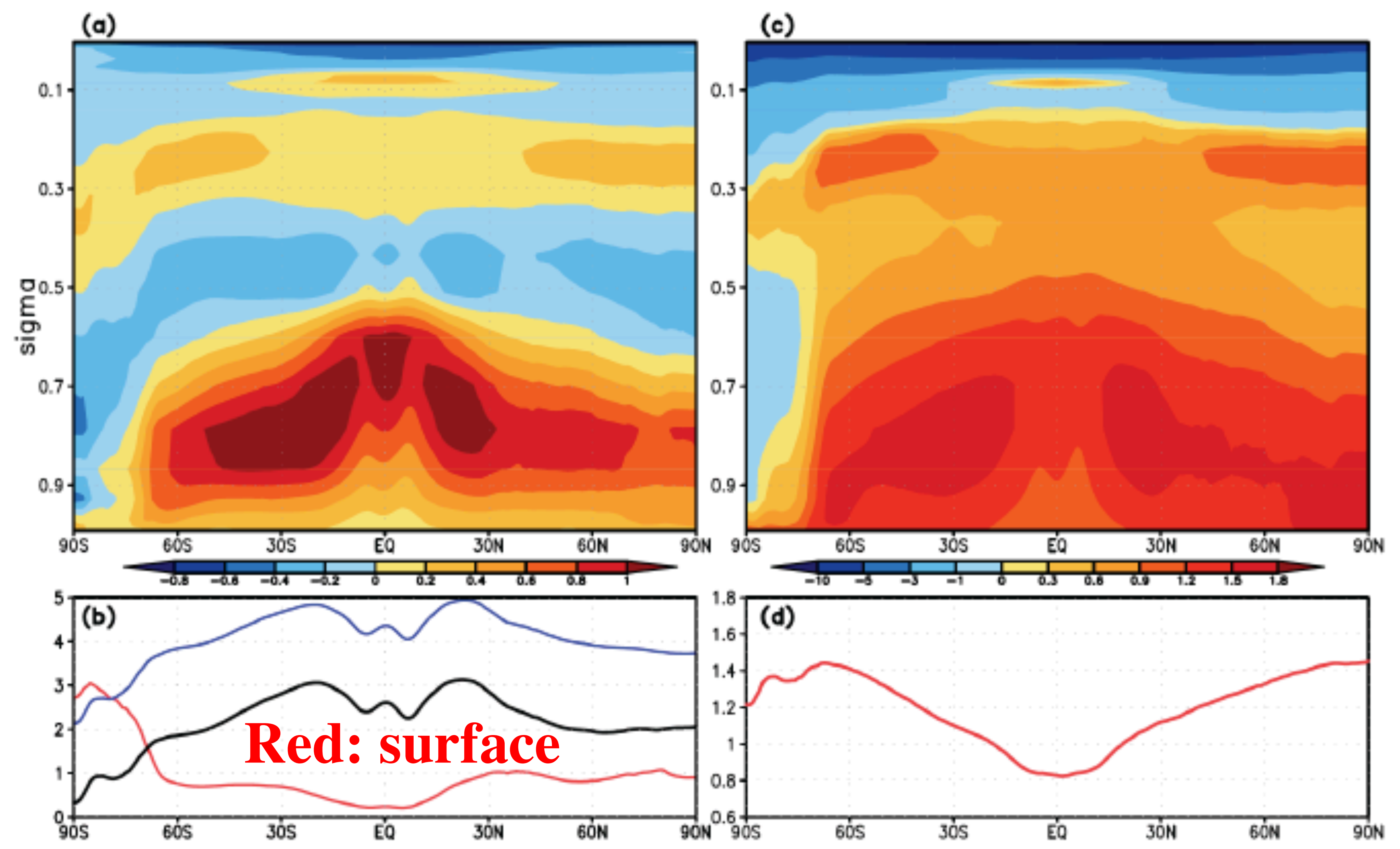
Changes in liquid clouds



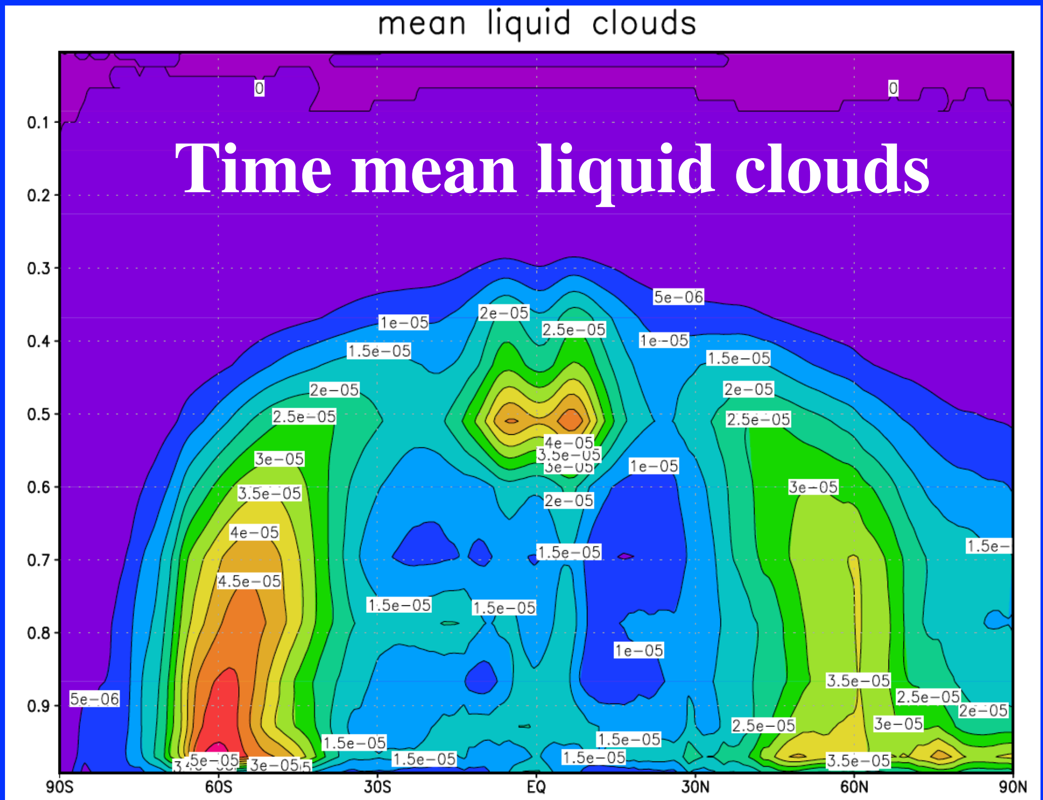
changes in ice cloud

Changes in ice clouds

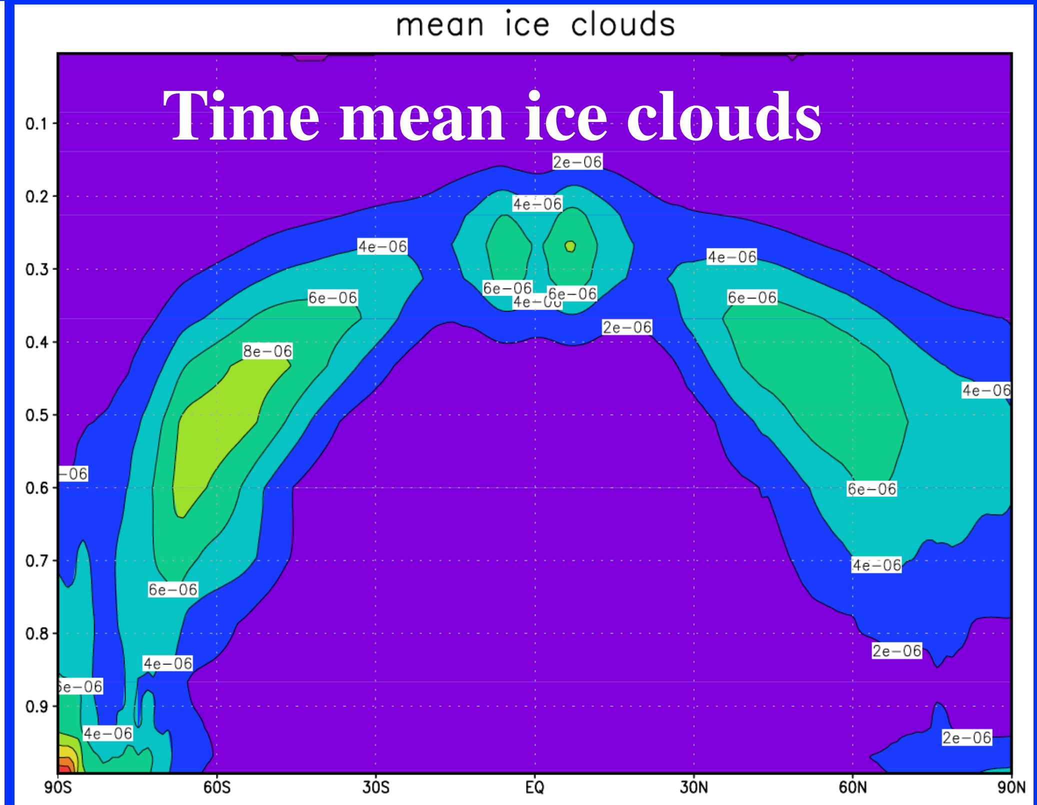
$[\Delta F^{(EXT)}]$ $[\Delta T^{(EXT)}]$



