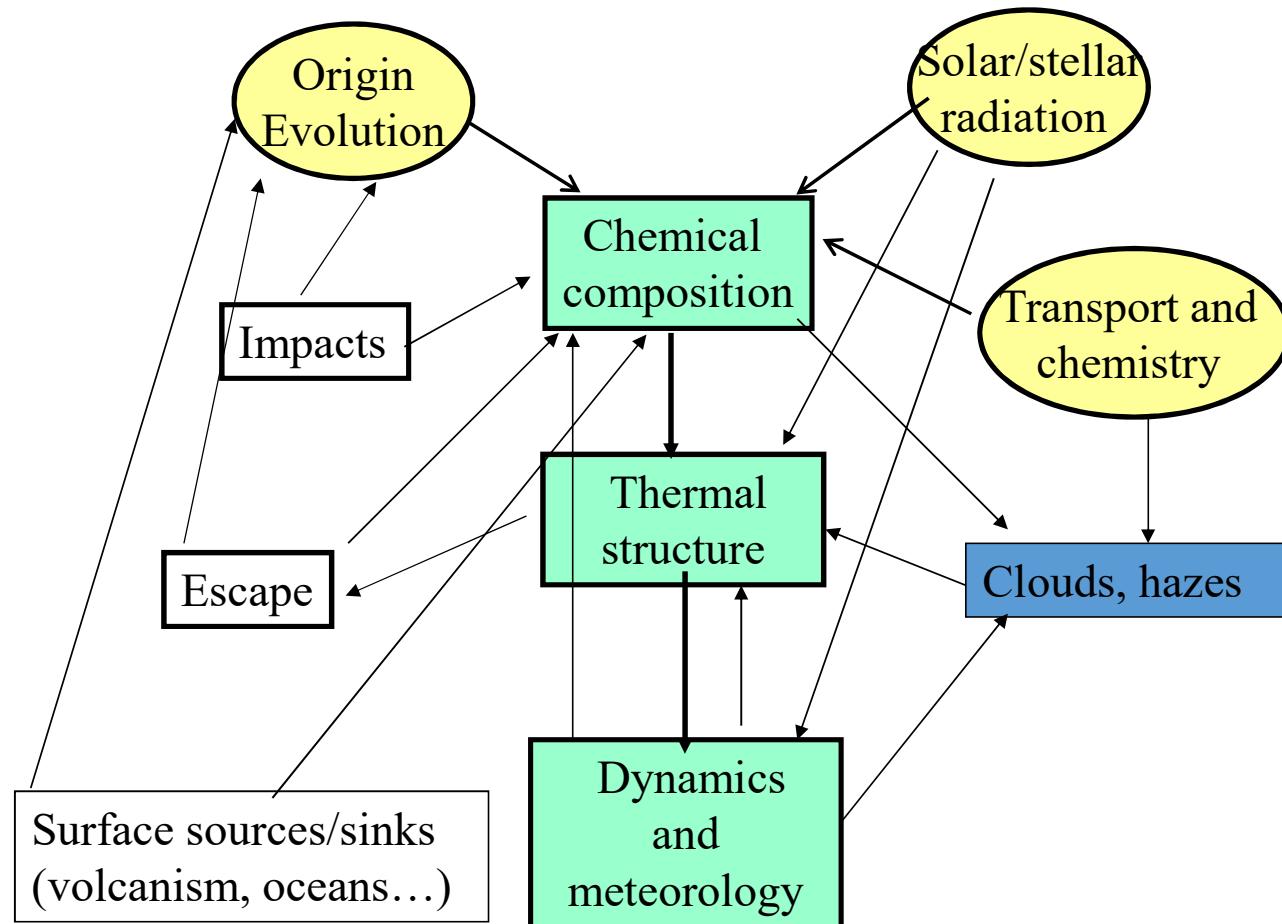


Planetary atmospheres

Emmanuel Lellouch
(LIRA, Observatoire de Paris)

Planetary atmospheres

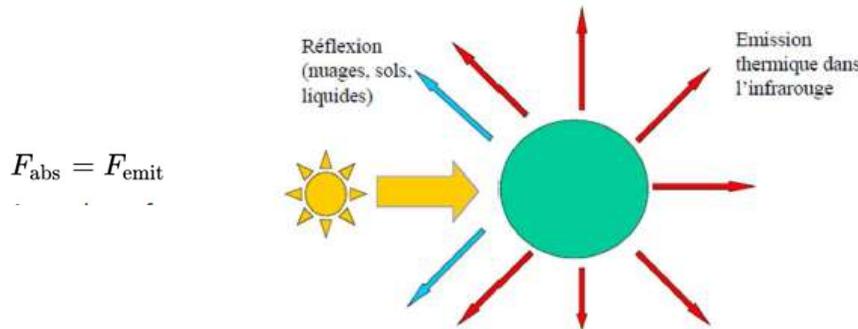


(Exo)-Planetary Atmospheres

Contents

- 1) basic concepts, nomenclature
- 2) diversity of atmospheres
- 3) effects on solar/stellar radiation
- 4) vertical transport, photochemistry, thermochemistry
- 5) methods of investigation
- 6) radiative balance, vertical structure, greenhouse effect

Basic concepts (1): Equilibrium temperature (1)



- **Equilibrium temperature T_{eq} = planet's temperature in equilibrium with solar/stellar radiation (theoretical)**

- Absorbed radiation = re-emitted radiation

$$\pi R^2 (1 - A_b) F_* = 4 \pi R^2 \sigma T_{\text{eq}}^4$$

where F_* = stellar flux (irradiance, W m^{-2}), A_b = Bond albedo, R = planet radius, σ = Stefan-Boltzmann constant
= $5,67 \cdot 10^{-8} \text{ W m}^{-2} \text{ K}^{-4}$

- Solar System: $F_* = F_{\text{sol}} = S / D^2$ where $S = 1368 \text{ W m}^{-2}$ = solar constant, and D is in au

$$\sigma T_{\text{eq}}^4 = (1 - A_b) F_{\text{sol}} / 4$$

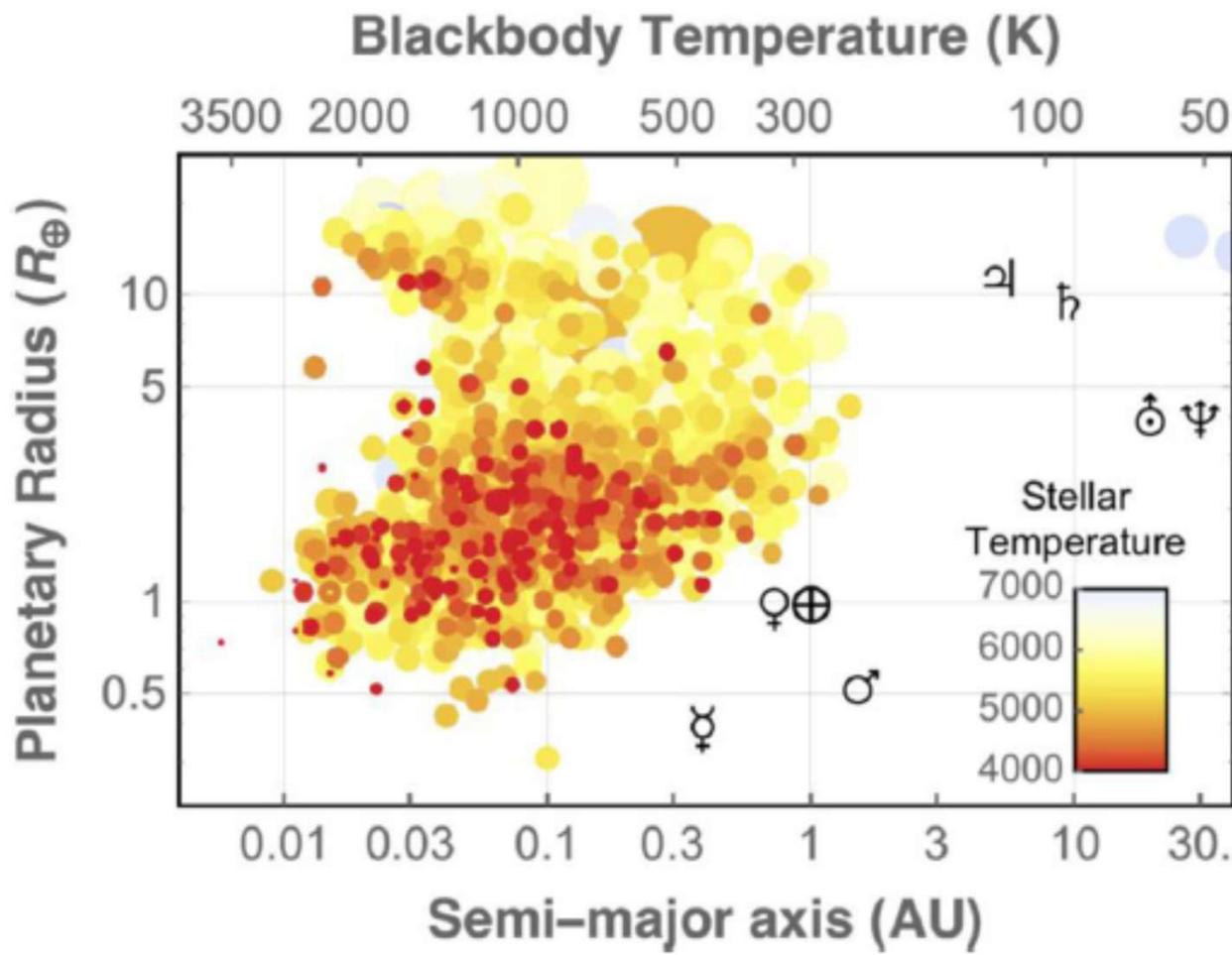
- Alternatively, $F_* = 4\pi R_{\text{star}}^2 \sigma T_{\text{star}}^4 / 4\pi a^2$, where a = planet-star distance and T_{star} = star temperature.

