Is the Atmosphere an Upside-down Ocean?

On the Troposphere and the Thermocline

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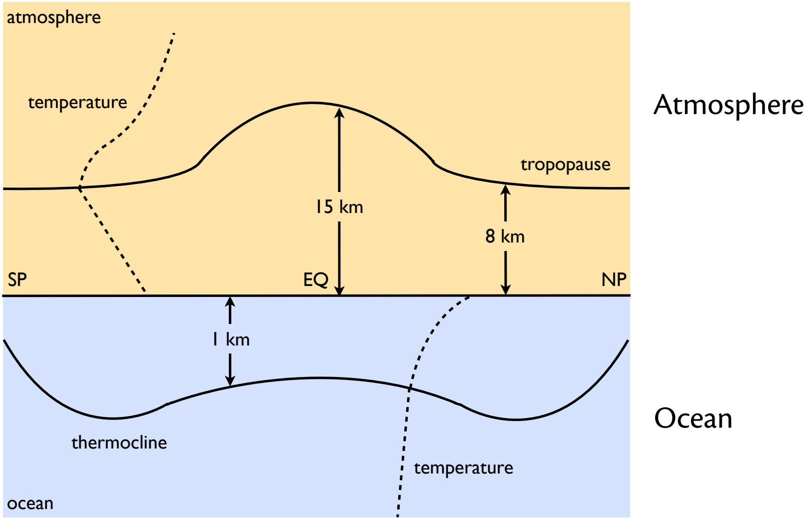
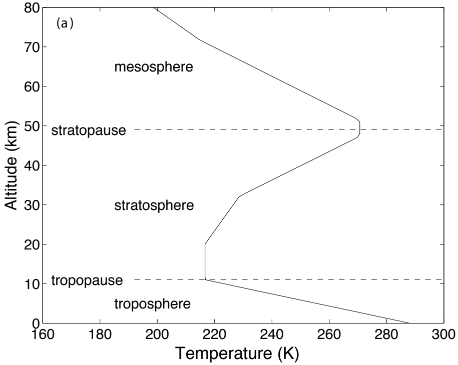
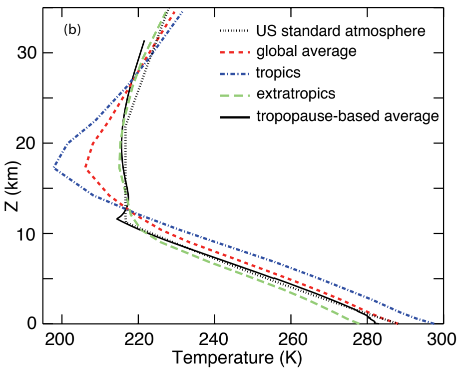
Atmosphere and Ocean - Schematic, Large Scale



The Standard Atmosphere



Observed Atmosphere



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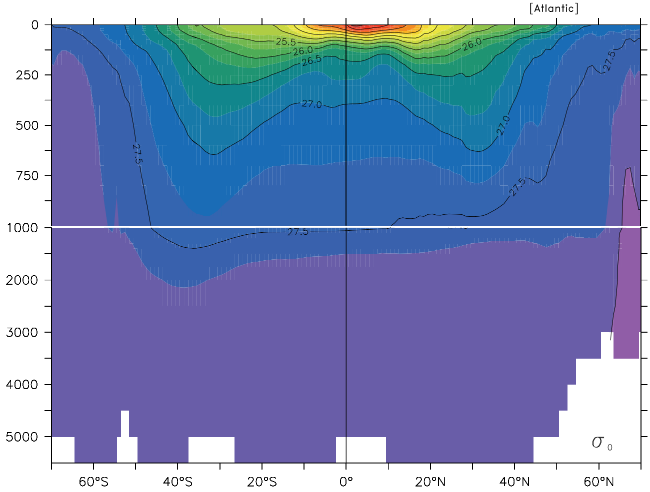
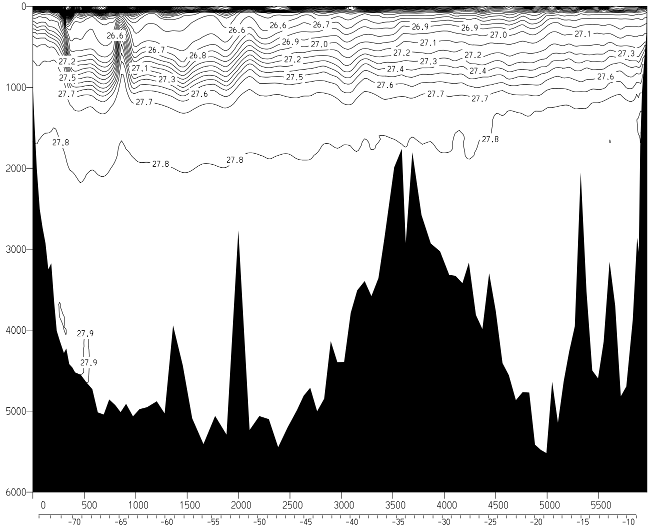
WMO definition: The tropopause is the lowest level at which the lapse rate decreases to 2 $\mathrm{K\,km^{-1}}$ or less, provided also that the average lapse rate between this level and all higher levels within 2 km does not exceed 2 $\mathrm{K\,km^{-1}}$.