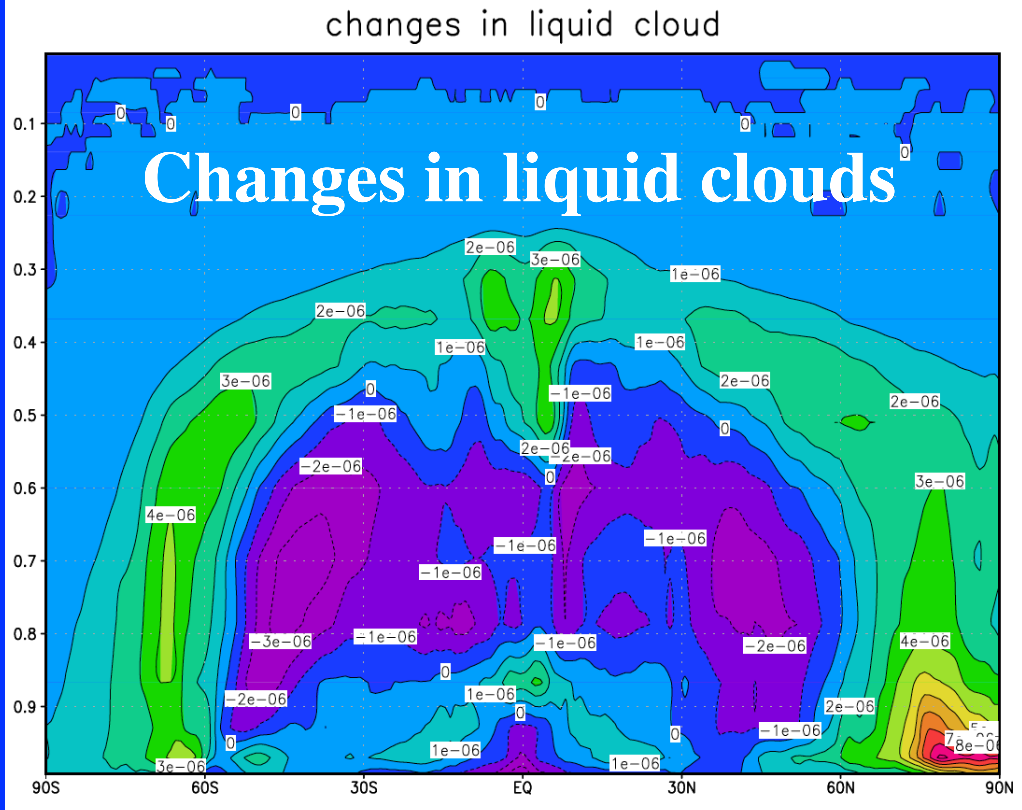
Red: surface


mean liquid clouds

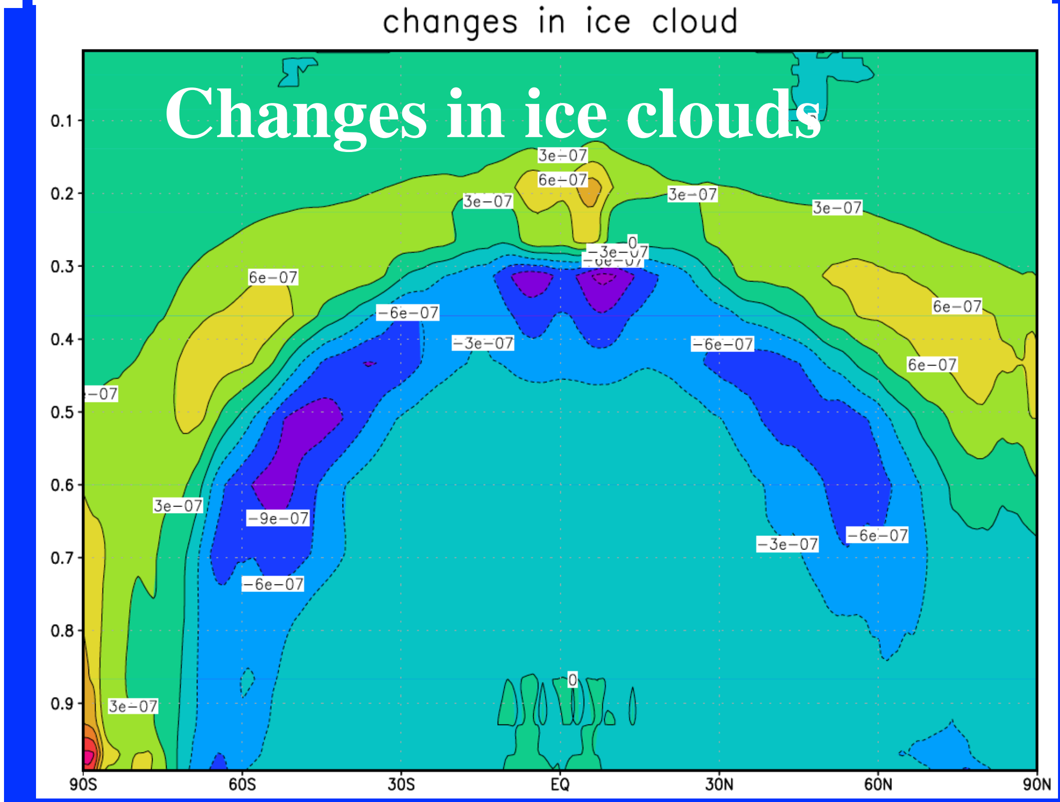
Time mean liquid clouds


mean ice clouds

Time mean ice clouds