Show that:

$$T_{\text{eq}} = T_{\text{star}} \sqrt{\frac{R_{\text{star}}}{2a}} (1 - A_B)^{1/4}$$

e.g. KELT-9b $T_{\text{star}} \sim 10000 \text{ K}$, $a = 0.035 \text{ au}$
 $T_{\text{eq}} \sim 4000 \text{ K}$!

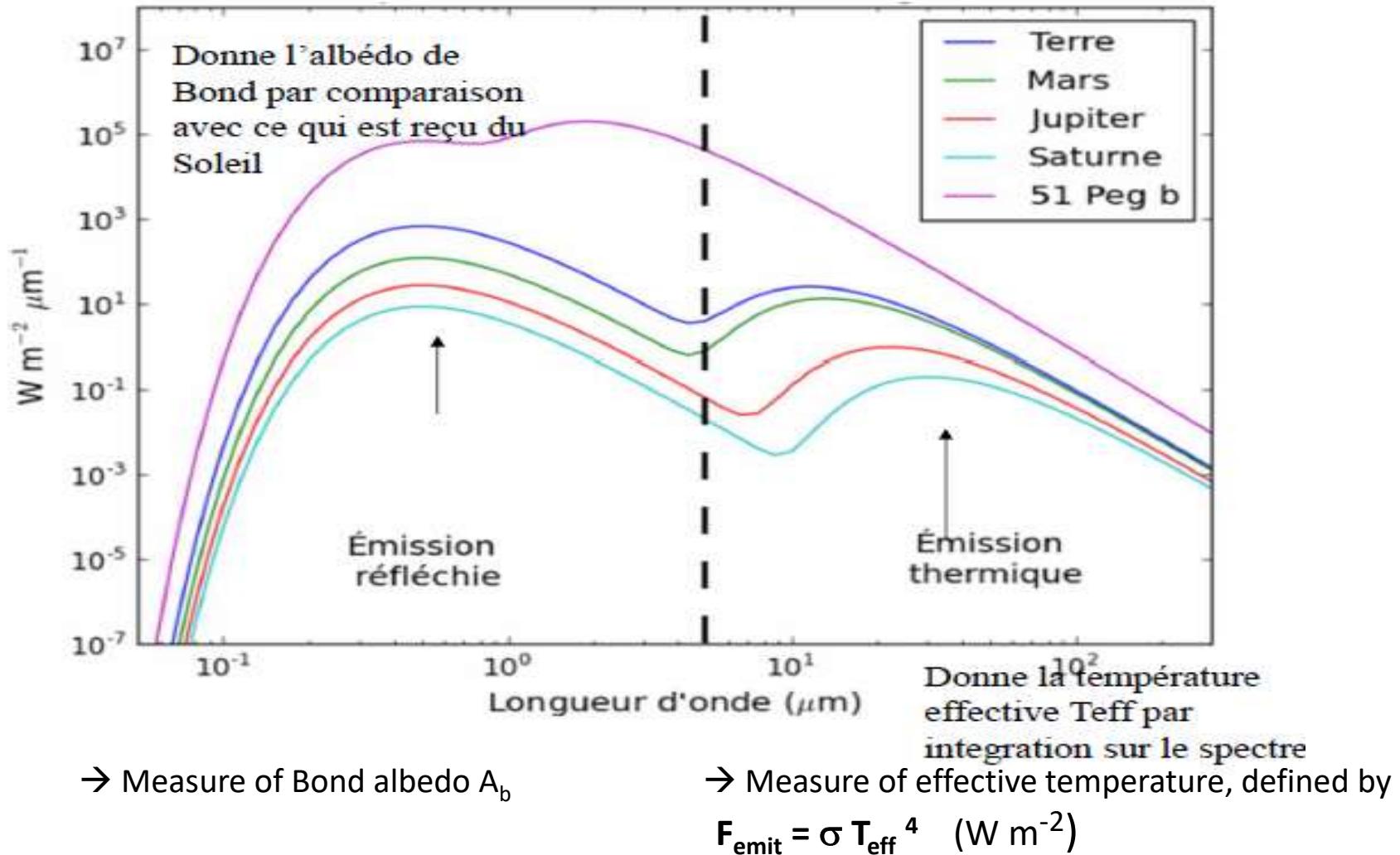
Exoplanets: equilibrium temperatures



In general
 A_b are not known
and assumed to be 0
here

Equilibrium temperatures are a measure of irradiation

Two component spectrum of a planet: schematics



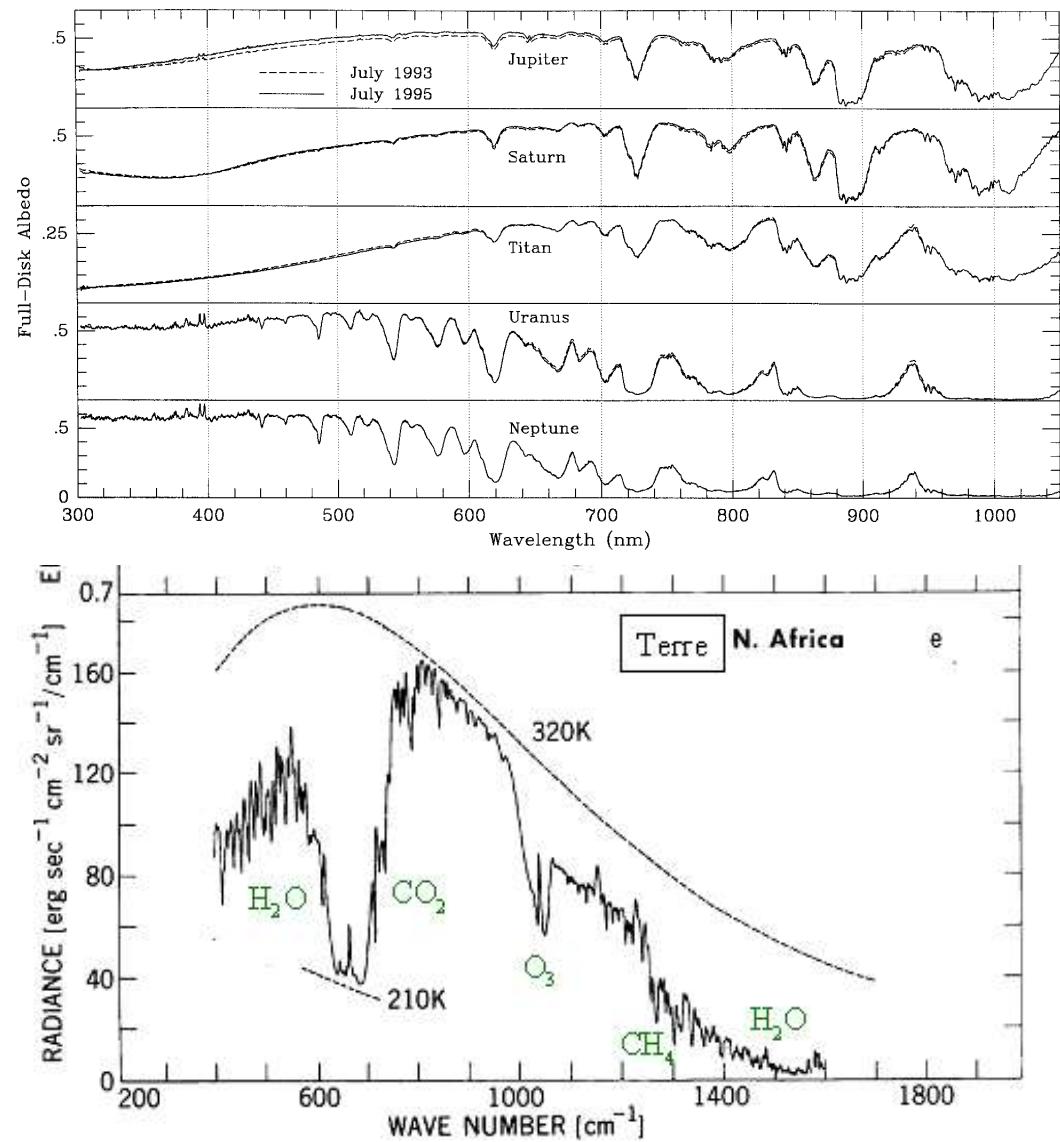
Real planetary spectra in the Solar System

Observational determination
of Bond albedo

and of

effective temperature

Effective temperature is a measure of thermal emission



Solar System planets: albedos and temperatures

	Bond albedo	Equilibrium Temperature (K)	Effective temperature (K)	Ground temperature (K)
Venus	0.75	231 !	231	735 !!
Earth	0.30	255	255	288
Mars	0.31	210	210	214 !?
Jupiter	0.343	109.5	124.4	165 * (1bar)
Saturn	0.342	82.3	95	135* (1bar)
Titan	0.29	82	82	94
Uranus	0.300	58.2	59.1	76* (1bar)
Neptune	0.31	45.9	59.3	72 * (1bar)

Calculate internal energy source in giant planets; and comment for a hot Jupiter at 0.05 au from a solar-type star