Potential Density, Atlantic zonal average, observed



Potential Density, section, Atlantic

Observed, equal contours



Observations and Questions

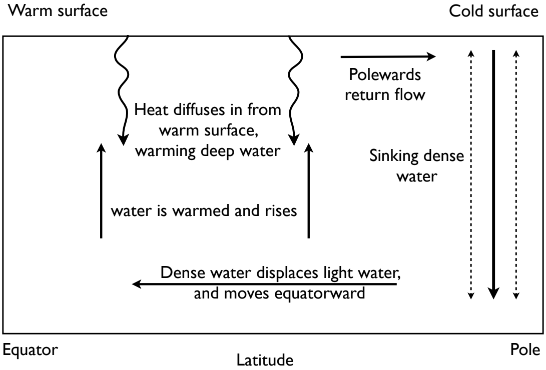
- The lower atmosphere (troposphere) is a region of fast dynamics, in which the stratification and temperature fields are set by fluid motion – convection or baroclinic instability.
- Above the troposphere lies the stratosphere, which at lowest order is in radiative equilibrium.
- The tropopause is the dividing line between the two.

- The upper ocean (thermocline) is a region of fast dynamics, in which the stratification and temperature fields are set by fluid motion – convection or subduction, perhaps baroclinic instability.
- Below the thermocline lies the abyss in which flow is sluggish with properties determined by where the water comes from.

- How are these systems the same, and how are they different?
- Particular emphasis on the thermocline and the tropopause.

Theories of the Thermocline and the MOC

A mixing-driven theory (Robinson, Stommel, Arons, Munk etc)



The ocean interior is *almost* adiabatic (but not quite).

$$\frac{\mathrm{D}T}{\mathrm{D}t} = \kappa \nabla^2 T$$

but κ is 'small'. That is

$$\kappa \ll UL, \quad \text{or} \quad \mathrm{P}_e \equiv \frac{UL}{\kappa} \gg 1.$$

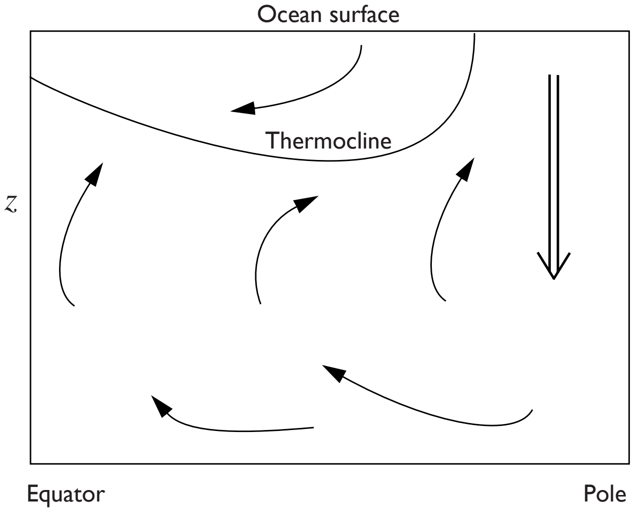
Dynamical balance between upwelling and diffusion

$$w\frac{\partial T}{\partial z} = \kappa \frac{\partial^2 T}{\partial z^2}$$

with solution (for κ and w given)

$$T = \Delta T\, \mathrm{e}^{wz/\kappa} + \text{constant}.$$

Thermocline with wind forcing too



So the thermocline is a *boundary layer,*
connecting the abyss to the surface.