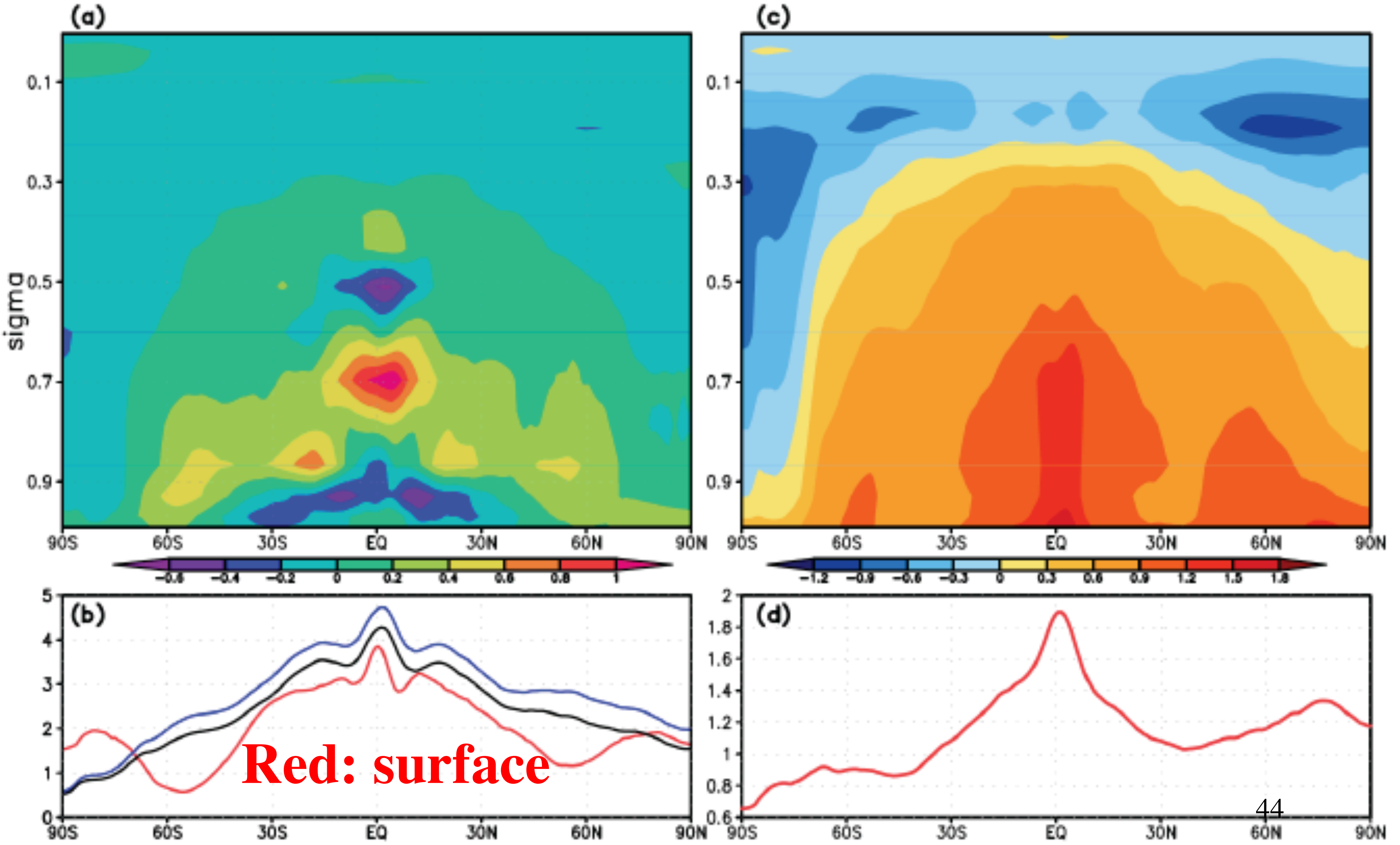

changes in liquid cloud

Changes in liquid clouds

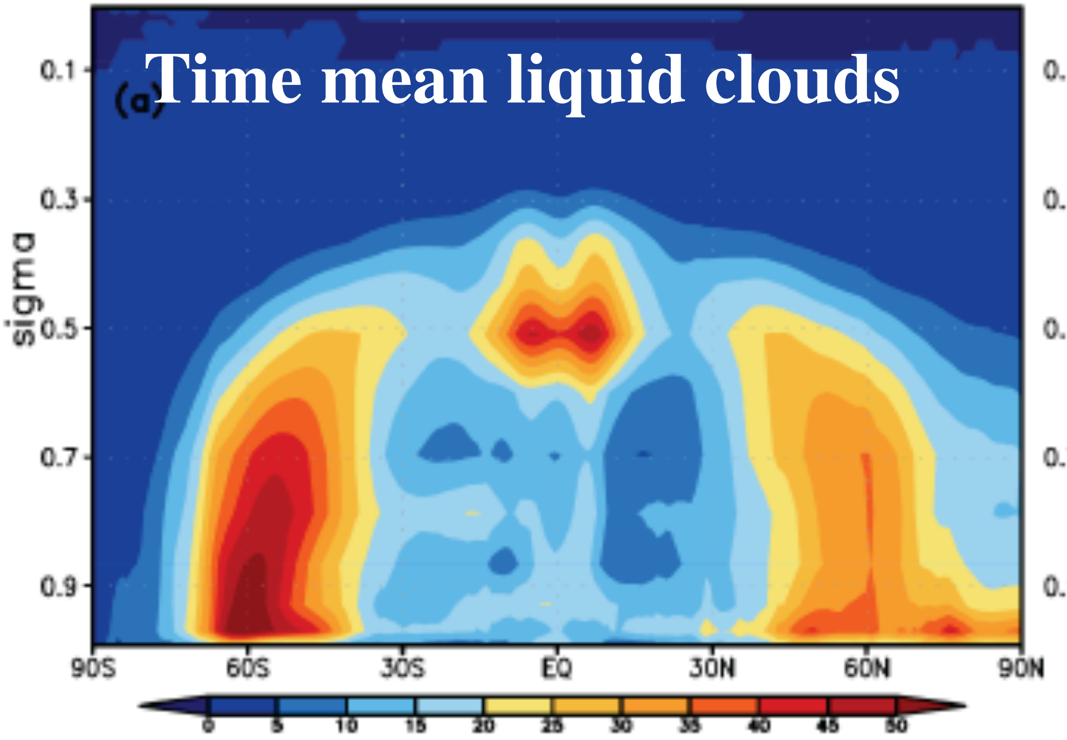

changes in ice cloud

Changes in ice clouds