Basic concepts (2): atmospheric escape

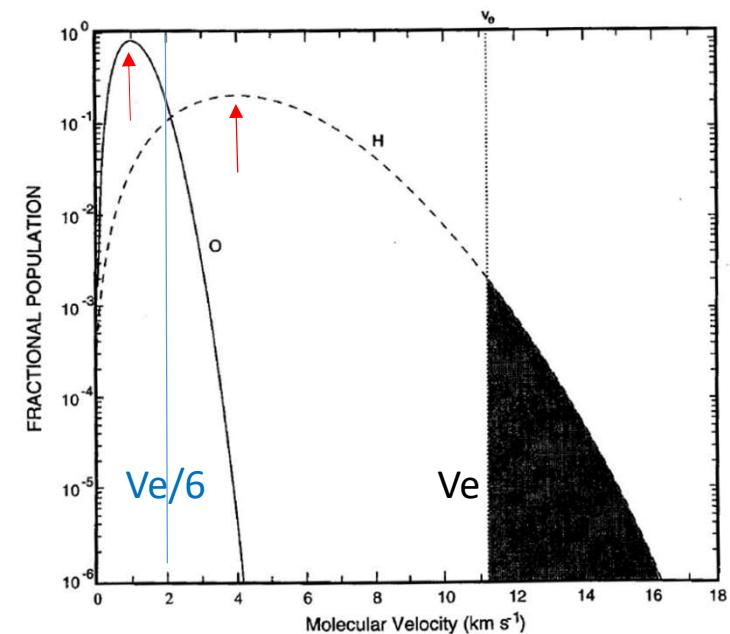
- **Escape velocity v_e** : $v_e = \sqrt{\frac{2GM_{pl}}{R_{pl}}} = \sqrt{2g_s R_{pl}}$ where g_s = surface gravity

- **Maxwell-Boltzmann distribution**

$$f(v) = 4\pi v^2 G(v) = \sqrt{\frac{2}{\pi}} \left(\frac{m}{k_B T} \right)^{3/2} v^2 e^{-\frac{mv^2}{2k_B T}}.$$

- **Thermal speed v_{th} (most probable speed in MB distribution)**

$$v_{th} = \sqrt{\frac{2kT}{m}}$$



- (Much) too simplistic approach: thermal escape important if $v_{th} > v_e$
- More realistic approximation: thermal escape important if $v_{th} > \sim v_e / 6$

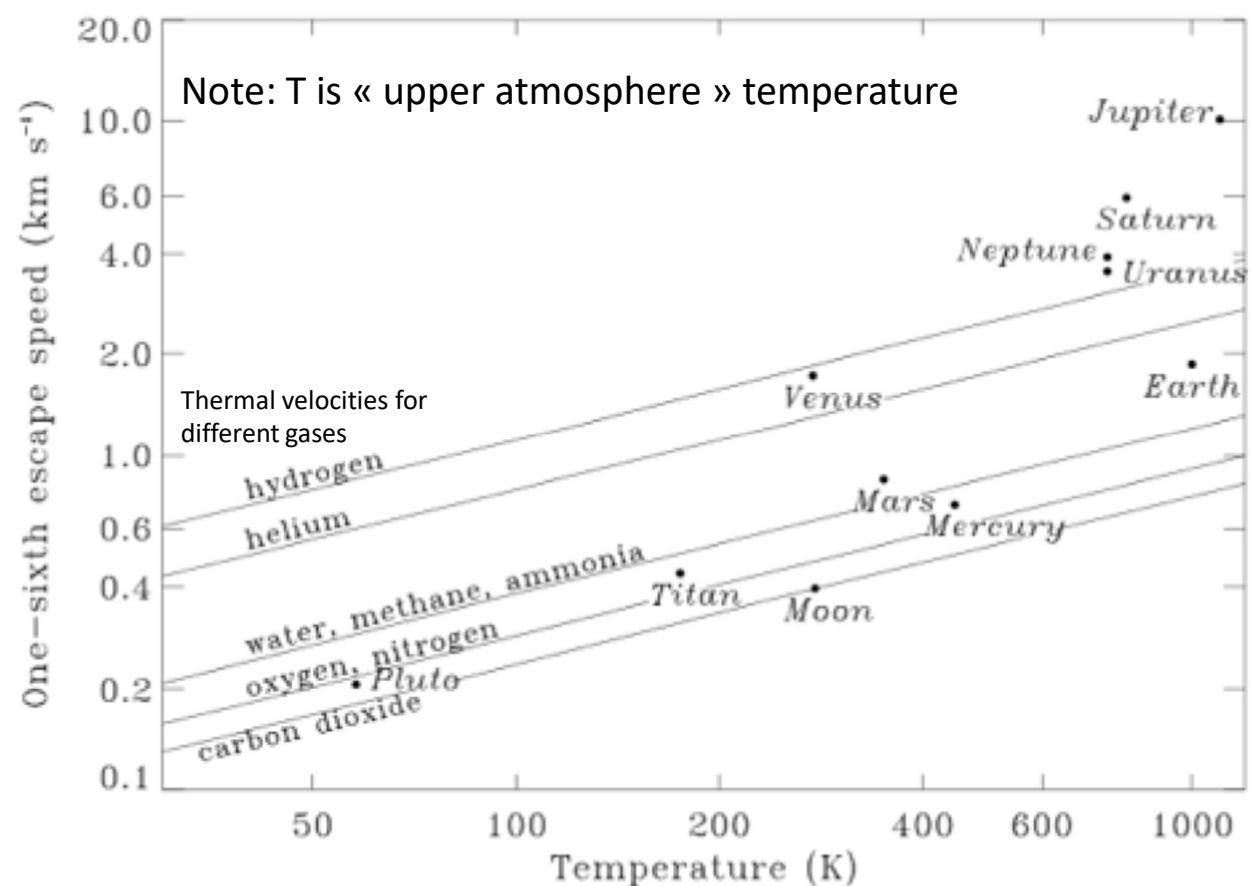
Basic concepts (2): atmospheric escape

More realistic approximation: thermal escape important if $v_{th} > \sim v_e / 6$

- Elementary comparisons

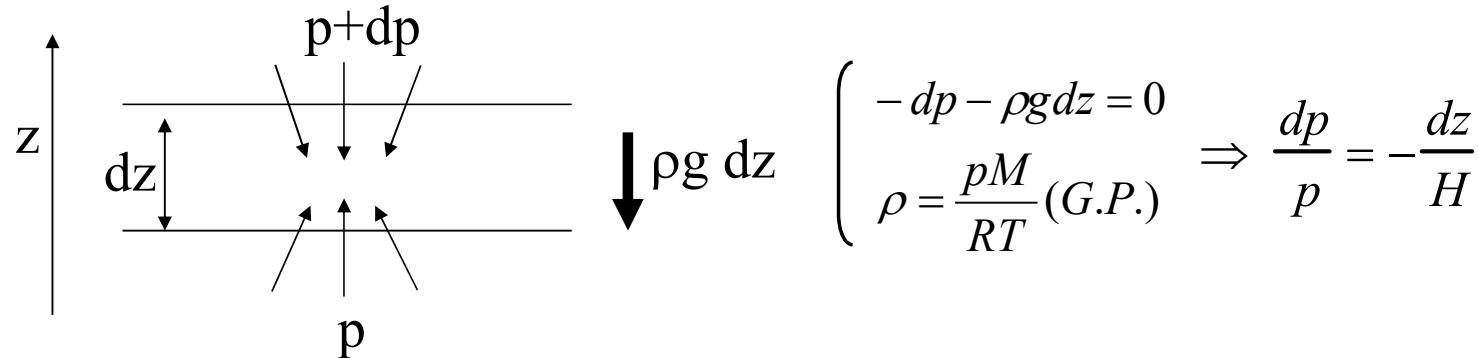
Earth/Moon, Mars/Mercury, H₂/O₂ on Earth

- Still simplistic because there are other escape processes...**



Basic concepts (3): Hydrostatic equilibrium

- Plane parallel approximation



- Scale height

$$H = \frac{RT}{Mg} = \frac{kT}{mg} \quad \text{where } R=8,3145 \text{ J mol}^{-1} \text{ K}^{-1}$$

$\rightarrow p = p_0 e^{-\int_{z_0}^z dz / H}$ (1) ; if T is constant $\rightarrow p = p_0 e^{-\frac{(z-z_0)}{H}}$ (2)

	Jupiter	Saturn	Uranus	Neptune	Venus	Mars	Earth	Titan	Triton	Pluto	Io	HD209458b (1.32 RJ, 0.69 MJ)
Main gas	H ₂ , He	CO ₂	CO ₂	N ₂ , O ₂	N ₂	N ₂	N ₂	SO ₂	H ₂ , He			
M (g)	2,3	2,3	2,6	2,5	43,4	43,3	29,0	27	28	28	64	2,3
g (m s ⁻²)	24	10	9	11	8,9	3,7	9,8	1,35	0,8	~0,7	1,8	10
T _{surf} (K) ^a	165	135	76	72	735	214	288	94	38	~40	110	1130 (Teq)
H (km)	25	48	27	22	16	11	8,4	21	14	~18	7,9	~400

a : 1 bar for Giant Planets

Column density: definition and expression (exercise)

$$a(z) = \int_z^{\infty} n(z') dz' : \text{number of molecules above some altitude } z \text{ per unit area nb de molécules/surface (mol cm}^{-2}\text{)}$$

Demonstrate that $a(z)$ is proportional to $p(z)$ and calculate value at Earth's surface.

Calculate Earth atmosphere mass and compare to oceans.

