Water vapour feedback and the Runaway Greenhouse

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Overview

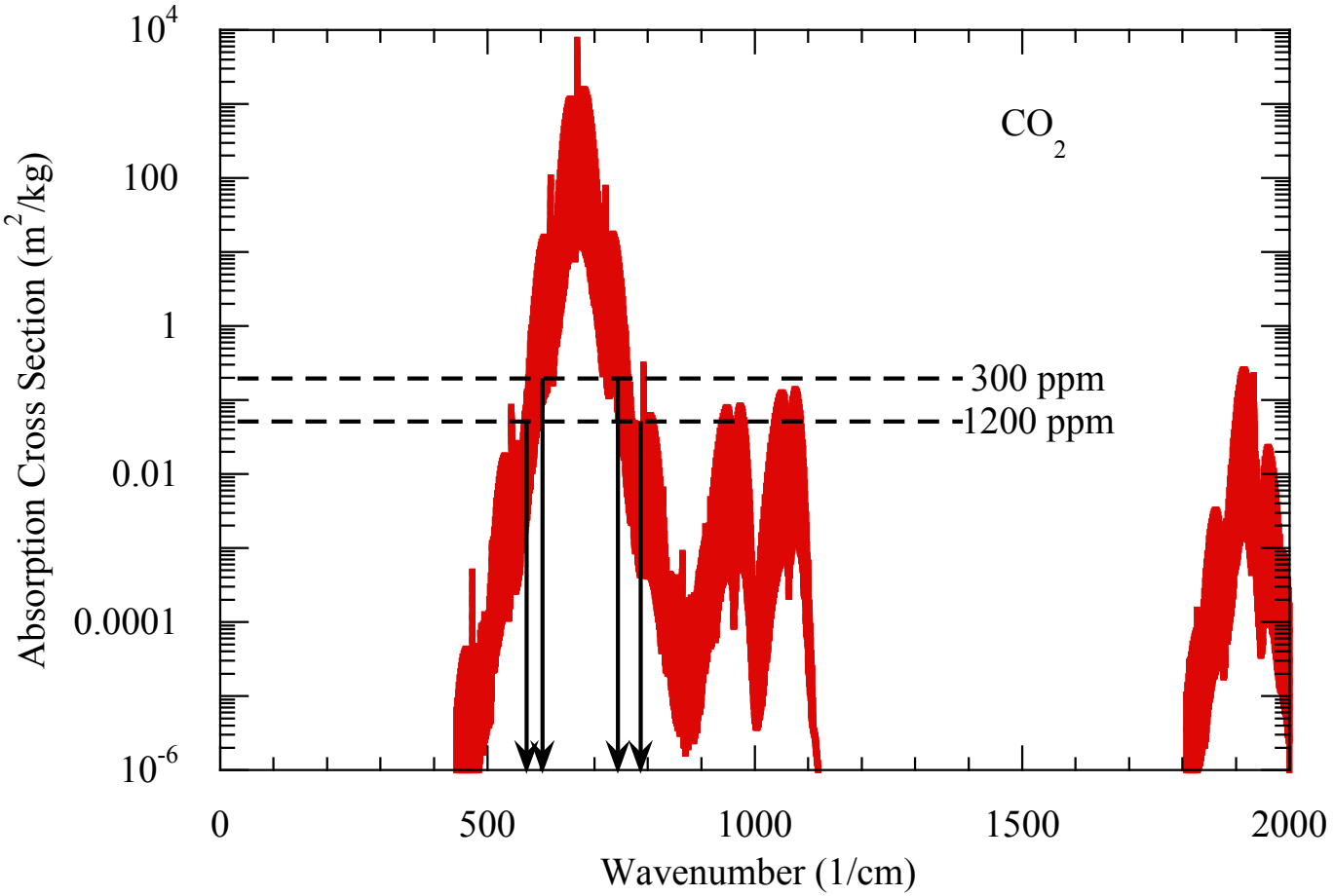
Water vapour spectroscopy

Condensation effects: Equilibrium with a surface ocean/glacier

Radiative forcing and climate response