Quantitative theory

Planetary geostrophic equations

$$-fv = -\frac{\partial \phi}{\partial x}, \qquad fu = -\frac{\partial \phi}{\partial y}, \qquad b = \frac{\partial \phi}{\partial z},$$

$$\nabla \cdot \boldsymbol{v} = 0, \qquad \frac{\partial b}{\partial t} + \boldsymbol{v} \cdot \nabla b = \kappa \nabla^2 b,$$

where b is the buoyancy. Define M such that $\phi = M_z$, $w = \beta M_x / f^2$ and obtain:

The 'M-equation': $$\frac{\partial M_{zz}}{\partial t} + \frac{1}{f} J(M_z, M_{zz}) + \frac{\beta}{f^2} M_x M_{zzz} = \kappa M_{zzzz}.$$

Restrict attention to steady flow of one horizontal dimension, then

$$\boxed{\frac{\beta}{f^2} M_x M_{zzz} = \kappa M_{zzzz}}$$

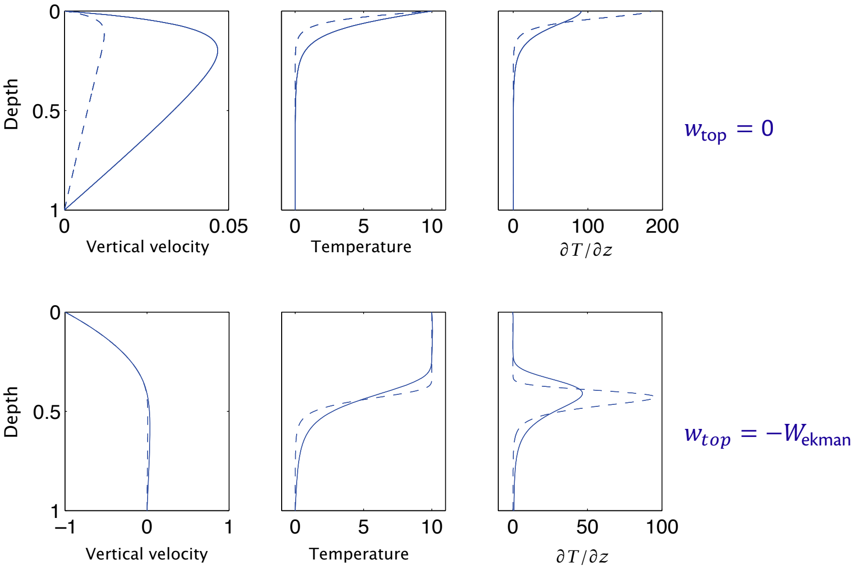
which, as $b = M_{zz}$, represents

$$w \frac{\partial b}{\partial z} = \kappa \frac{\partial^2 b}{\partial z^2}.$$

Apply boundary conditions on w (M_x) and b (M_{zz}) at top and bottom.

Solutions

(Dashed lines have smaller diffusivity)



If $w_{\text{top}} \neq 0$ the thermocline is internal to the fluid (an 'internal boundary layer').

Thickness and Depth

- The thermocline has a *depth* and a *thickness*.
  - These are dynamically distinct, but less so in actuality.
- Depth, D, is determined by wind:

$$D = W_E^{1/2} \left(\frac{f^2 L}{\beta \Delta b} \right)^{1/2}$$

- Thickness, δ, is determined by the diffusivity.

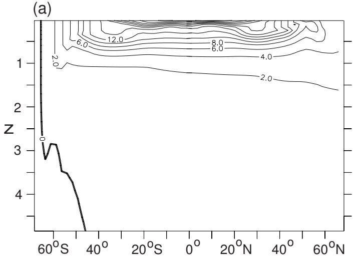
$$\delta = \kappa^{1/2} \left(\frac{f^2 L}{\Delta b \beta W_E} \right)^{1/4} .$$

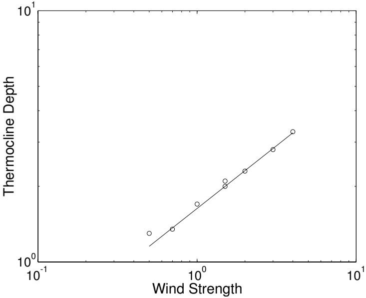
As $\delta \to 0$ temperature changes discontinuously.