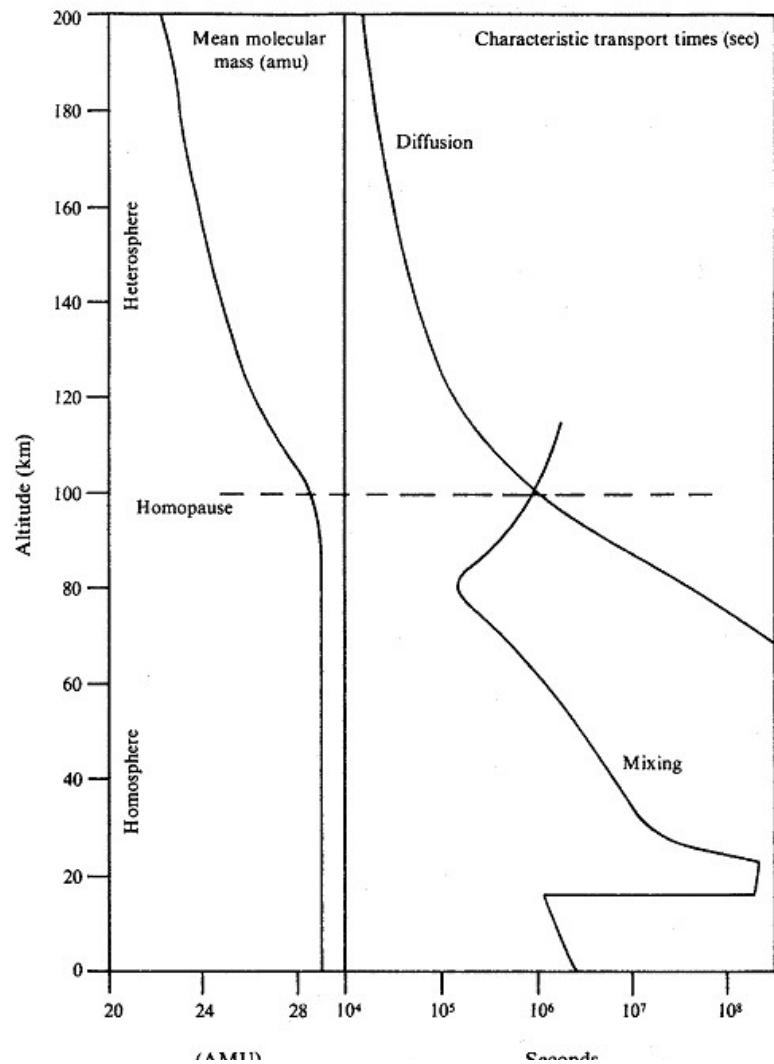
Gravitational segregation

- Hydrostatic equilibrium: $H = RT/Mg = kT/mg$. What is m ? One m per molecule? H_i ?
- **Microscopic point of view:** why an equilibrium given that molecules have thermal velocities and that they are subject to gravitational field?
- Thermal velocities are randomly oriented and re-initiated at each collision
 - Mean free path = mean distance between 2 collisions = (exercise)
$$l = 1 / (n Q)$$
, n = concentration, Q = collision cross section $\sim 3 \cdot 10^{-15} \text{ cm}^2$
- Downward flux of molecule i falling in gravitational field
 - Friction exerted by collisions \rightarrow vertical motion has a limiting **velocity** $w = g \tau$, where $\tau = l / v_{th}$ = time between collisions.
Associated flux (downward) = $-w n_i$, tending to enhance molecule concentration at low altitudes
- Concentration gradient \rightarrow diffusion flux in direction opposed to gradient, counteracts gravitational separation
$$\Phi_i = -D_i dn_i / dz$$
 (upward), D_i = **diffusion coefficient** ($\text{cm}^2 \text{ s}^{-1}$) . It can be shown that $D_i = 1/2 l v_{th}$ $v_{th} = \left(\frac{2 k T}{m} \right)^{1/2}$
- Total flux : $F = -w n_i - D_i dn_i / dz$
- At equilibrium: $F=0 \rightarrow dn_i^e / n_i^e = - (w / D_i) dz = - 2 g / v_{th}^2 dz = -m_i g / (kT) * dz = -dz / H_i$
- **DIFFUSION EQUILIBRIUM:** each gas has its own scale height H_i

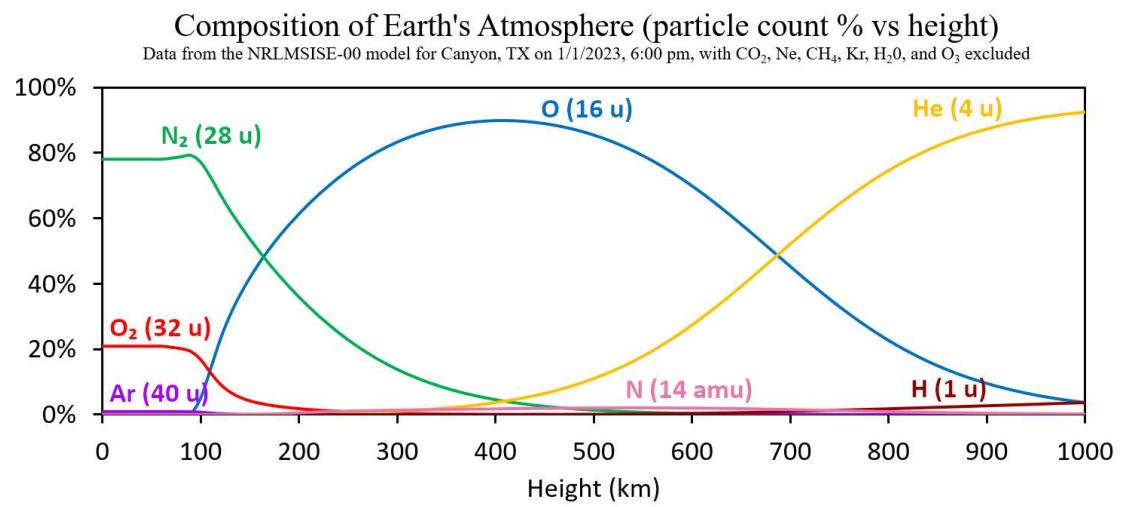
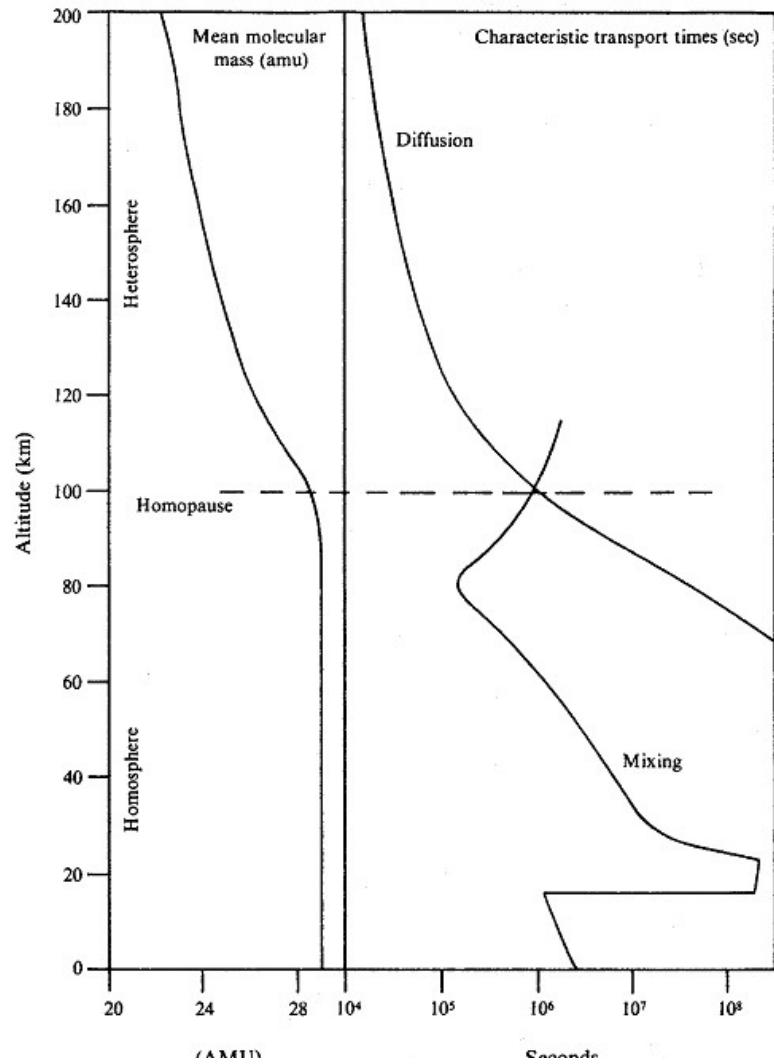
Gravitational separation vs atmospheric mixing

- However, in addition to molecular diffusion, the atmosphere can also be mixed by turbulent motions (vertical or horizontal) – bottle (water+oil) analogy
 - Typical length L of mixing depends on mechanism for turbulence but can be from ~ 1 m to \sim scale height. Mixing characterized by eddy diffusion coefficient K (also in $\text{cm}^2 \text{ s}^{-1}$): $K \sim L v_{\text{eddy}}$, where v_{eddy} is characteristic turbulent velocity
- Transport timescales for molecular diffusion and eddy mixing
 - Over 1 scale height: $\tau_d = \frac{H^2}{D_i}$ $\tau_K = \frac{H^2}{K}$ $\tau_d \propto \frac{l}{v_{th}}$ $\tau_K \propto \frac{L}{v_{\text{eddy}}}$
 - If $\tau_K < \tau_d$, turbulence mixes gases more efficiently than gravitation separates them.
 - Low altitudes (homosphere)
 - At Earth surface, $l = 10^{-7}$ m. D_i very small $\rightarrow \tau_K \ll \tau_d$. Gases are uniformly mixed (except for chemical reactions (e.g. ozone, cloud condensation...)). Hydrostatic equilibrium applies **in average**, with mean molar mass M
 - High altitudes (heterosphere)
 - The mean free path l increases as $1/n$ (e.g. at 100 km in Earth's atmosphere, $l = 1$ m), and so does D_i , and τ_d decreases accordingly. When $\tau_K \gg \tau_d$, diffusion equilibrium. Each gas has its own scale height.
 - General case: coexistence of diffusion and turbulent fluxes \rightarrow progressive change of regime (mixing \rightarrow diffusion).