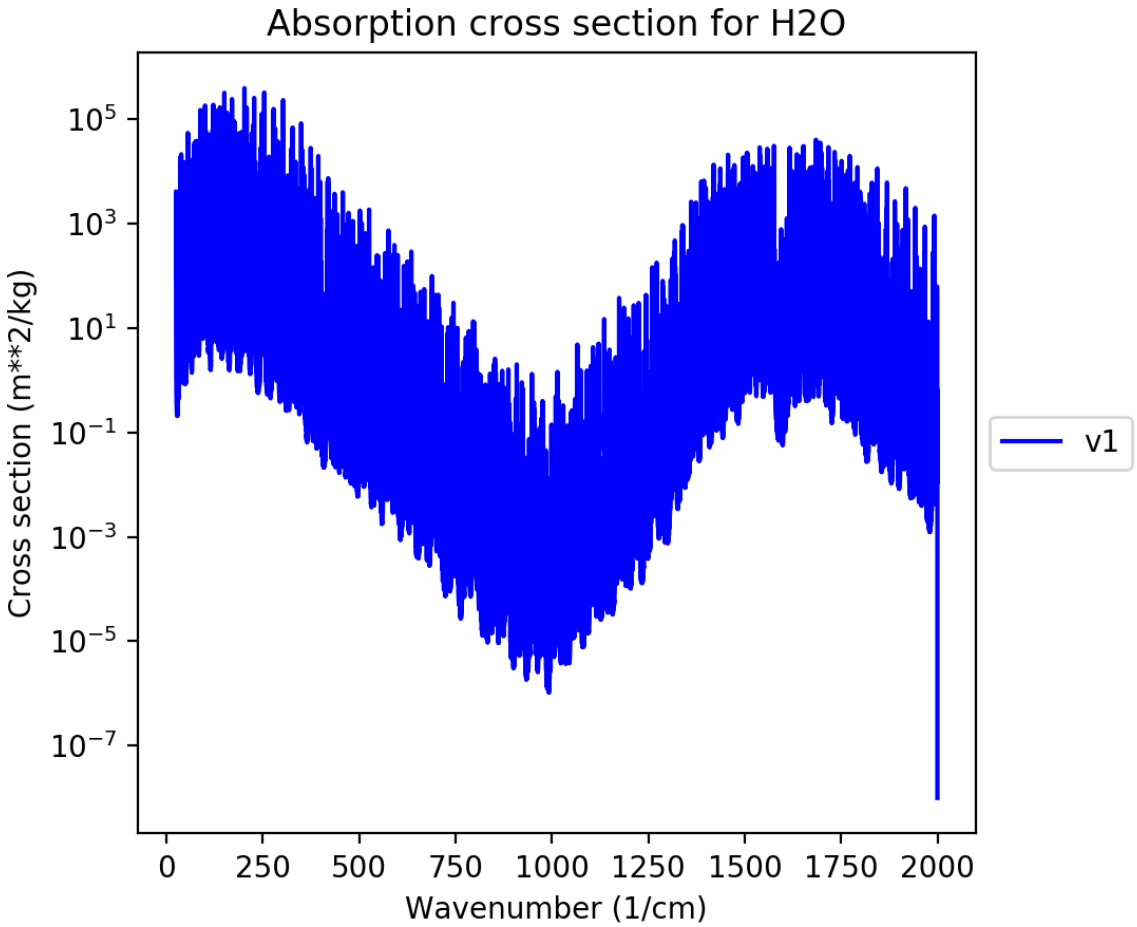
The runaway greenhouse

Reprise: CO_2 absorption spectrum

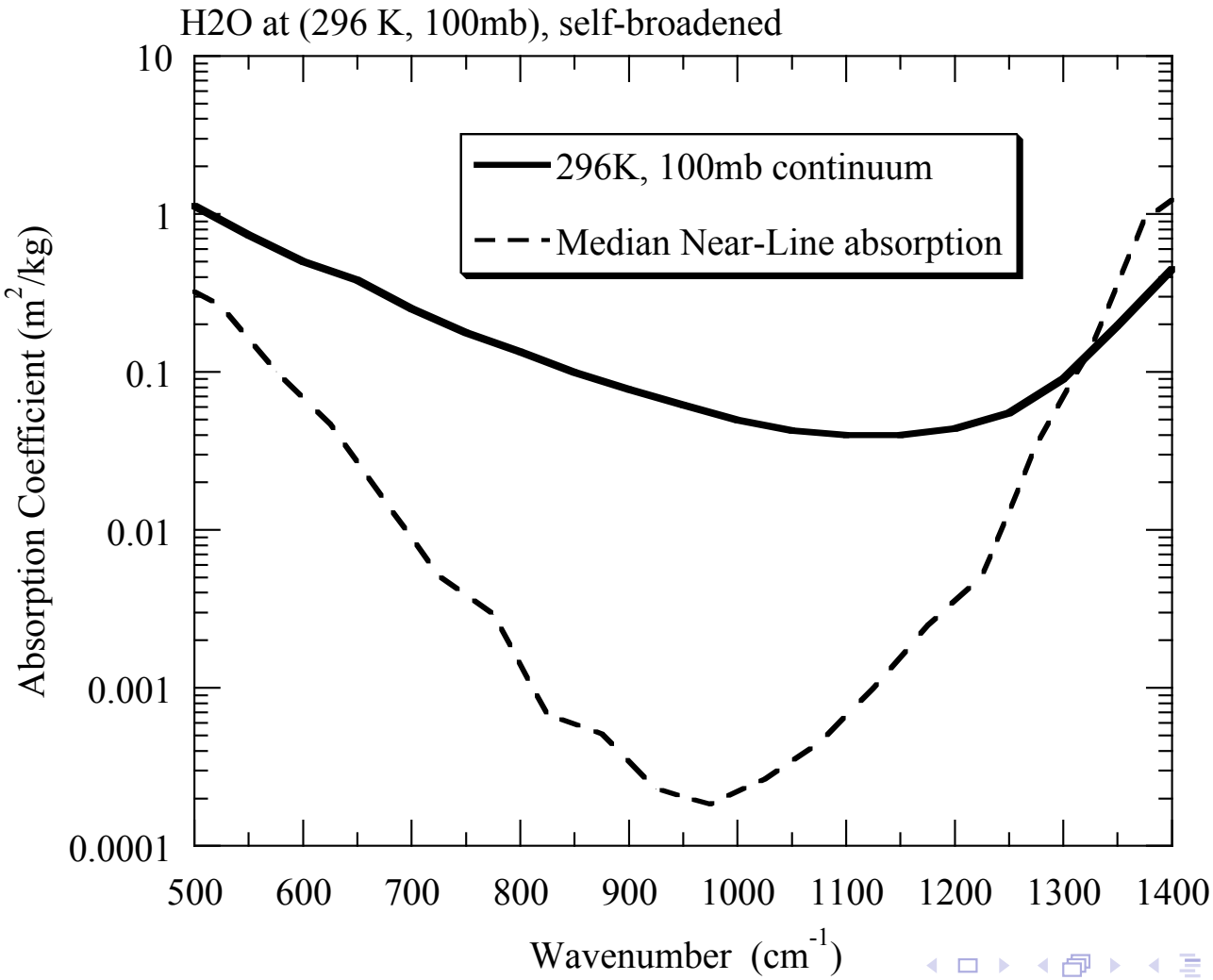


Water vapour spectroscopy

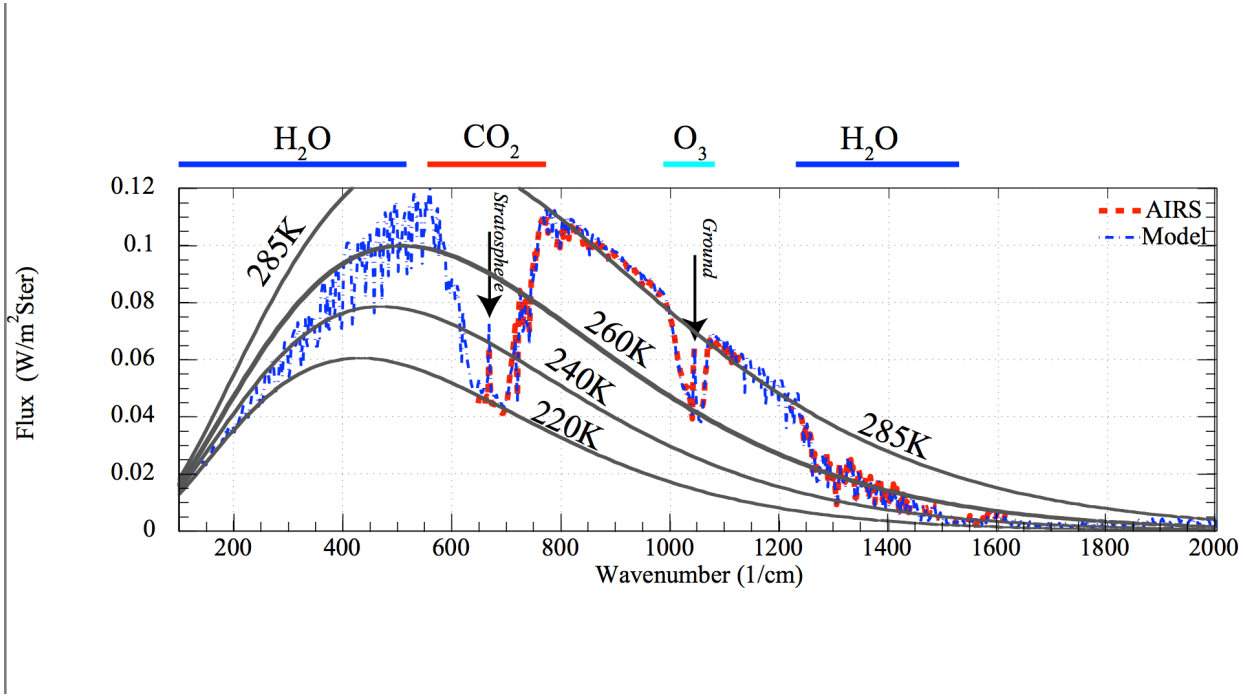
Live explorations in Python

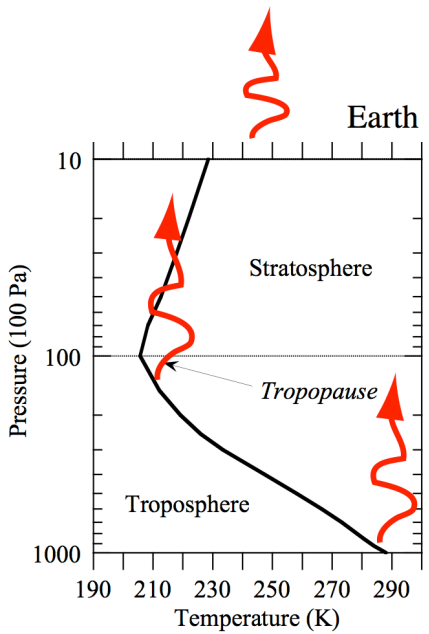