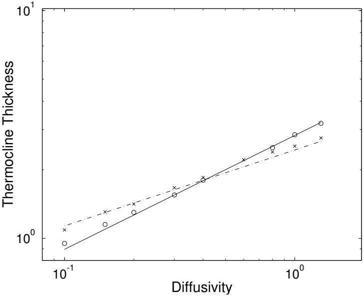
In the limit of zero diffusivity the temperature changes discontinuously at the base of the thermocline

Scaling vs GCM tests

Thickness and depth of the thermocline





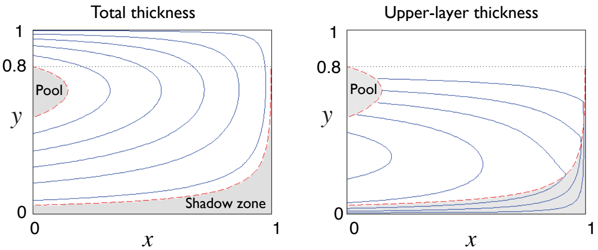


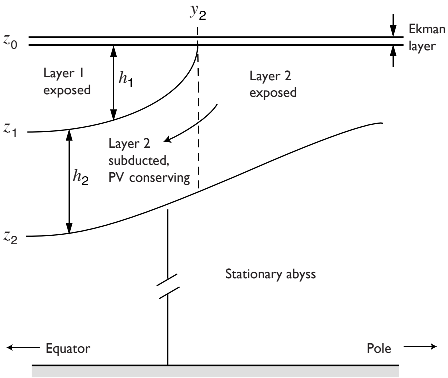
Complications

In fact there is a whole load of dynamics going on *above* the internal boundary layer. This is the region of ocean gyres! Potentially:

1. Stommel-like gyres,
2. A ventilated thermocline (LPS theory).
3. Homogenization of potential vorticity by mesoscale eddies (Rhines & Young).

The ventilated thermocline





But the diffusive layer at the bottom of the ventilated thermocline is unavoidable!

A diffusive layer is required to connect the temperatures at the bottom of the thermocline and the abyss.

Paradigms lost

Alas, the oceanic diffusivity is too small to support a deep circulation as strong as that observed.

We need $\kappa = 10^{-4}\ \mathrm{m^2\ s^{-1}}$. We observe $\kappa = 10^{-5}\ \mathrm{m^2\ s^{-1}}$.

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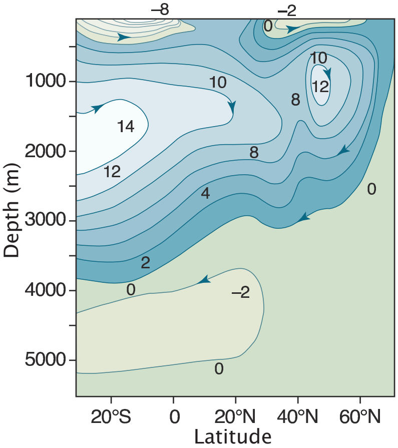
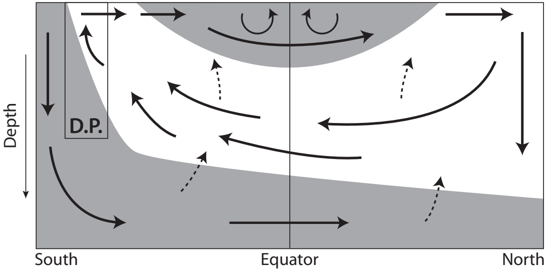
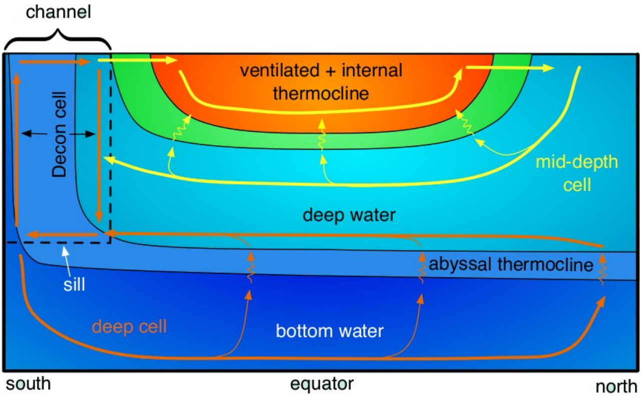