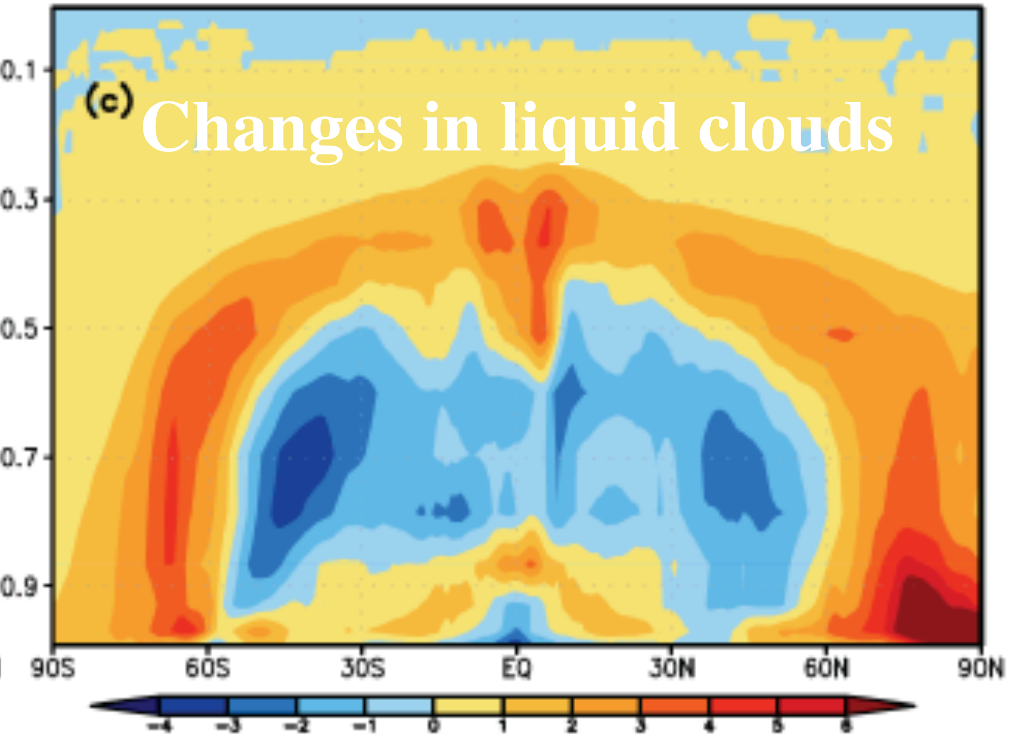
$[\Delta F^{(WV)}]$ $[\Delta T^{(WV)}]$



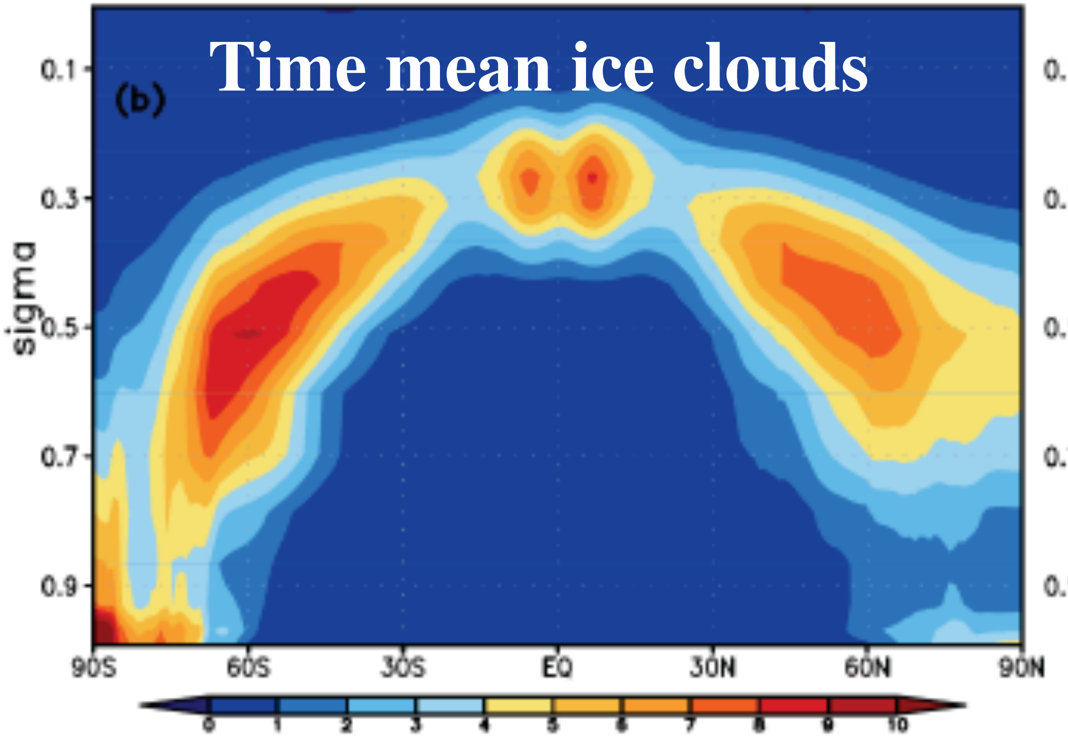
Red: surface



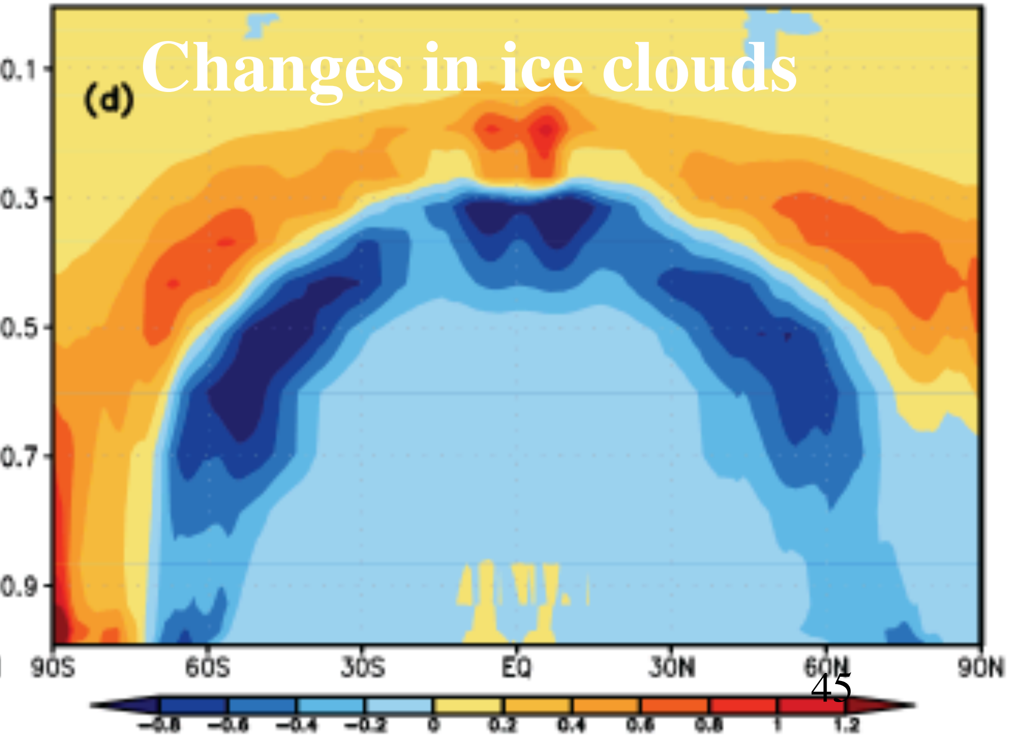