The water vapour continuum



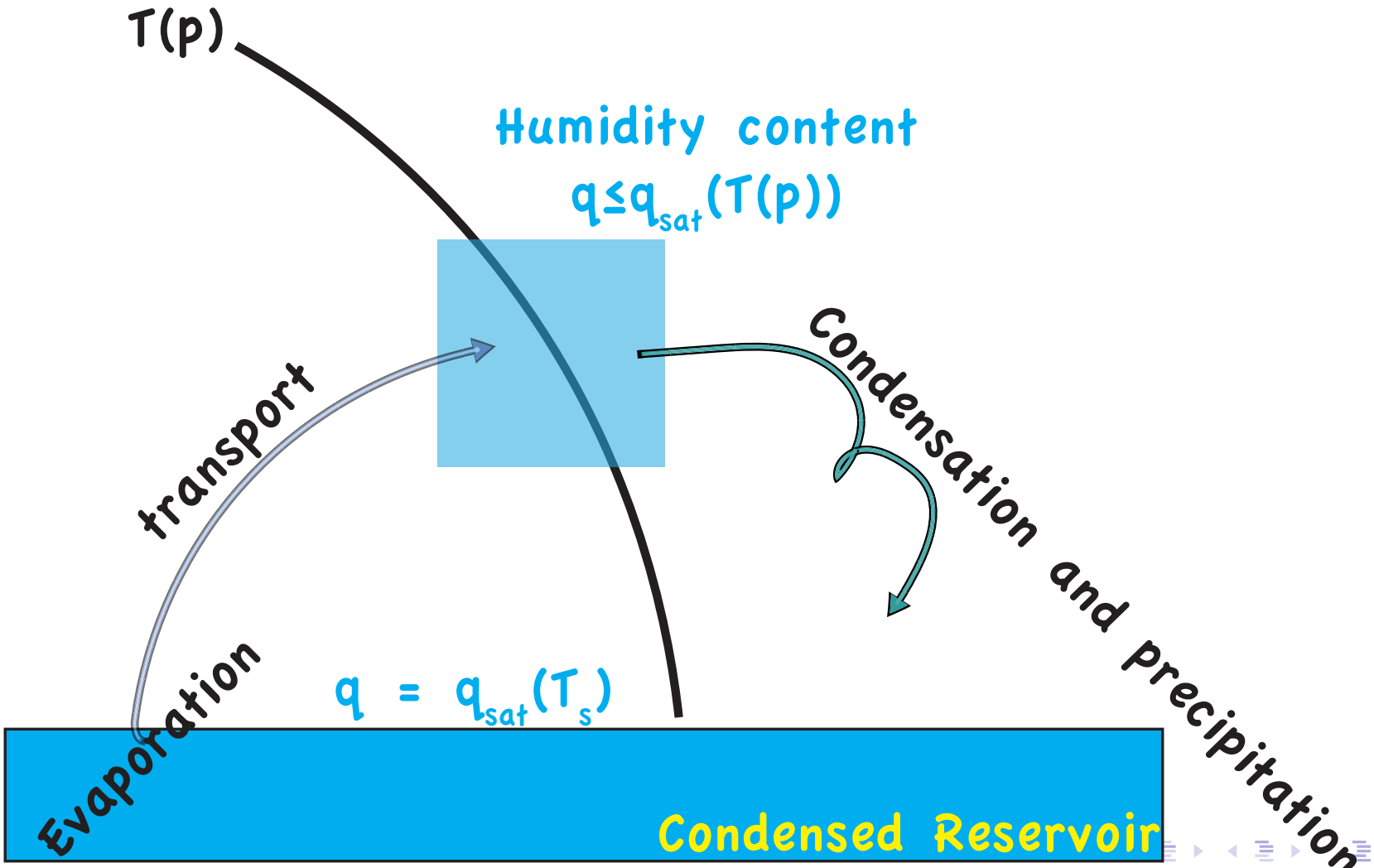
Observed *OLR* spectrum for Earth





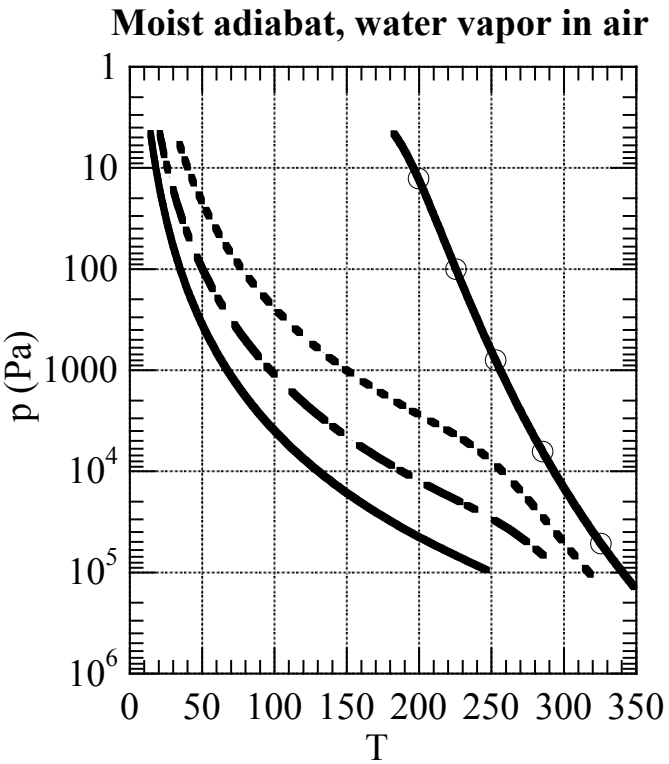
Clear sky

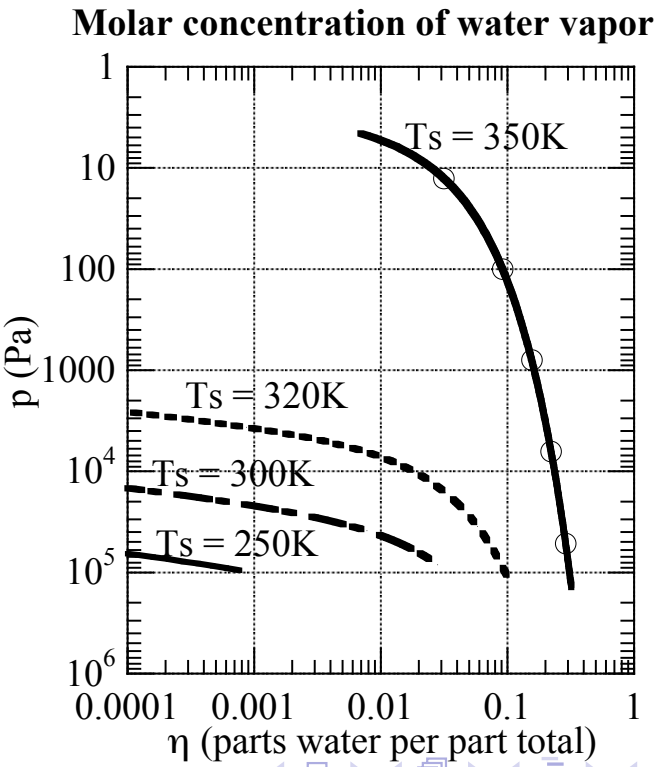
What controls water vapour content of an atmosphere?