State estimation (reanalysis)



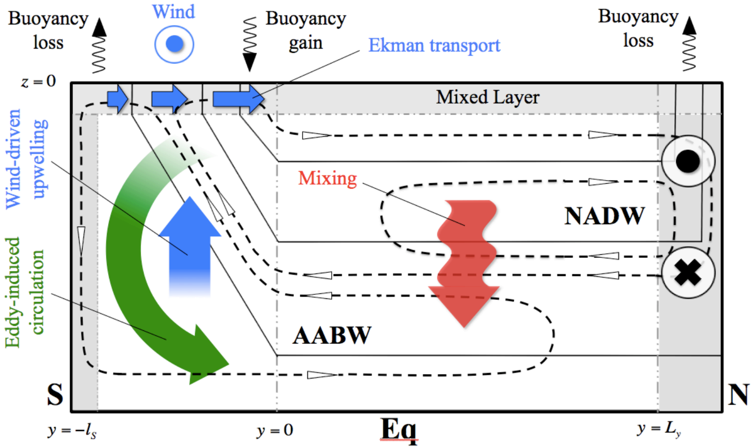
Schematic

A Theory for Deep Stratification   the conceptual model



- In channel, the wind generates a clockwise 'Deacon Cell' that tries to make the isopycnals vertical.
- Balanced by baroclinic eddies try to make the isopycnals horizontal.
- Vertical destabilization by the wind (ocean) vs radiation (atmosphere).
- The circulation in the channel must connect smoothly with that in the basin.

Schematic for Interhemispheric Flow



Circumpolar Current region: wind-driven upwelling balanced by eddy-induced circulation

Ocean interior: downward mixing of buoyancy balanced by upward vertical advection

North Atlantic high latitudes: convection due to buoyancy loss

The three regions are matched so the solutions are smooth.

The Thermocline --- summary

- The thermocline is a boundary layer, connecting the deep cold abyss to the warm surface.
- It will be an *internal* boundary layer if there is a wind forcing.
- In general, as the diapycnal diffusivity goes to zero, the base of the thermocline will be marked by a *discontinuity* in temperature.
- Scaling seems to work, compared to (non-eddying) OGCMs.
- Zeroth order dynamics can be described in the absence of baroclinic eddies.
- The abyssal flow is in part wind-driven (!) but that is a story for another day.

The thermocline separates (or contains) *fast* dynamics. Most of it is advective, with

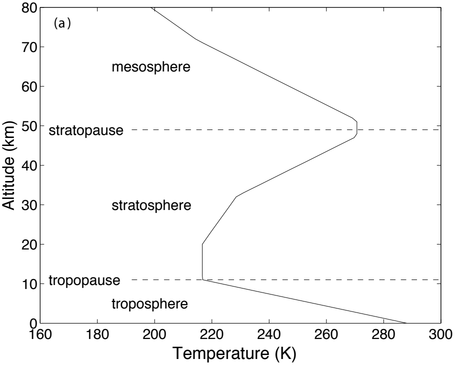
$$T \sim L/V = 1000\,\mathrm{km}/1\,\mathrm{cm\,s^{-1}} \approx 10^8\,\mathrm{s} \approx \text{a few years.} \tag{1}$$

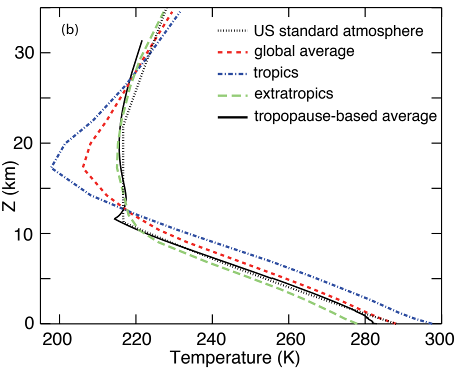
Timescales in the abyss are up to *two orders of magnitude longer.*