Time mean liquid clouds



Changes in liquid clouds



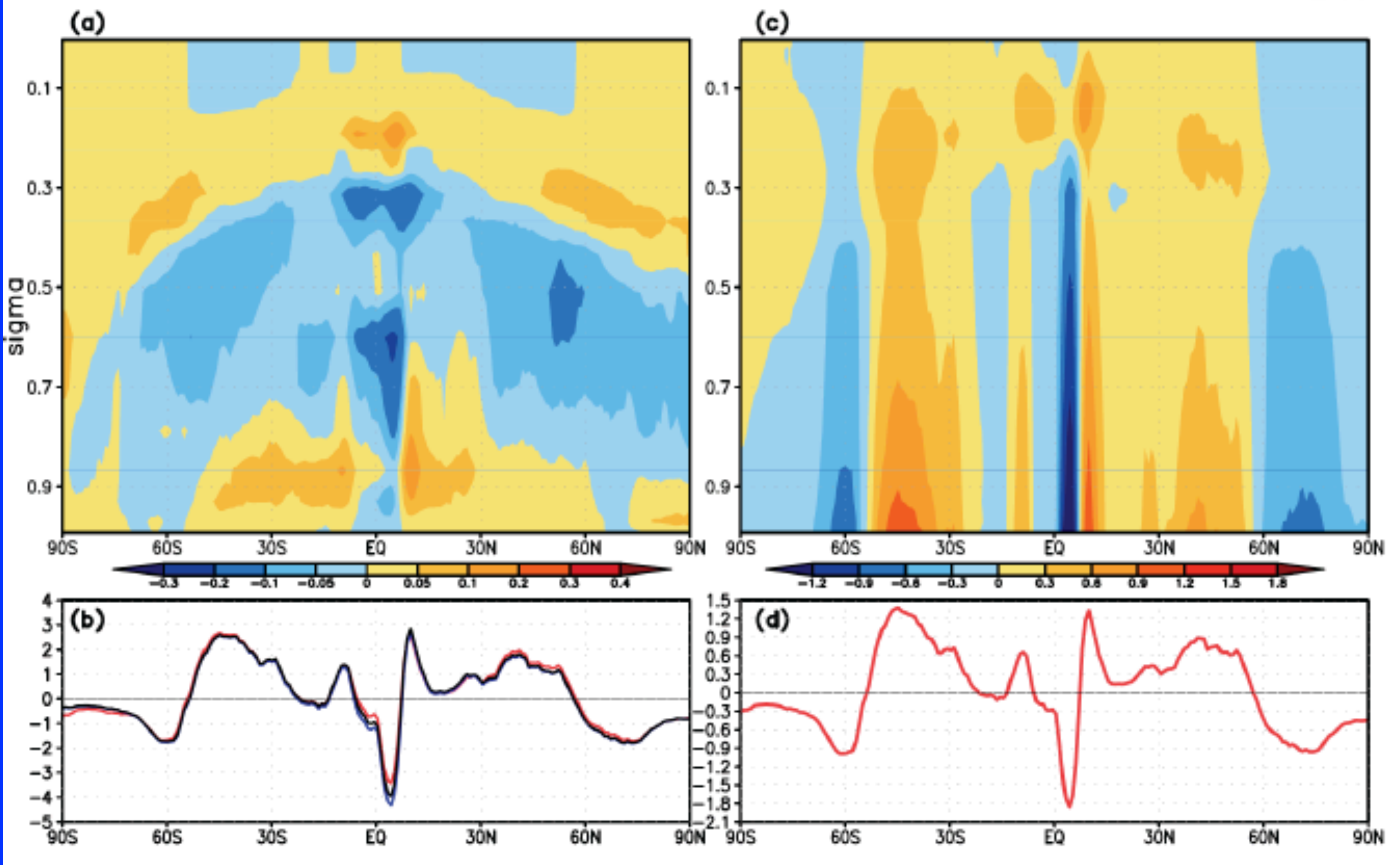
Time mean ice clouds



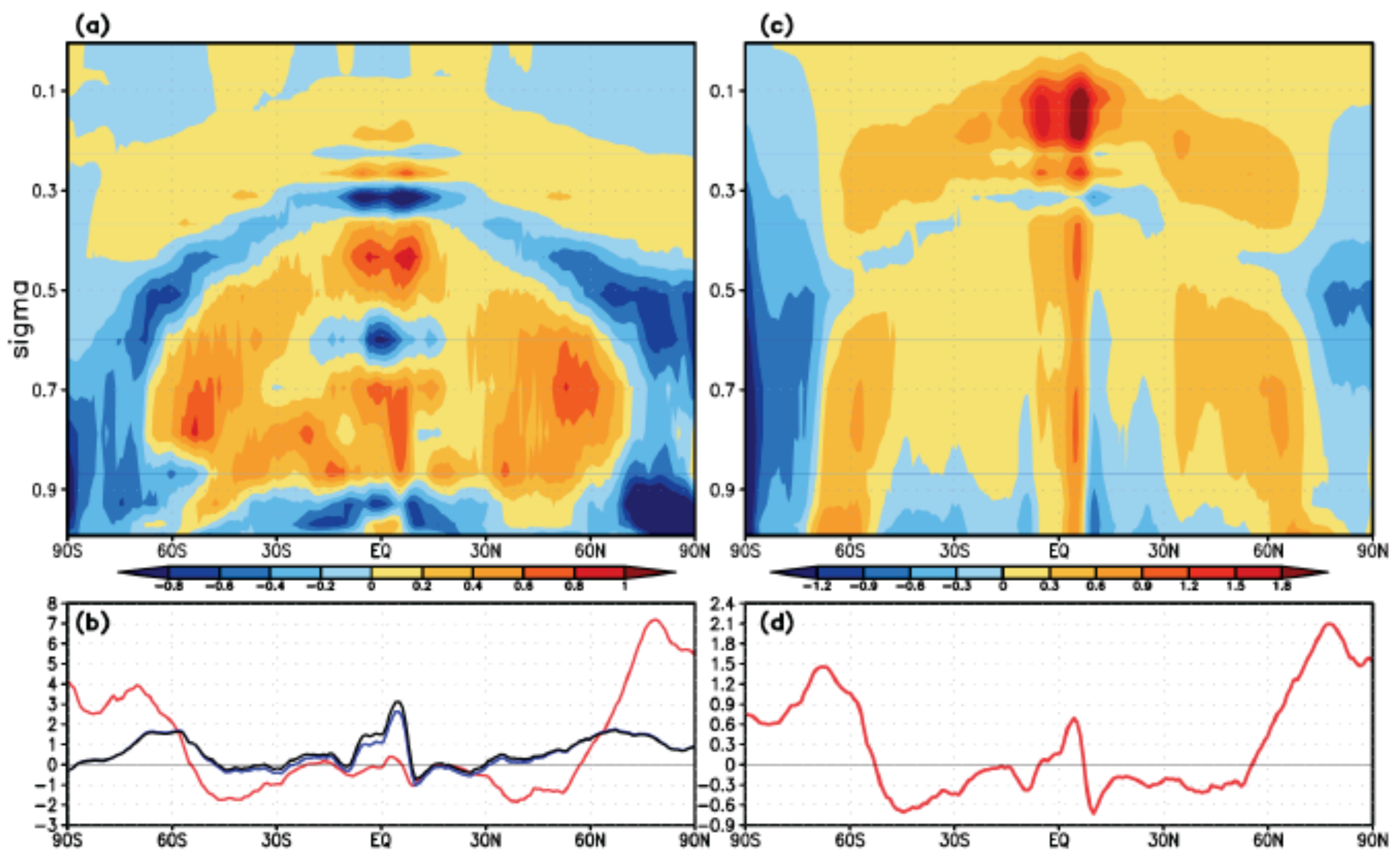
Changes in ice clouds