Water content of a saturated atmosphere increases with temperature

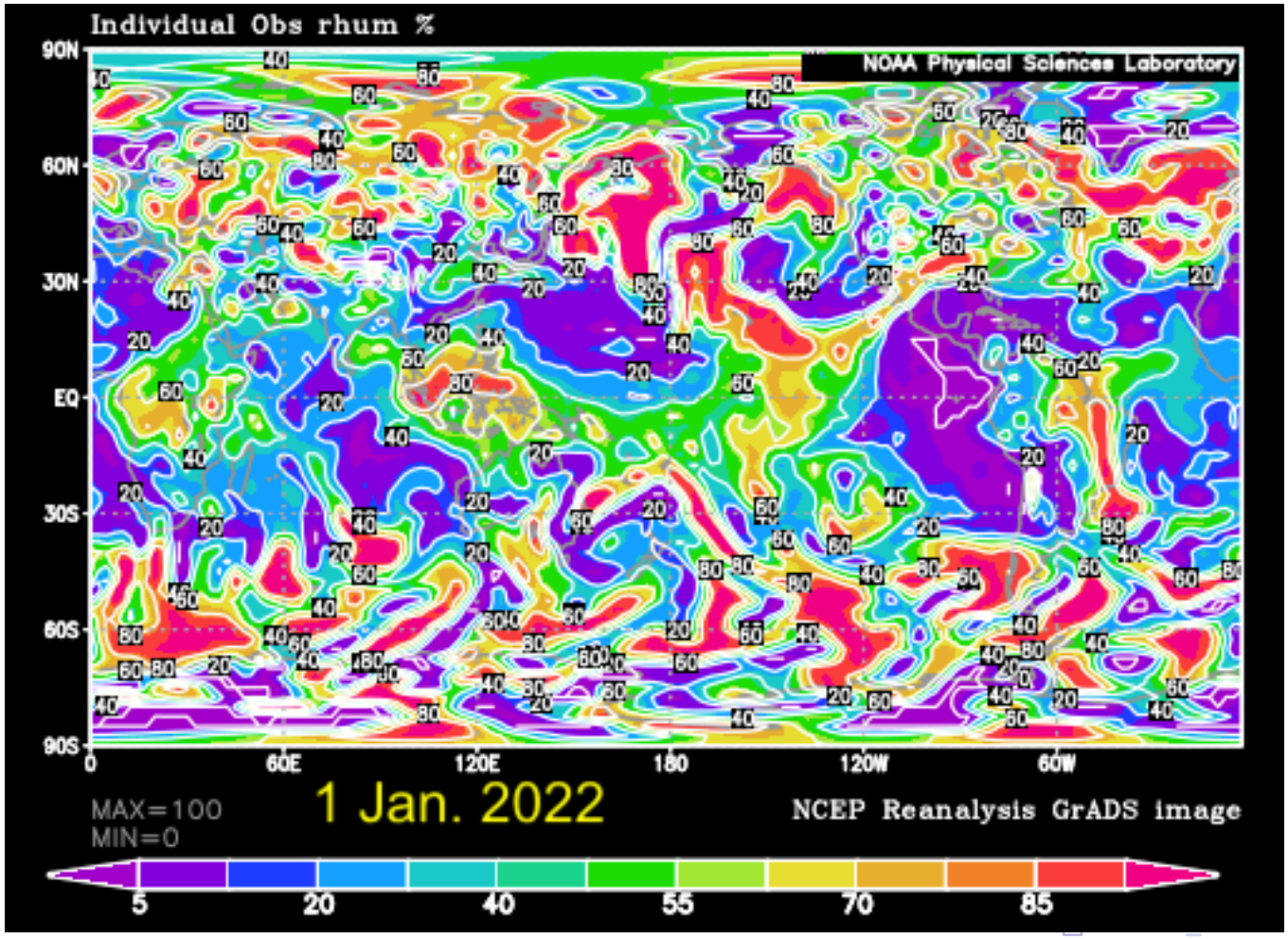
1 bar N_2/O_2 background atmosphere





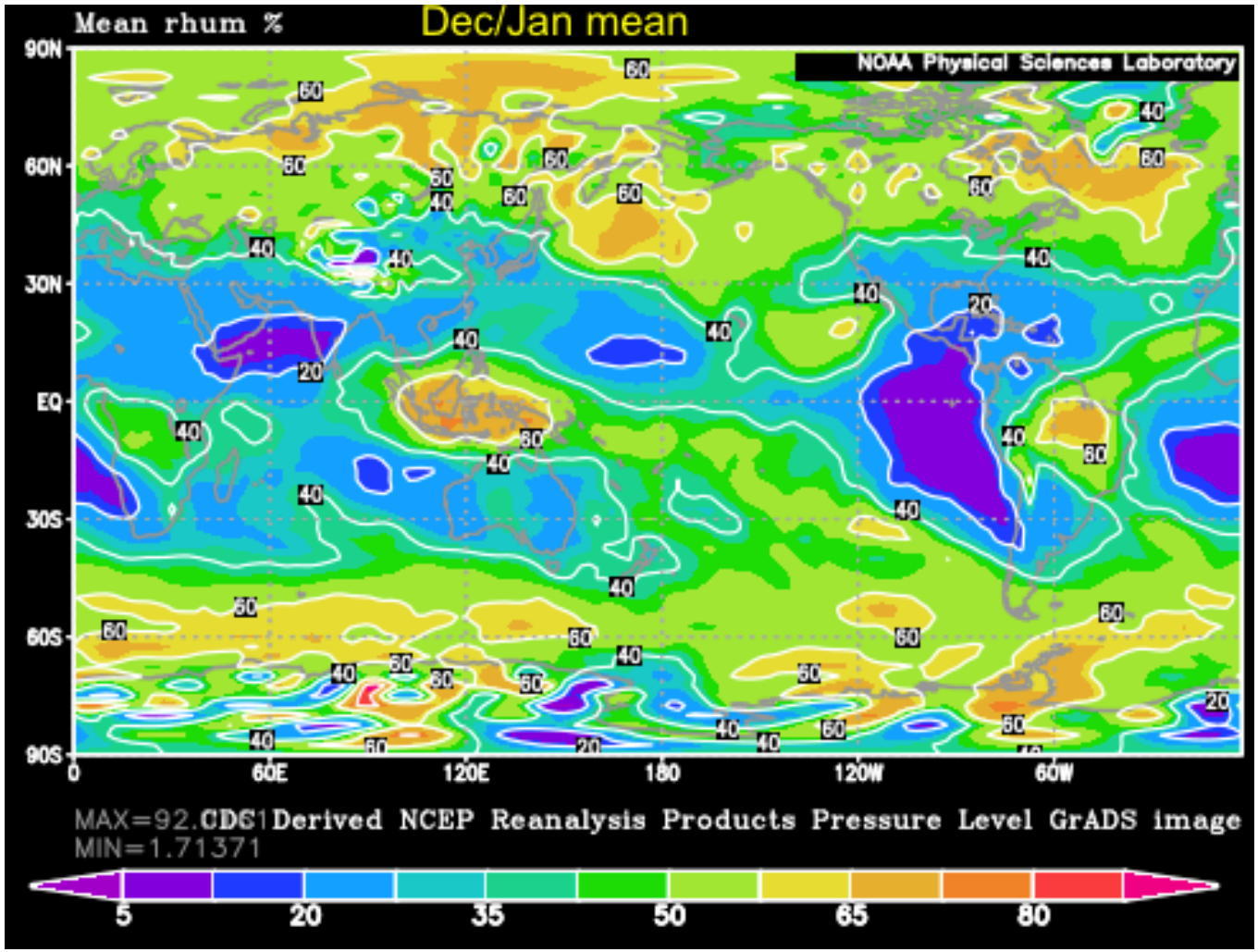
The atmosphere is considerably undersaturated

500mb relative humidity snapshot (100% = saturated)



The atmosphere is considerably undersaturated

500mb relative humidity Dec/Jan average (100% = saturated)



How to deal with relative humidity?

- In 3D general circulation models, relative humidty is *computed*
- For use in simplified modeling, an assumption that relative humidity remains fixed adequately captures behaviour, at least over moderate temperature ranges