THE ATMOSPHERE

The Standard Atmosphere

Observed Atmosphere





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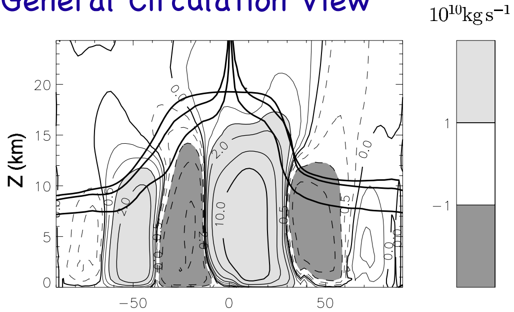
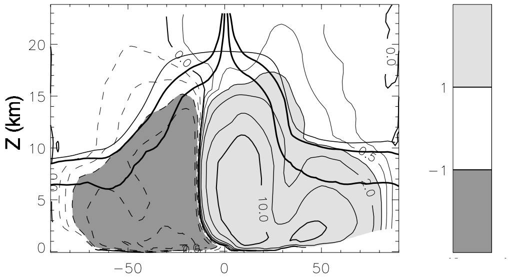
A General Circulation View



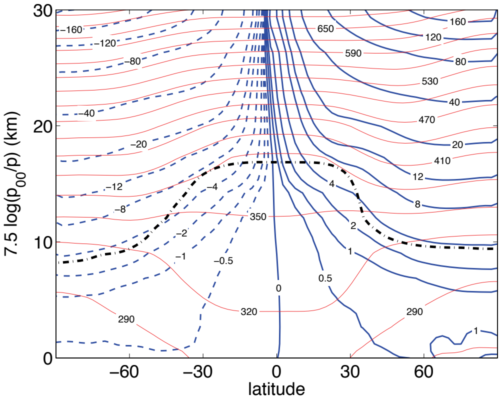
Eulerian MOC and tropopause.



Residual MOC and tropopause.

The troposphere is the boundary layer in which the dynamical redistribution of energy and momentum by eddies takes place. Adapted from Juckes, 2004.

Troposphere and stratosphere



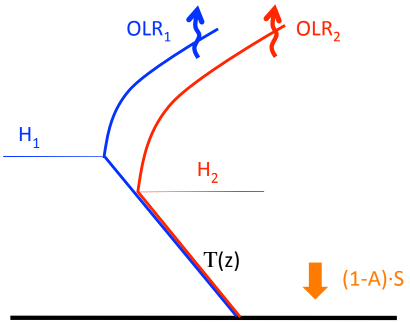
- The troposphere is dynamically active, fairly well mixed tracers. 'Fast' dynamics and short residence times (young air)
- Stratosphere has slow dynamics, in near radiative equilibrium, long residences times.

- So the tropopause is a kind of *front.* No need for convection, or ozone, but these will almost certainly be important in practice.
- Define the (mid-latitude) tropopause as a given isocontour of poleward mass flux.
  - Is this the same as the WMO definition?

Tropopause Height

A radiative constraint

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



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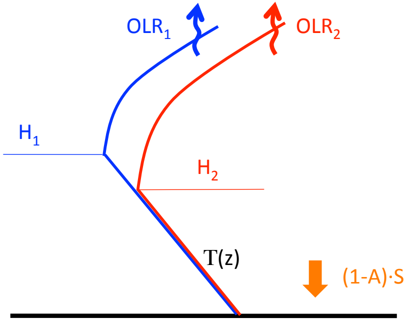

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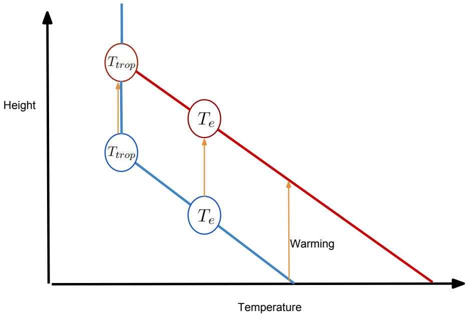


Tropopause height increases with increased COT.