Molecular diffusion and eddy mixing on Earth



Molecular diffusion and eddy mixing on Earth



Hydrostatic equilibrium at large distances

- Far away from surface: plane-parallel \rightarrow spherical symmetry, and g varies

For isothermal T

$$\nabla p + \rho g(r) = 0$$
$$p = p_s e^{\frac{r_s}{H_s} \left(\frac{r_s}{r} - 1 \right)}$$
$$g(r) = g_s \left(\frac{r_s}{r} \right)^2 \rightarrow \frac{dp}{p} = -\frac{1}{H_s} \left(\frac{r_s}{r} \right)^2 dr$$

- Atmospheric expansion

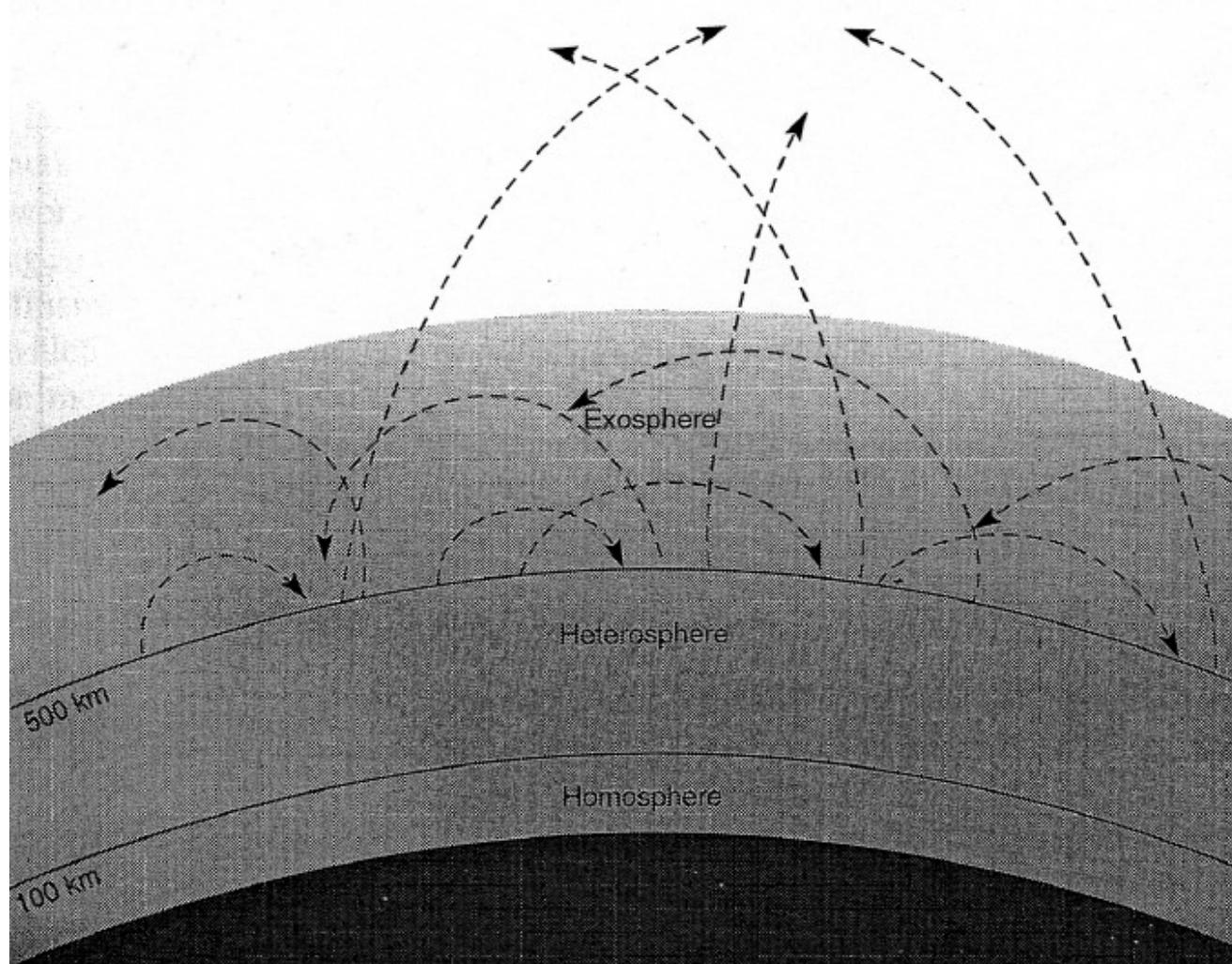
- The above leads to $p(r = \infty) = p_s e^{-r_s/H_s} \neq 0$! Obviously wrong
- Breakdown of hydrostatic equilibrium
- Happens because collisions are too rare to eliminate directional velocities
- Equation of motion becomes $\rho \frac{dv}{dt} = -\nabla p - \rho g > 0$ ($v > 0$ upwards)
- Ballistic and escape trajectories

- This happens when mean free path becomes comparable to scale height H

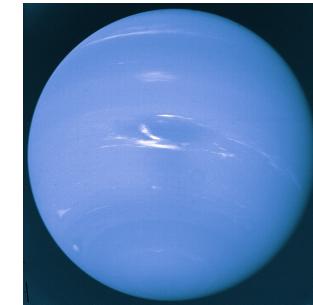
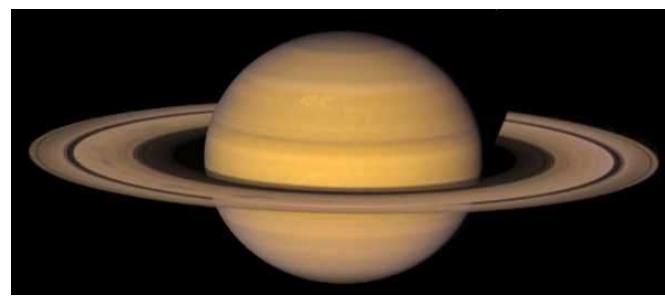
- Exobase $l = 1/(n_e Q) = H_e$. Exobase column density $N_e \sim n_e H_e = 1/Q \sim 3 \times 10^{14} \text{ cm}^{-2}$
- Earth $p_e \sim 10^{-11} \text{ bar}$, $T_e \sim 1000 \text{ K}$, $z_e \sim 500 \text{ km}$

\rightarrow Exosphere: upper limit of the atmosphere, where escape occurs

Homosphere, heterosphere and exosphere on Earth



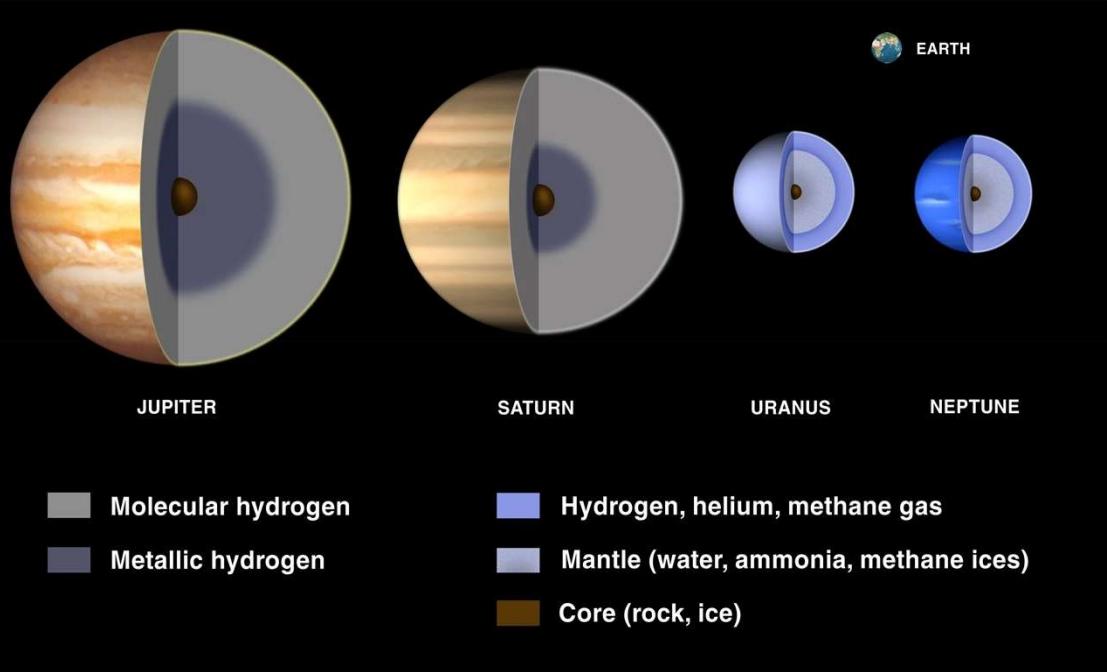
Diversity of Solar System atmospheres (1) Giant Planets