- Nonetheless, the way relative humidity changes as the climate changes (e.g. through increasing CO_2 can significantly affect water vapour feedback.

The take-home point

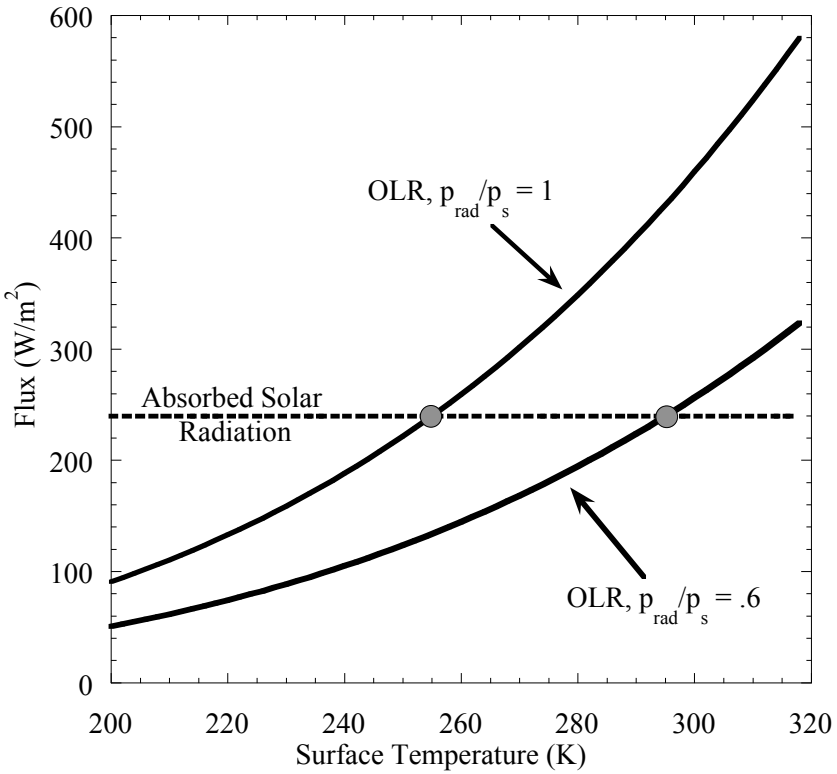
- For an atmosphere in contact with a surface condensed reservoir (ocean or glacier), water vapour is a *feedback*, not an imposed climate forcing.
- That's in contrast to CO_2 in Earthlike conditions, at least over time scales of a few millennia
- (On sufficiently long time scales CO_2 is actually also a feedback, with concentration determined by the carbon cycle)
- Water vapour feedback is a *fast feedback*, since it is determined by fast processes such as evaporation and condensation.
  (Equilibration time is on the order of weeks or less).