Tropopause Height

With Gray Radiation and a `Thin' Stratosphere

$$\frac{\partial U}{\partial \tau} = U - B \qquad \frac{\partial D}{\partial \tau} = B - D,$$

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

$$\frac{\partial}{\partial \tau}(U - D) = U + D - 2B, \qquad \frac{\partial}{\partial \tau}(U + D) = U - D$$

Stratosphere in longwave radiative equilibrium:

$$D = \frac{\tau}{2} OLR, \qquad U = \left(1 + \frac{\tau}{2}\right) OLR, \qquad B = \frac{1 + \tau}{2} OLR.$$

and if $\tau \ll 1$

$$D = 0, \qquad U = OLR = 2B, \qquad B = OLR/2.$$

Stratosphere is isothermal. Tropopause temperature fixed by OLR.

Height of the tropopause

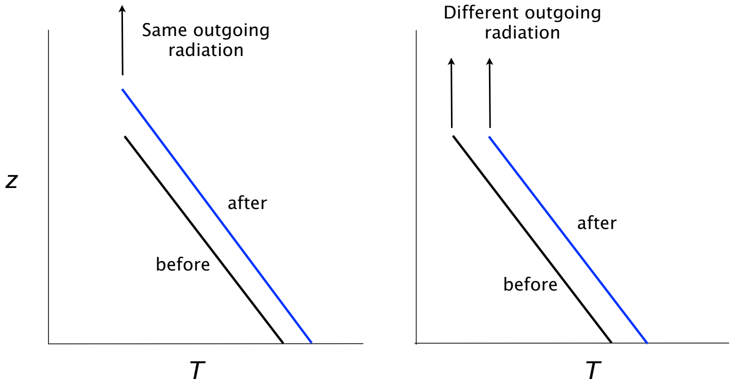
(also Thuburn and Craig c. 2002, Santer)

Even as we add greenhouse gases, the OLR is *fixed* independently of optical depth.

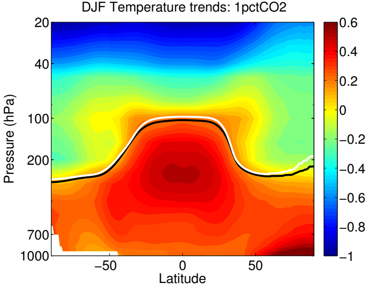
In the troposphere $T = T_s - \Gamma z, \qquad z \leq H_T$

and at the tropopause $U = OLR = 2B = \text{constant}.$

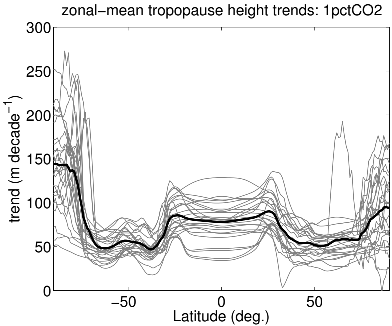
Tropopause temperature fixed and height must increase: $\boxed{\Delta H_T = \dfrac{\Delta T}{\Gamma}}$

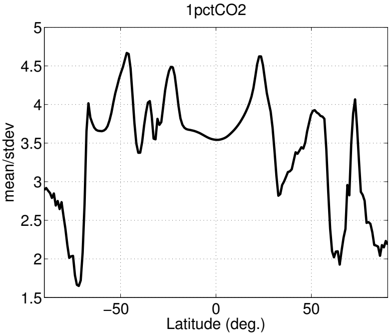


Predicted Temperature and tropopause height changes

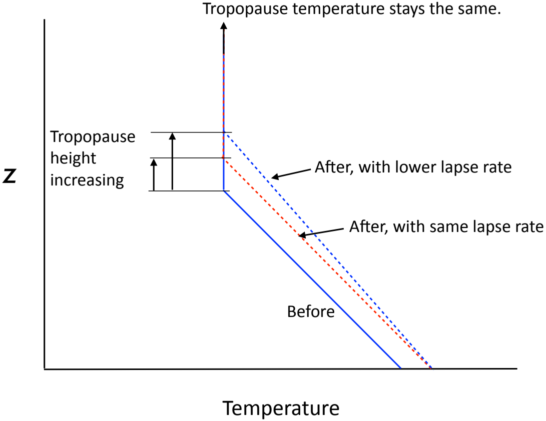


Increase in tropopause height is common across models. The change in height is greater than the model standard deviation, especially in low latitudes.