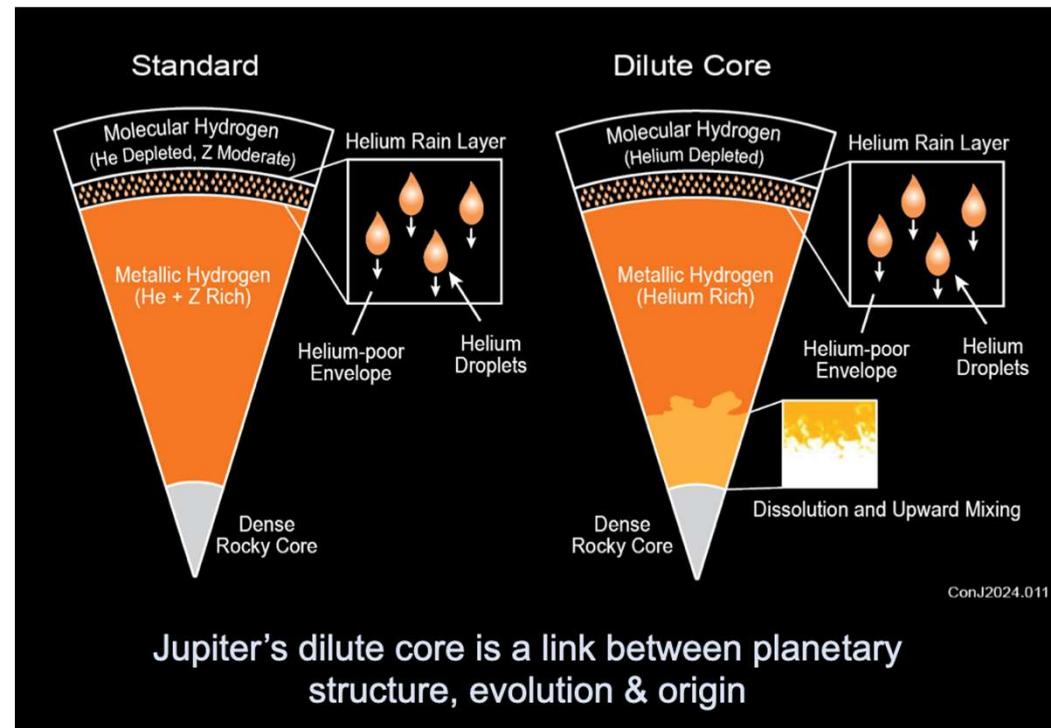
- Fluid, no surface interacting with atmosphere
- H₂ – He dominated
- Predomimantly chemically reducing (CH₄, H₂O, NH₃, H₂S...)
- Primitive (captured from protosolar nebula gas), and little evolved since formation (little escape, no interaction with surface, recycling of photolysis products)
- Measurable internal energy source ($T_{\text{eff}} > T_{\text{eq}}$), except Uranus
- Two classes
 - Gas giants (Jupiter and Saturn): most of mass is in H₂ et He (> 95%)
 - Icy giants (Uranus et Neptune) : most of ice in the form of ices and rocks (>80%)

Giant Planets internal structure

Classic view



Modern view (Jupiter, Saturn)

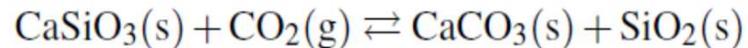


Diversity of Solar System atmospheres (2) Telluric planets

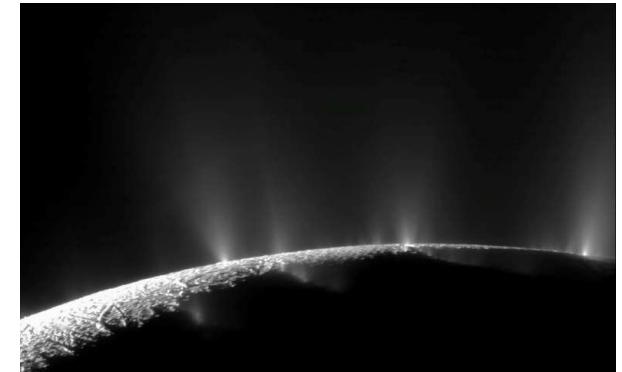
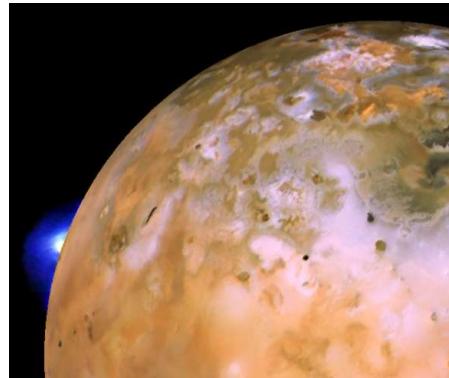


- Dominated by heavy compounds (N₂, O₂, CO₂, H₂O...)
- Strongly evolved : any primitive H₂-He atmosphere has been lost to escape
- **Secondary atmospheres**, formed after formation
 - Catastrophic outgassing of the volatiles (CO₂, CH₄, H₂O, N₂) present in the accreting planetesimals (~100 MYr)
 - Planetesimal impacts and melting → magma ocean phase → cooling → solidification of magma ocean and outgassing of volatiles
 - Continuous volcanic degassing of volatiles still present in mantle (geological timescales)
 - Enrichment by external sources (asteroid and cometary impacts)
- Strong interaction with the surface
 - Chemical equilibria of atmosphere with surface rocks: weathering
 - Example : trapping of CO₂ in surface / subsurface from weathering of silicates
 - Biogeochemical reactions; conversion of CO₂ into O₂ (Earth)
- **Long term diverging evolution: Venus: 90 bar, Earth: 1 bar, Mars: 6 mbar**

- Titan (N₂ + CH₄ atmosphere, 1.5 bar)
 - An initially NH₃ atmosphere produced from outgassing of planetesimals, later converted to N₂
 - Episodic outgassing of methane from “cold volcanoes”



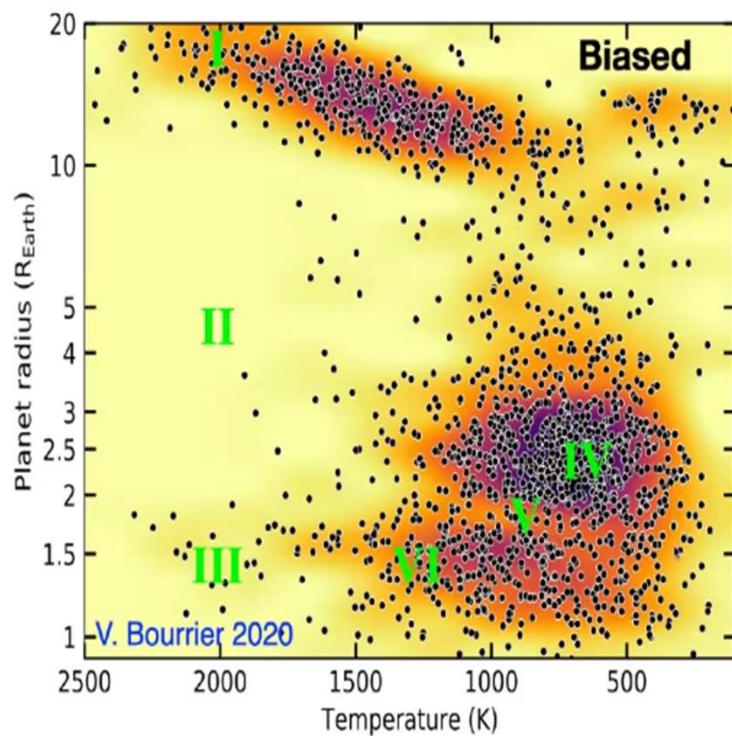
Diversity of Solar System atmospheres (3) Tenuous atmospheres



- 12 other “planetary size” objects Solar System bodies have detected atmospheres: Moon, Mercury, Ceres, Io, Europa, Ganymede, Europa, Callisto, Enceladus, Dione, Rhea, Triton Pluto, Dominated by heavy compounds (N_2 , CH_4 , O_2 , SO_2 , H_2O ...)
- Different classes : “true” atmospheres (Pluto, Triton, Io) vs. exospheres (most others)
- Weak gravitational field → importance of escape

- Immediate sources of these atmospheres
 - Sublimation of surface ices (N_2 on Pluto and Triton, SO_2 on Io)
 - Volcanism (Io) or cryo-volcanism (Enceladus)
 - Surface sputtering : impact of energetic particles (solar wind, micrometeorites, charged particles from magnetospheres) on surfaces

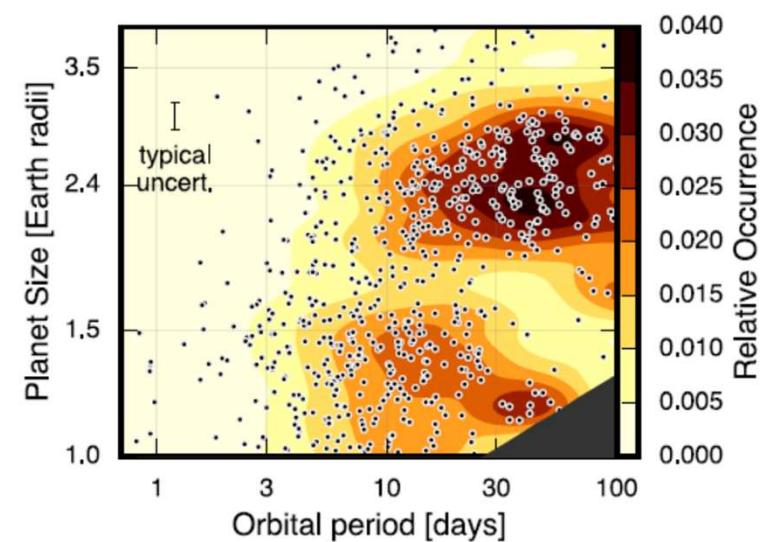
What to expect for exoplanets ?