Reprise: The *OLR* curve and climate equilibrium

Suppose we have a theory that relates $T(p)$ to the surface temperature T_s, e.g. dry adiabat, pure radiative equilibrium or mix of the two.

Then, from Schwartschild, we can compute $OLR(T_s)$. Increase of greenhouse gas concentration lowers the $OLR(T_s)$ curve, and shifts equilibria to warmer temperatures

Radiative forcing for greenhouse gases



Radiative forcing is the amount by which the *OLR* curve is shifted downward at a given T_s (the unperturbed equilibrium), when the greenhouse gas amount is increased with everything else held fixed.

Perturbed energy balance

Suppose OLR depends on a parameter λ (e.g. CO_2 concentration).
Suppose $T(p)$ can be computed uniquely from surface temp T_s.

$$0 = G \equiv OLR(T_s, \lambda) - (1-\alpha)\frac{L_\circledast}{4}$$

$$0 = \frac{d}{d\lambda} G = (\partial_{T_s} G)(\partial_\lambda T_s) + \partial_\lambda G$$

$$\partial_\lambda T_s = -\frac{\partial_\lambda G}{\partial_{T_s} G} = -\frac{\partial_\lambda OLR}{\partial_{T_s} OLR}$$

$-\partial_\lambda OLR$ is the *differential radiative forcing*

Computed with $T(p)$ and everything else held fixed

$\partial_{T_s} OLR$ is the *climate sensitivity parameter*

Feedback Analysis

Suppose OLR depends also on a parameter r (e.g. water vapour content), which in turn is a function of T_s.

$$OLR = OLR(T_s, \lambda, r(T_s))$$

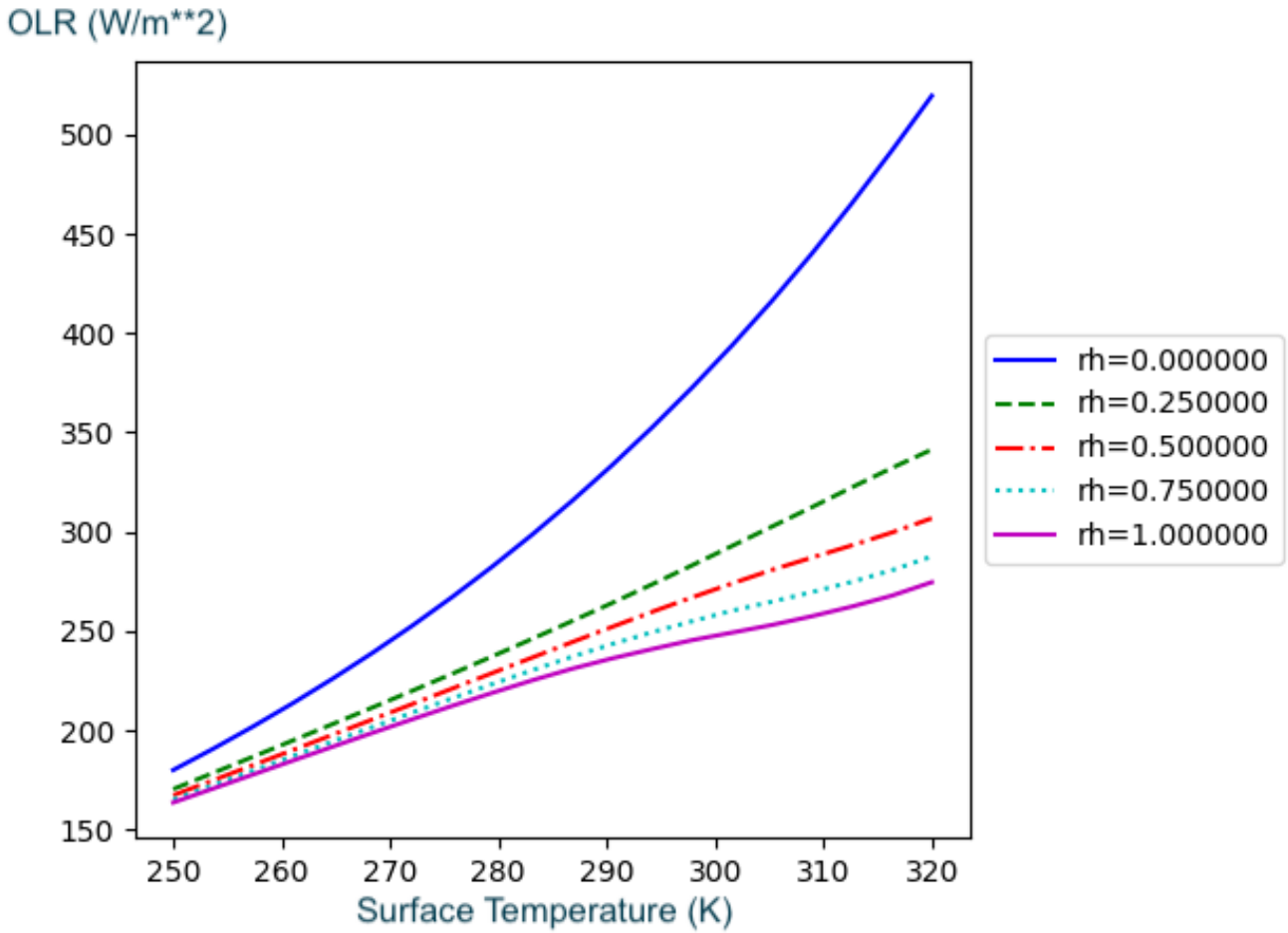
$$0 = G \equiv OLR(T_s, \lambda, r(T_s)) - (1-\alpha)\frac{L_\circledast}{4}$$

$$0 = \frac{d}{d\lambda} G = (\partial_{T_s} G|_\lambda)(\partial_\lambda T_s) + \partial_\lambda G$$

$$\partial_{T_s} G|_\lambda = \partial_{T_s} G|_{\lambda,r} + \partial_r G|_{\lambda,T_s} \partial_{T_s} r = \partial_{T_s} G|_{\lambda,r} \left(1 + \boxed{\frac{\partial_r G|_{\lambda,T_s}}{\partial_{T_s} G|_{\lambda,r}} \partial_{T_s} r} \right)$$