$[\Delta F^{(Cloud\_SW)}]$

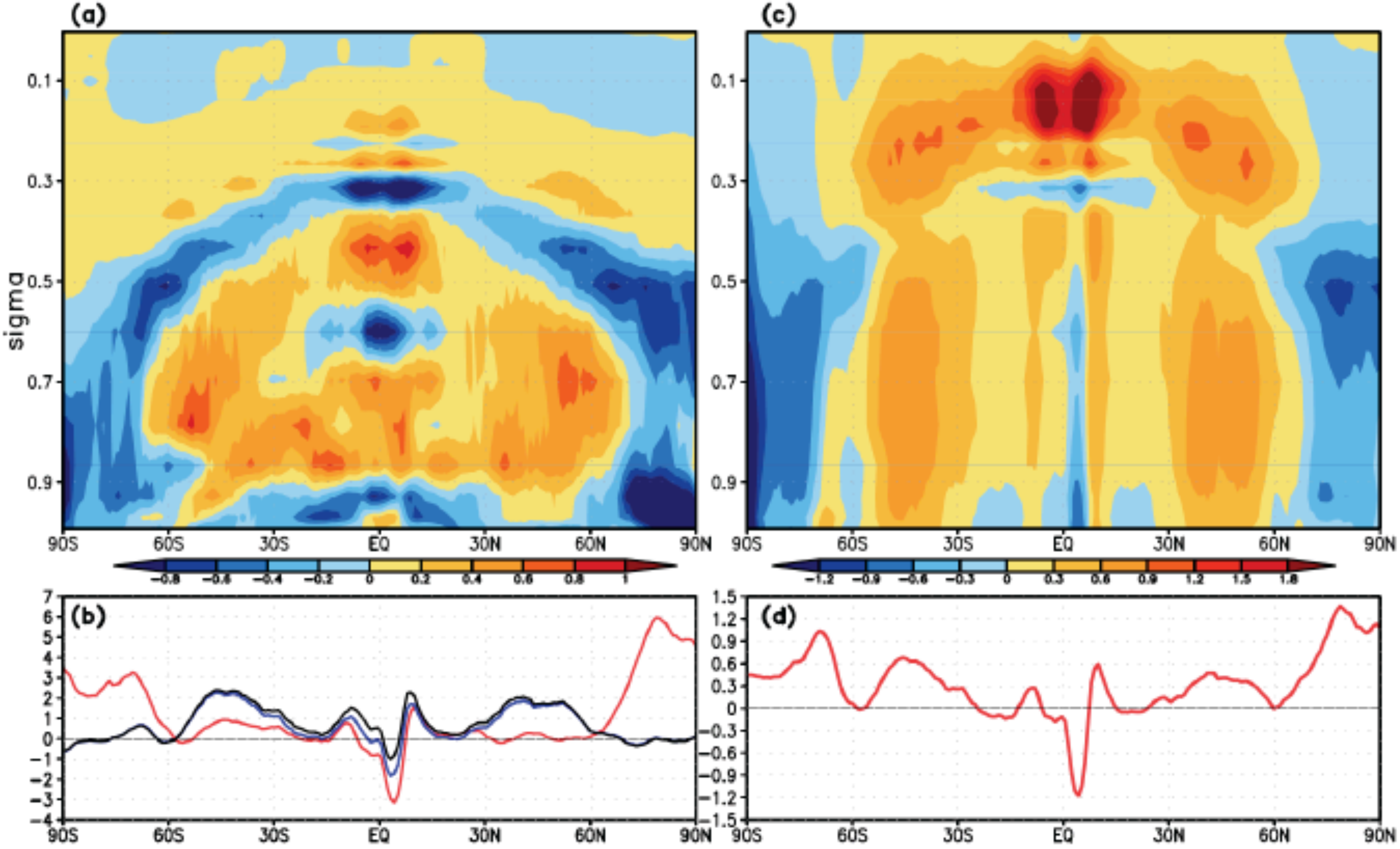
$[\Delta T^{(Cloud\_SW)}]$



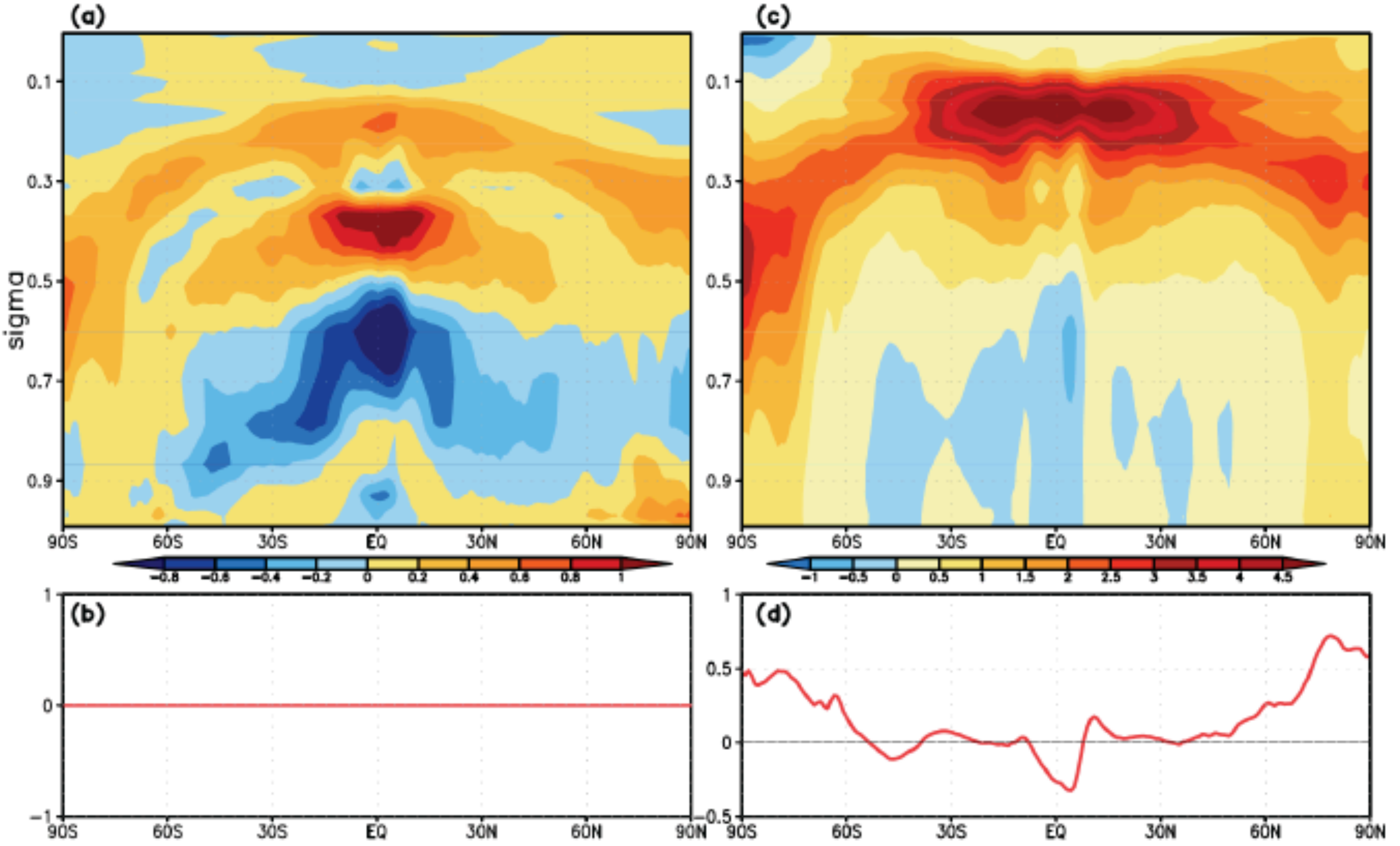
$[\Delta F^{(Cloud\_LW)}]$ $[\Delta T^{(Cloud\_LW)}]$