Changes in tropopause height:
Lapse rate and temperature effects



$$\Delta H_T = \frac{\Delta T}{\Gamma} - \frac{H_T \Delta\Gamma}{\Gamma}$$

where

T_T is the tropopause temperature,

ΔT is the increase in temperature at a given height in the troposphere

$\Delta\Gamma$ the change in the lapse rate.

That is:

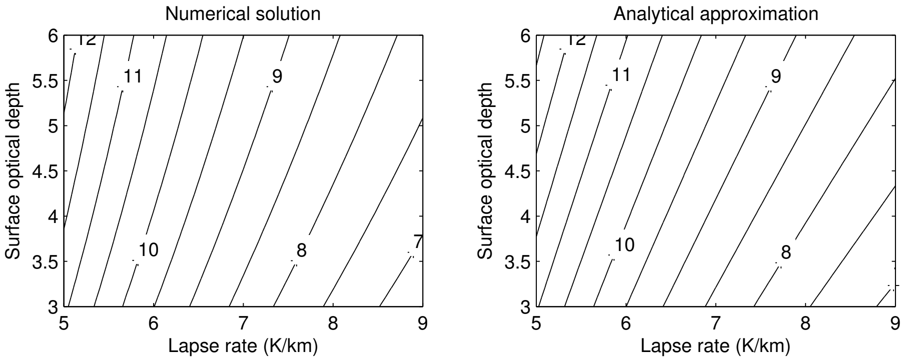
$$\boxed{\frac{\partial H_T}{\partial T} = \frac{1}{\Gamma} - \frac{H_T}{\Gamma}\frac{\partial \Gamma}{\partial T}}$$

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

Tropopause Height: Theoretical Prediction

After some algebra…

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

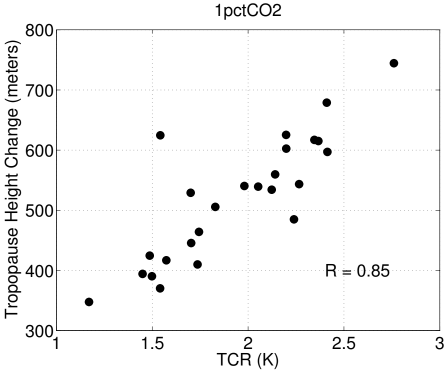


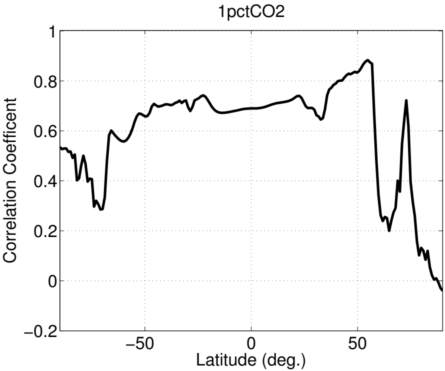
Optically thick and thin limits:

Thick: $H_T \approx \sqrt{\frac{T_T\tau_s H_a}{8\Gamma}}$ Thin: $H_T \approx \frac{1.38T_T}{8\Gamma}$.

CMIP5, results

Tropopause height vs Climate Sensitivity





Change in tropopause height is correlated to the climate sensitivity, both locally and in the mean.

Final Remarks

- The troposphere is a fast, turbulent layer, capped by the tropopause. In midlatitudes the tropopause marks the vertical extent of baroclinic eddies.
  - The thermocline is a fast layer, too, and the slow abyss is akin to the stratosphere.
  - But the stratosphere is in radiative equilibrium (not adiabatic!), the abyss is a near-isothermal, ventilated pool.
- Tropopause will in general mark a change in stratification ($\partial T/\partial z$), although not always sharp.
  - Temperature itself is continuous at the edge of the troposphere because the stratosphere is in radiative equilibrium.
  - In contrast the oceanic abyss is almost adiabatic, and so takes the temperature of its origin (e.g., at the poles). Temperature is discontinuous at the base of the thermocline in the adiabatic limit.
- Tropopause will robustly rise with global warming, and the stratosphere will cool.

Questions

- Where does the Earth's atmosphere sit in parameter space?
- Why is the Earth's mid-latitude tropopause so sharp?
- Is water vapor important because it is a diabatic effect, or because it changes the neutral stability, or something else?
- Suppose we remove ozone. What is the nature of the resulting tropopause? (Experiment with comprehensive GCM?)