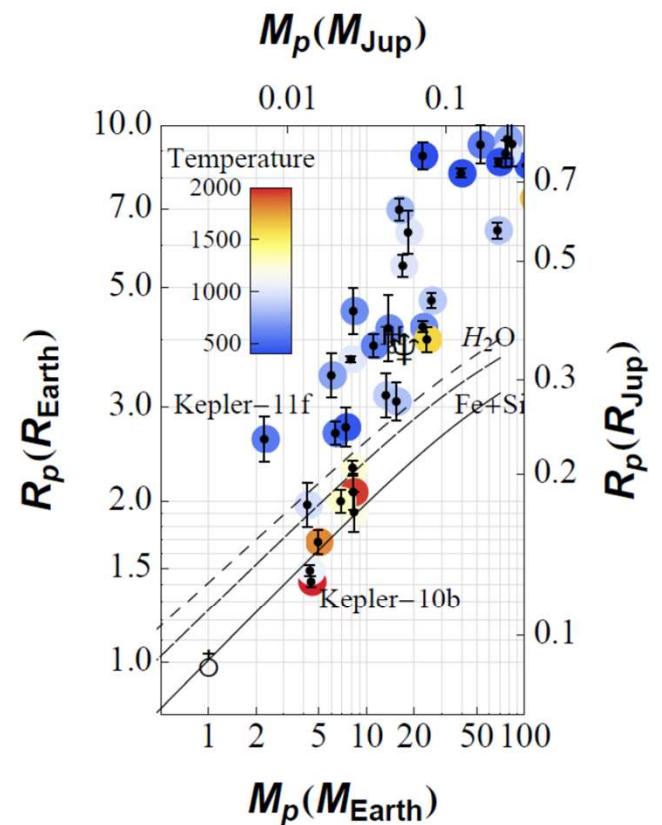
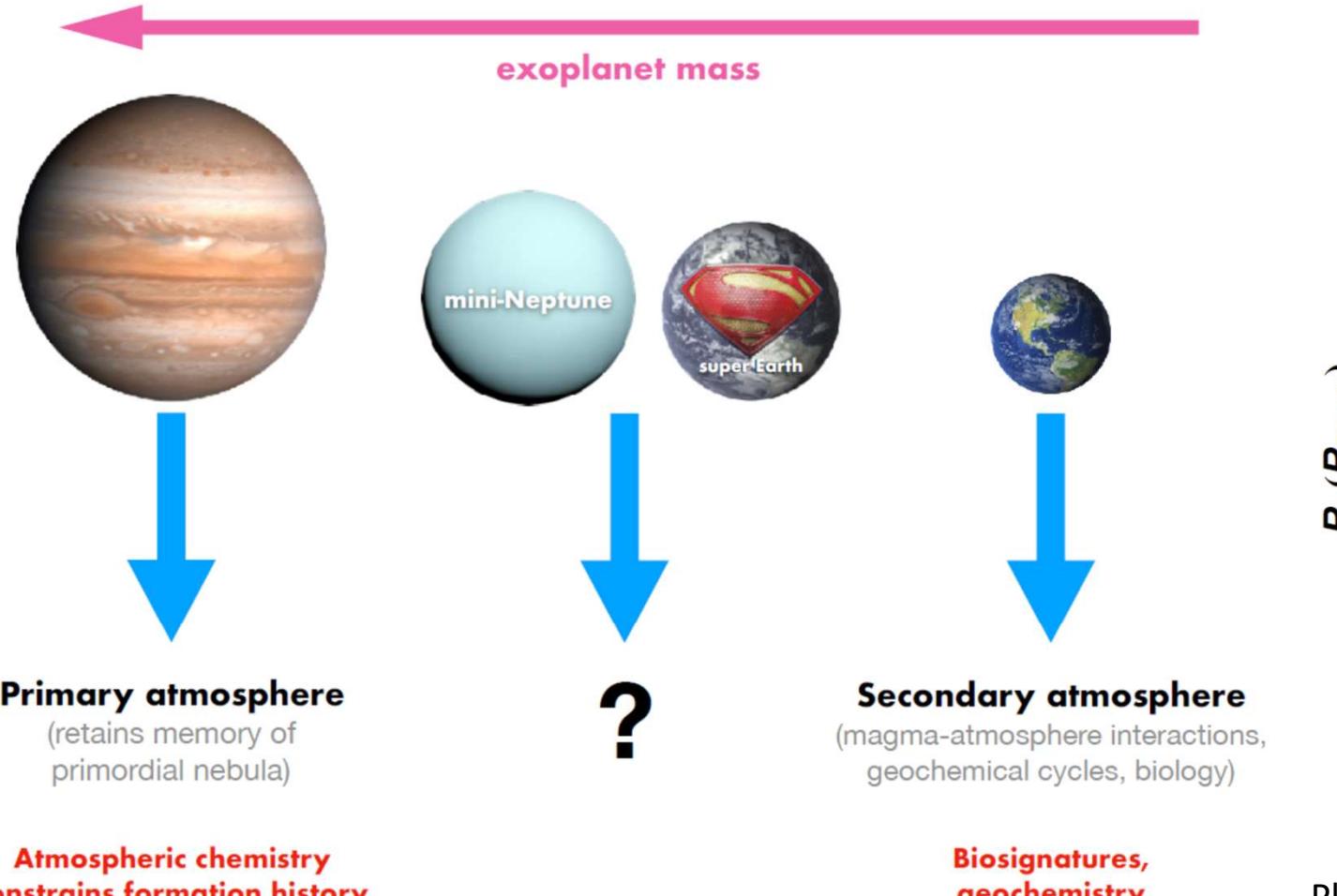
Best known sub-population (obs. bias).

Morphological features:

- I. Bloated hot Jupiters
- II. Sub-Saturnian desert
- III. Magma planets
- IV. Sub-Neptunes
- V. Radius valley
- VI. Super-Earths

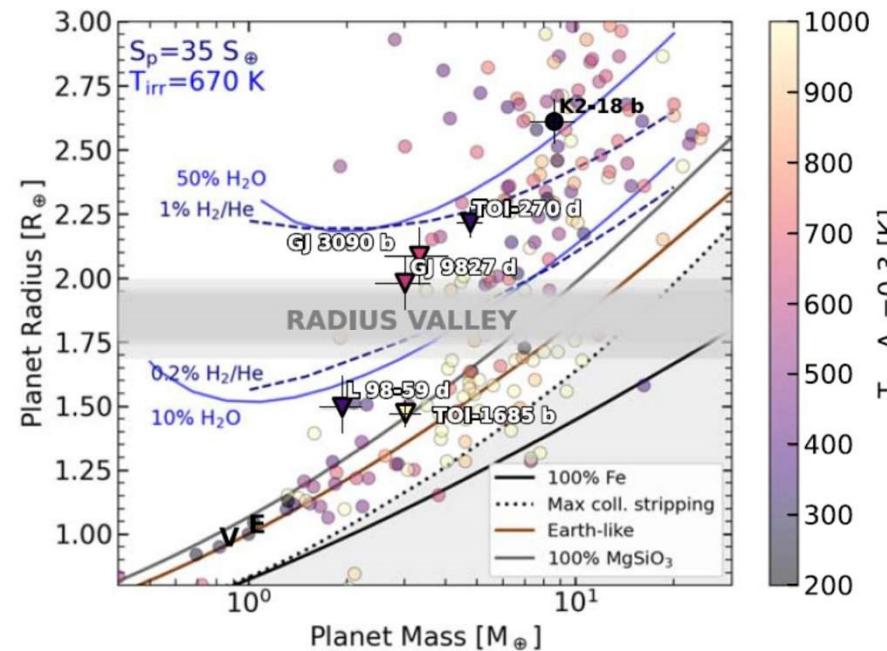


What to expect for exoplanets ?

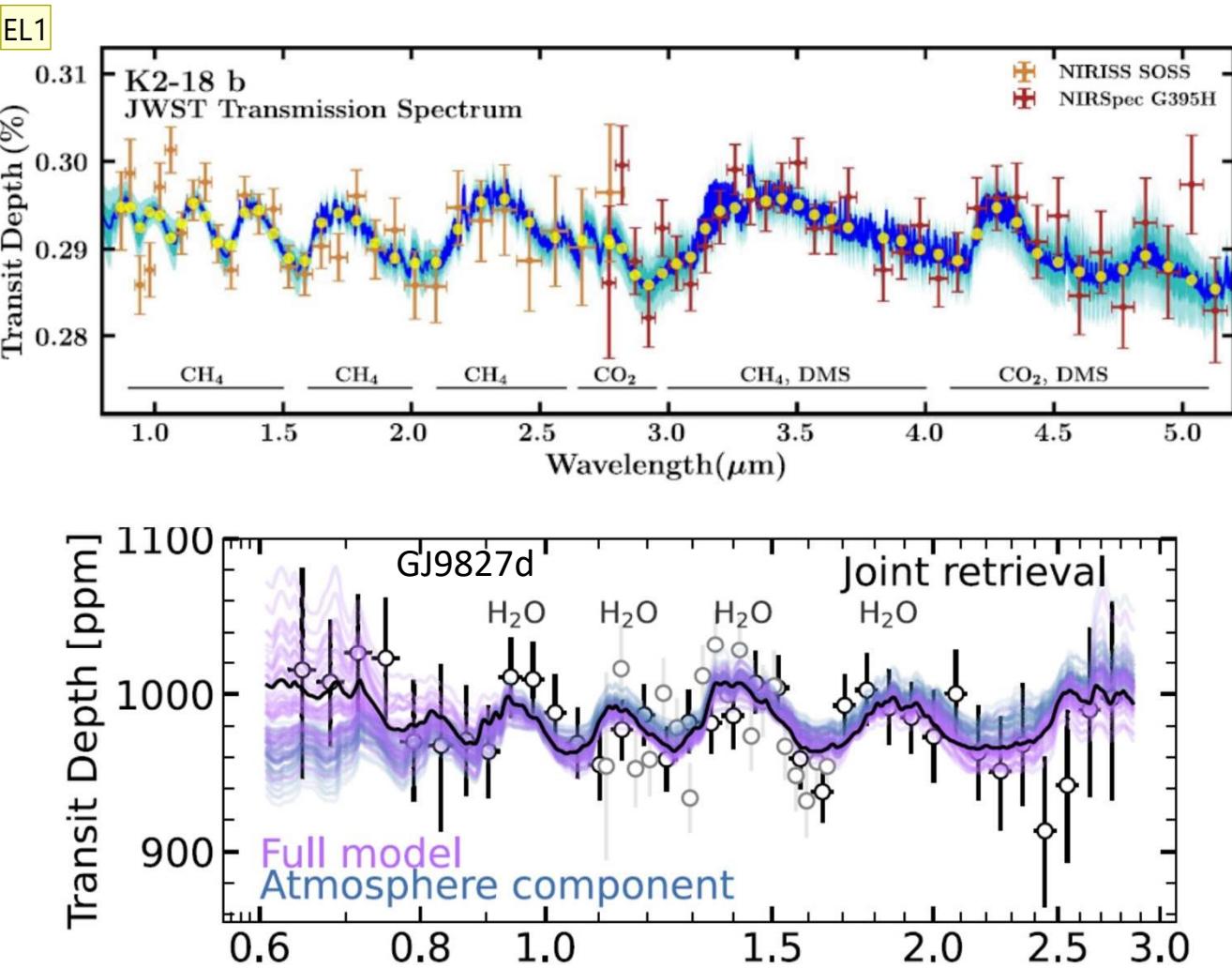


Planets as light as a few Earth masses may have large gas envelopes

Exoplanet atmospheres : lifting the degeneracy (JWST 2024)