Effect of water vapour on OLR curves

CO_2 held fixed at 400 ppmv



Motivation

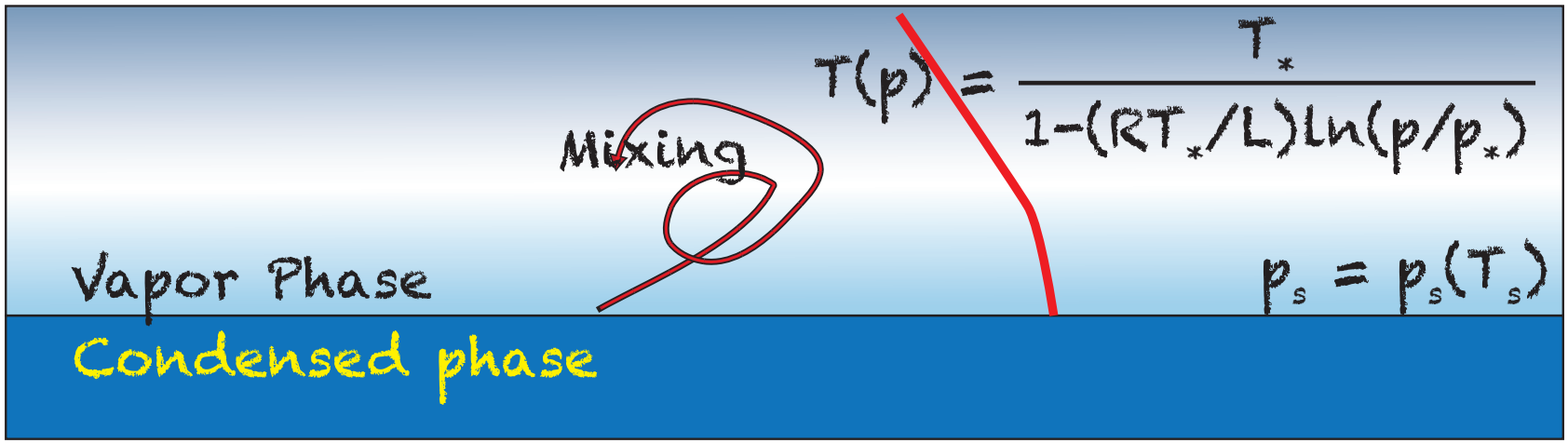
- How did Venus wind up with a 100 bar CO_2 atmosphere containing only traces of water?
- Inner edge of the (nominal) habitable zone:
  How much instellation can a planet get (i.e. how close to star can it be) and still support a liquid water ocean at the surface?

The pure steam limit

- In this lecture, we will discuss the physics of the runaway greenhouse for a single-component pure water vapour (steam) atmosphere
- This isn't actually much of a limitation, since any atmosphere will become steam-dominated if it gets hot enough, provided you don't run out of ocean first.
- e.g. if the atmosphere has a 1 bar N_2 noncondensible component, by the time the surface temperature is 400K, the saturation partial pressure of water is 3.2 bars, and the atmosphere s mostly water vapour.
- By the time the critical point temperature (647.1K) is reached, 221 bars of water vapour has gone into the atmosphere (if you haven't run out of ocean first), dominating anything but an extremely massive noncondensible background atmosphere

Limiting case I: Moist pseudo-adiabat for a single-component atmosphere

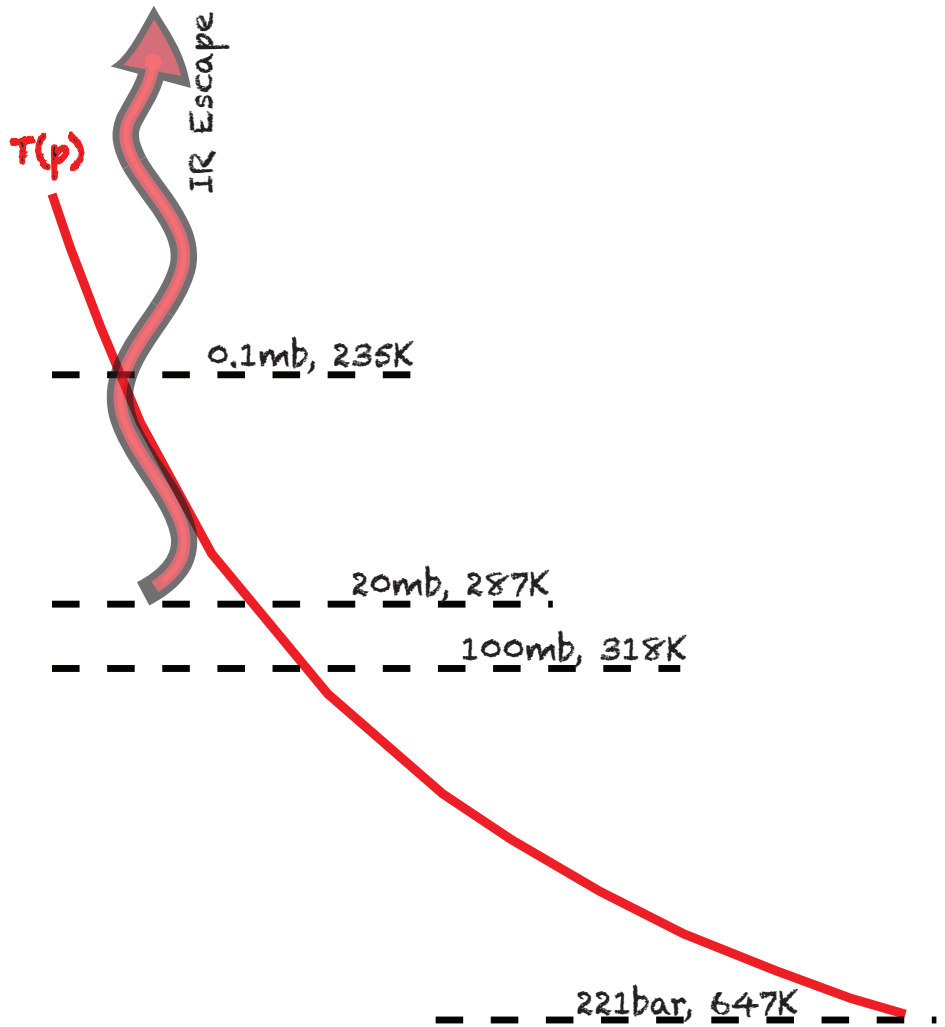
Solve Clausius-Clapeyron for $T(p)$. This is also the *dew point* (or *frost point*) formula. For a single component saturated atmosphere total pressure p is same as p_{sat}.



(constant L approximation)

Mass of atmosphere determined by surface temperature!