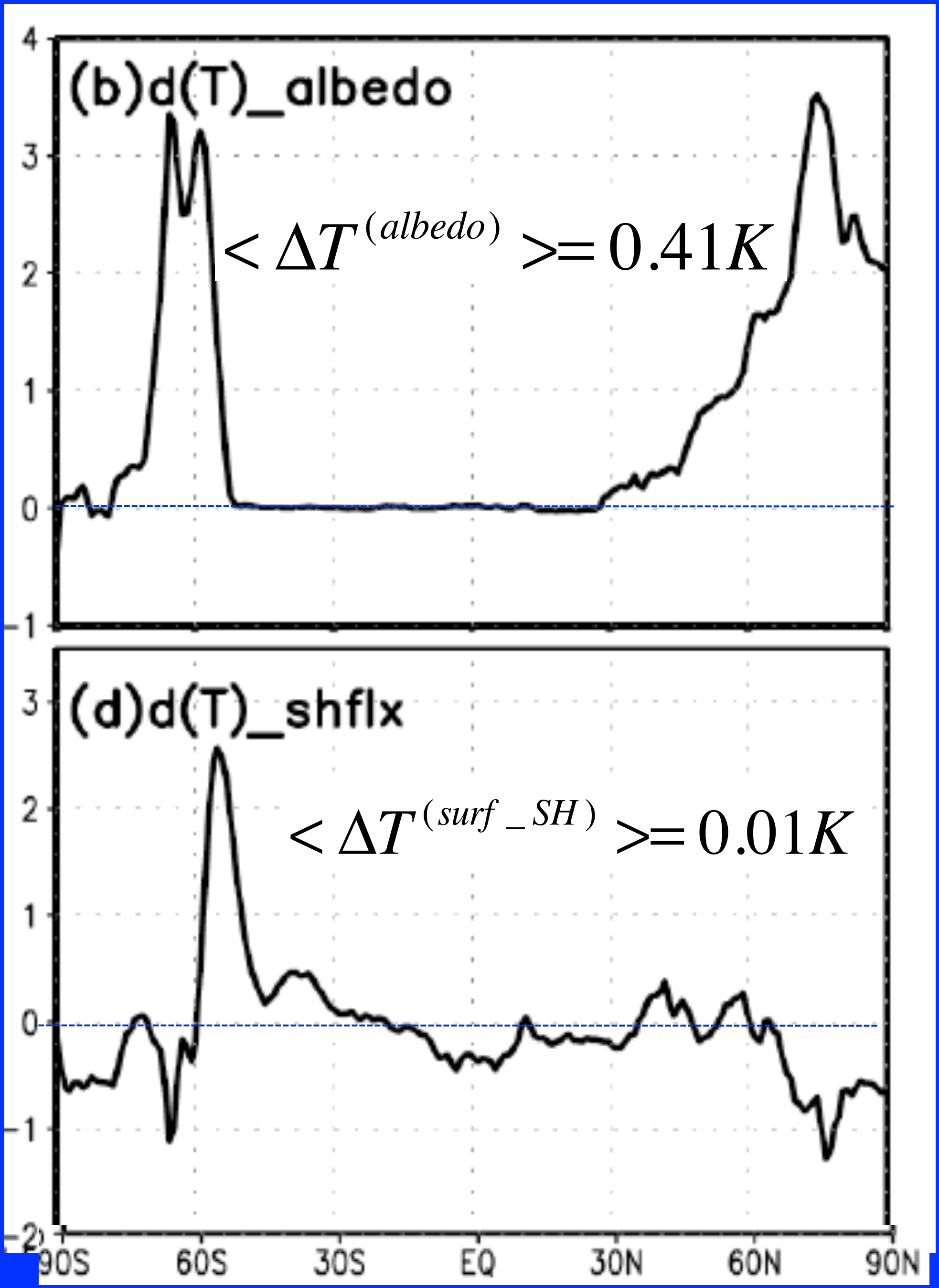


$[\Delta F^{(Cloud\_NET)}]$ $[\Delta T^{(Cloud\_NET)}]$



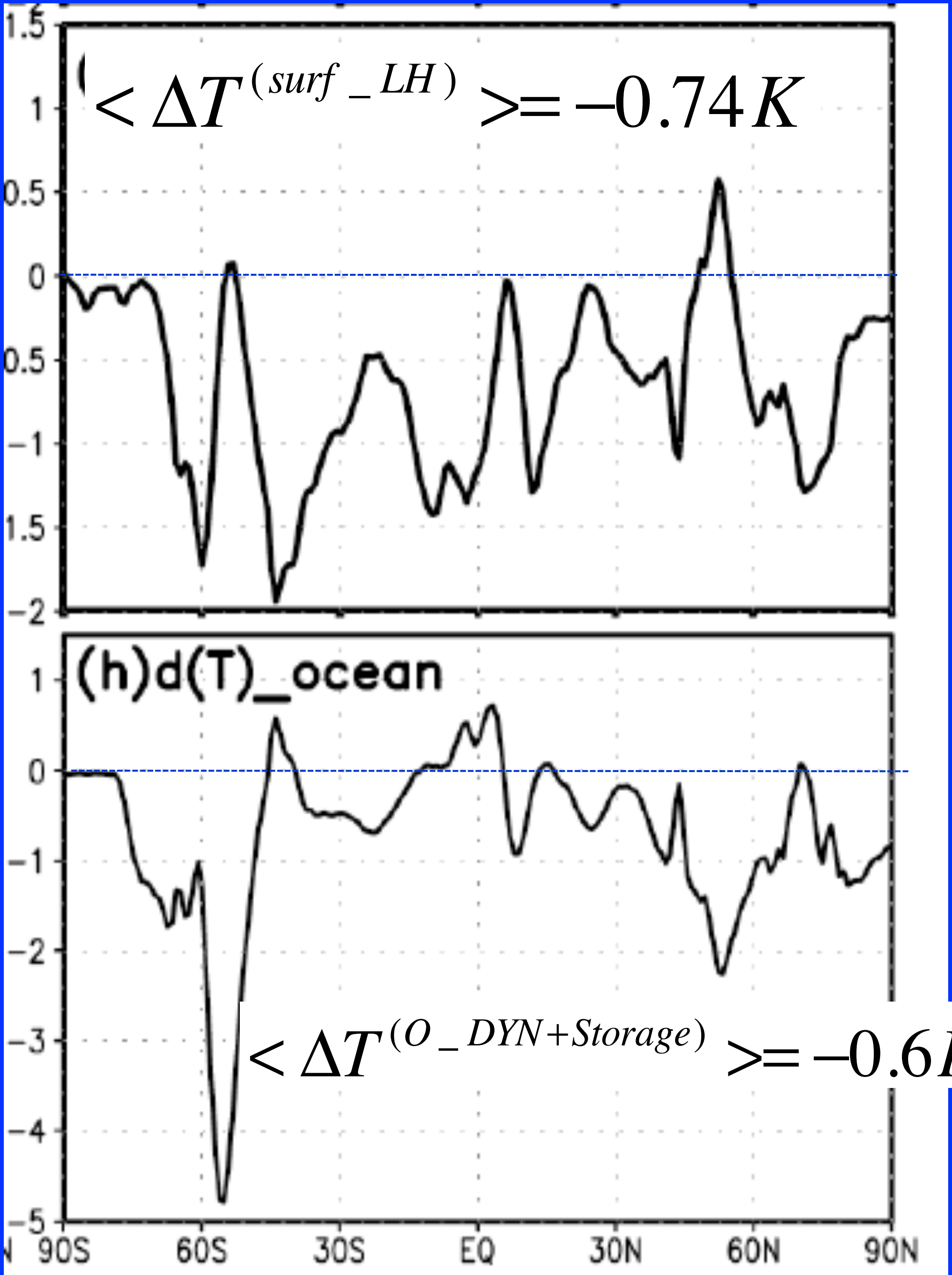
$[\Delta F^{(Atmos\_DYN)}]$ $[\Delta T^{(Atmos\_DYN)}]$