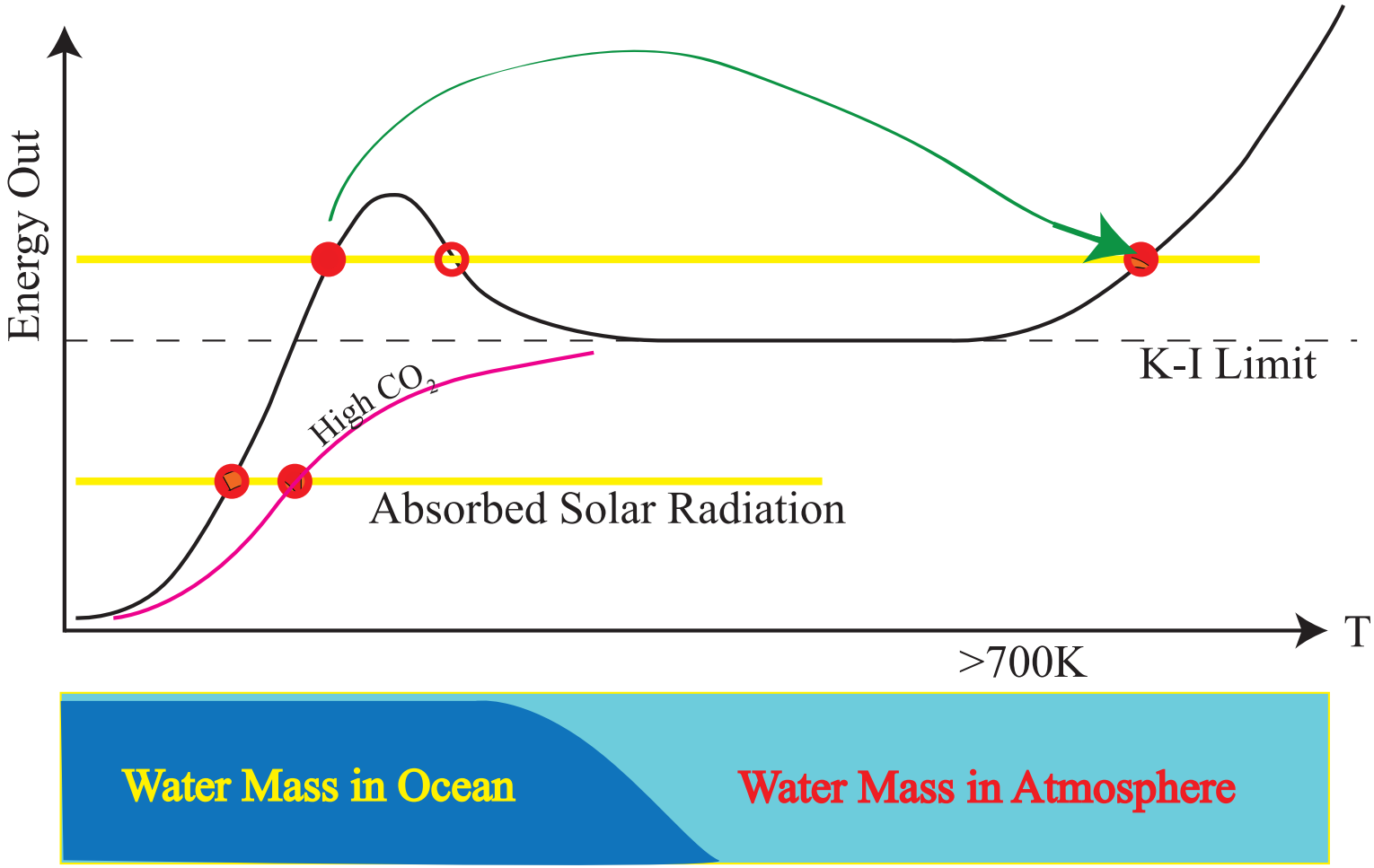
Limiting *OLR* for a steam atmosphere



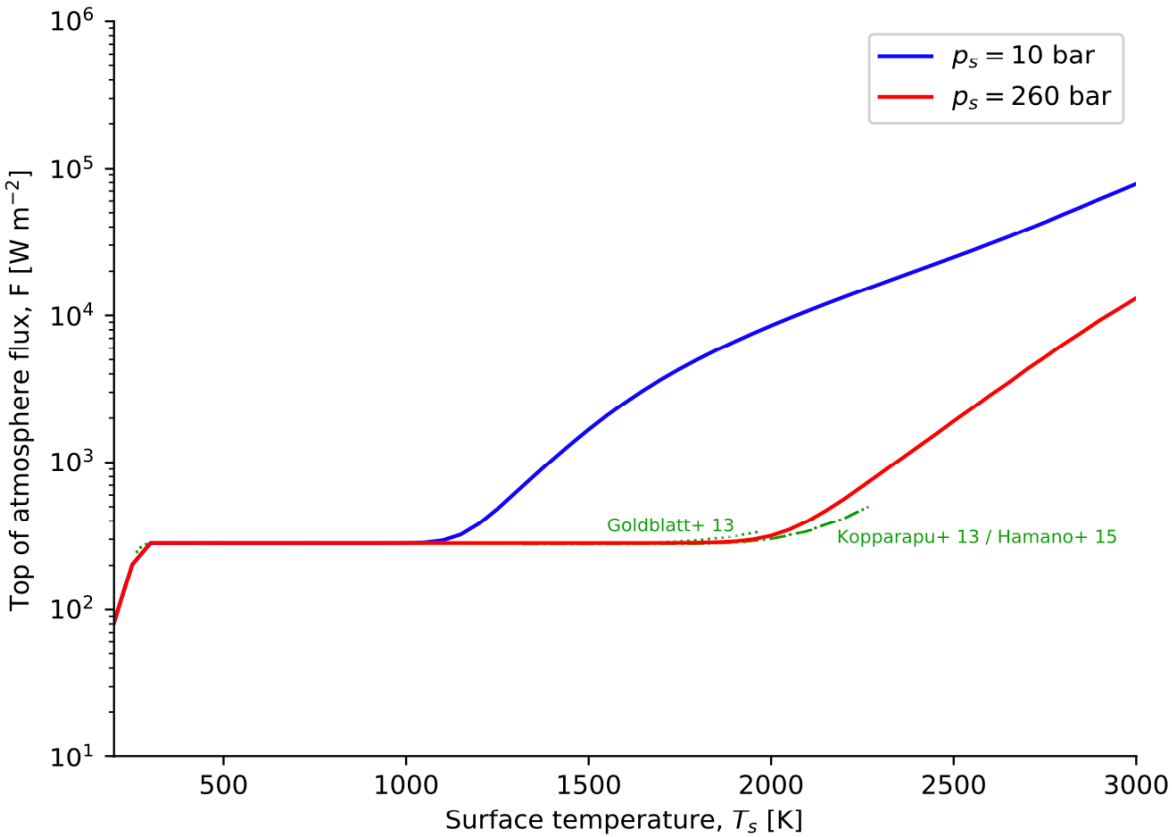
Grey-gas estimate of OLR_∞

Derivation on BLACKBOARD

Equilibration of the runaway



A real-gas calculation, pure steam



Boukrouche *et al* 2021

The aftermath: Irreversible water loss