$$< \Delta T^{(albedo)} >= 0.41K$$

$$< \Delta T^{(surf\_SH)} >= 0.01K$$

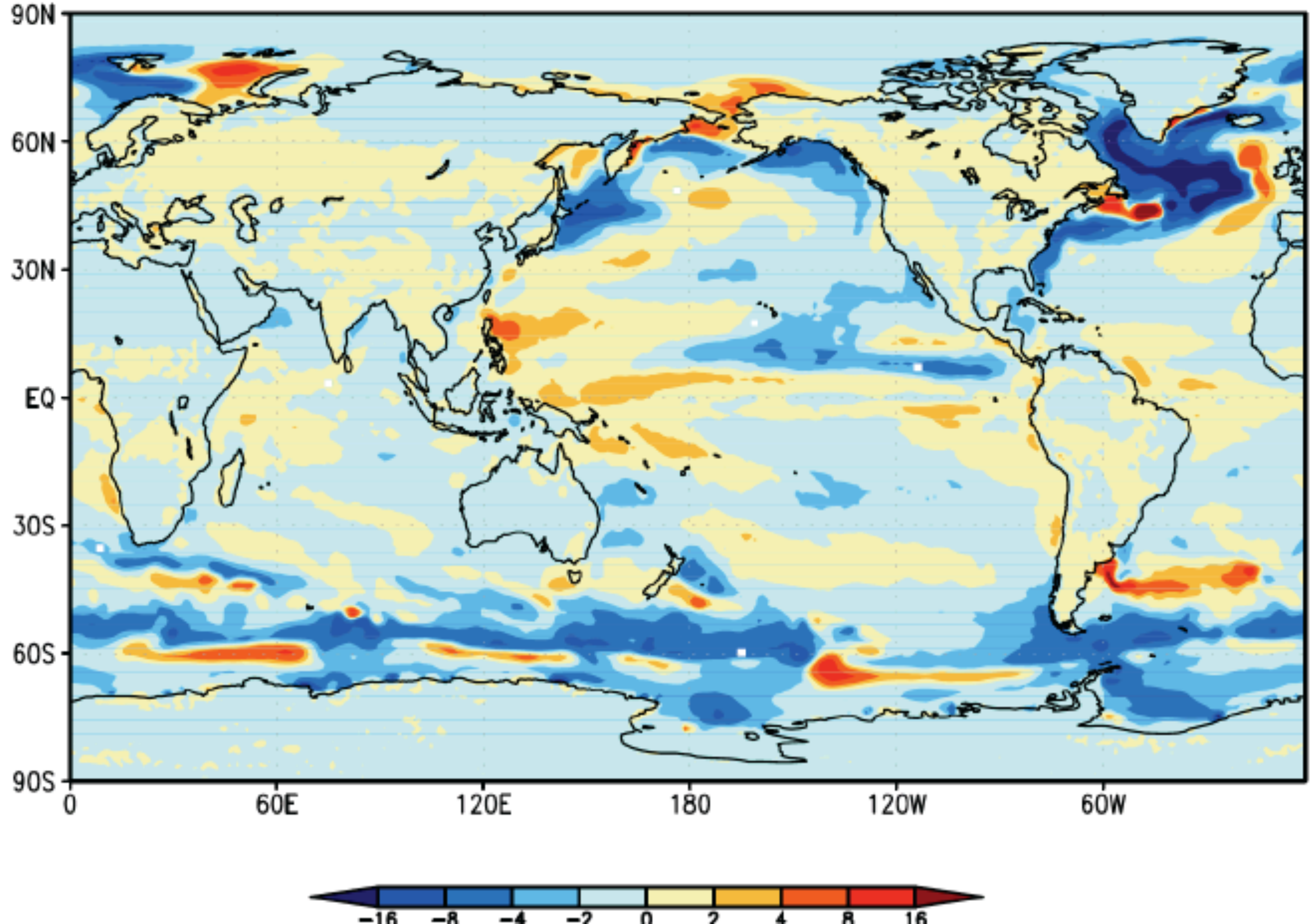


$$< \Delta T^{(surf\_LH)} >= -0.74K$$

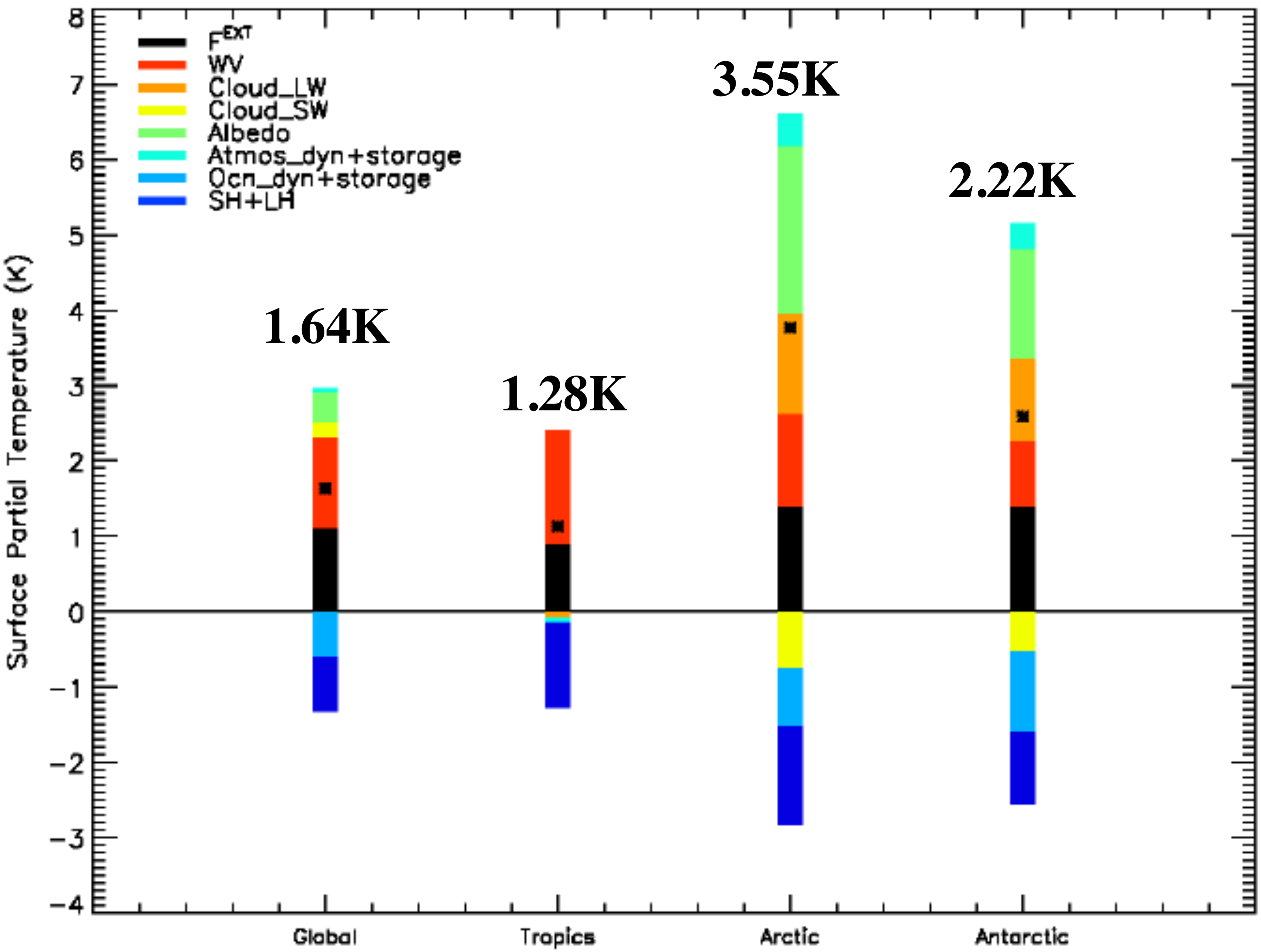
$$< \Delta T^{(O\_DYN+Storage)} >= -0.6K$$

Ocean transport and storage term

ΔT_ocean_dyn_heat_storage> = -0.60K



Attributions



1.64K

1.28K

3.55K

2.22K

Summary 2

- **The linearization of radiation transfer model is a good approximation for global warming climate feedback analysis.**
- **Sum of partial temp. changes is very close to the total temp. change (validation of CFRAM).**
- **2CO$_2$ forcing and water vapor feedback tend to create largest warming at the lower troposphere and surface.**
- **Evaporation feedback acts to reduce warming over the vast global surface while cloud shortwave radiative feedback mainly reduces warming over the warm pool area.**
- **Cloud longwave radiative and atmospheric dynamical (non-radiative) feedbacks tend to place larger warming in upper troposphere.**
- **2CO$_2$ forcing, cloud longwave, and surface albedo radiative feedbacks and atmospheric dynamical feedbacks all contribute to stronger surface warming at high latitudes.**
- **Atmospheric circulation expands upward and poleward.**