K2-18b: CH₄ gas in a H-He dominated atm.
GJ 9827d (one of the smallest exoplanets with detected atm): H₂O, steam atmosphere



Diapositive 25

EL1

Emmanuel Lellouch; 15/01/2025

Vertical structure of a planetary atmosphere (nomenclature)

TERRE

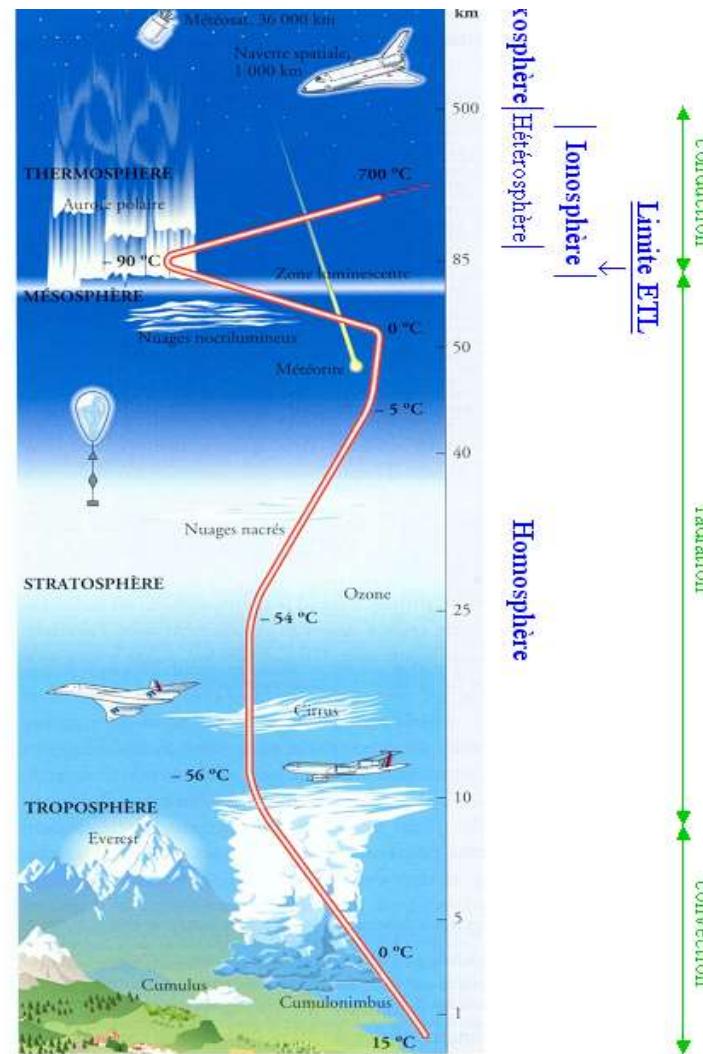
Thermosphère

Mésopause
Mésosphère
Stratopause

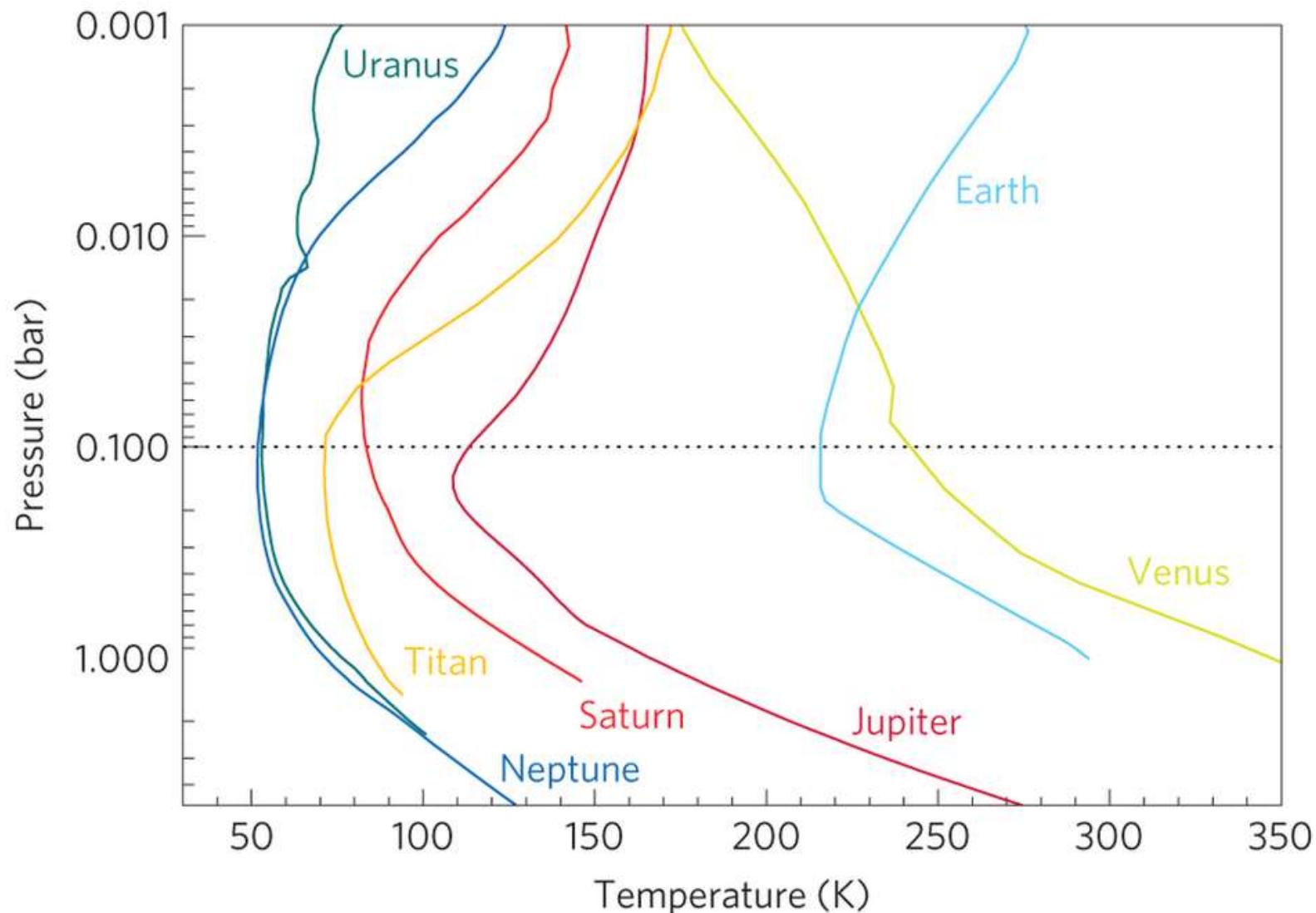
Stratosphère

Tropopause

Troposphère



Tropospheres and stratospheres in the Solar System



Stability of hydrostatic equilibrium and convection

Temperature gradient for adiabatic exchanges

- Dry air:

$$dQ = dU - dW = C_v^* dT + pdV$$

$$\left. \frac{dT}{dz} \right|_{ad,sec} = -\frac{g}{C_p^*}$$

C_p^* = specific heat capacity at constant pressure (J / kg / K)

$$T P^{-\kappa} = \text{cte}$$

$$\kappa = \frac{R}{M C_p^*} = \frac{R}{C_p} = \frac{\gamma - 1}{\gamma}$$

Exercise: calculate C_p^* and κ for N2

- Wet air:

$$dQ = dU - dW = C_v^* dT + L^* dw_s + pdV$$

$$\left. \frac{dT}{dz} \right|_{ad} = \frac{-g/C_p^*}{1 + L^*/C_p^* dw_s/dT}$$

Wet adiabatic lapse rate is smaller than dry lapse rate.

where L^* is latent heat of condensation (J / kg)
 w_s = mass of condensable gas per unit mass of air

Earth :

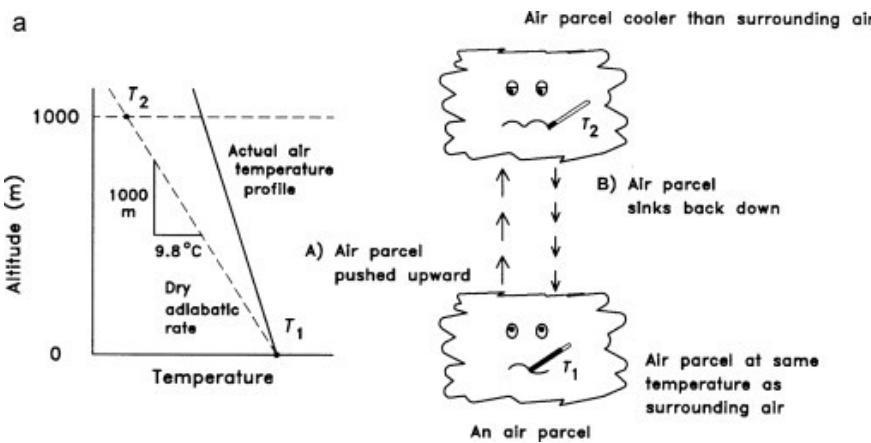
$$(dT / dz)_{dry} = -9.8 \text{ K/km}$$

$$(dT / dz)_{wet} = -5.5 \text{ K/km for air saturated in water}$$

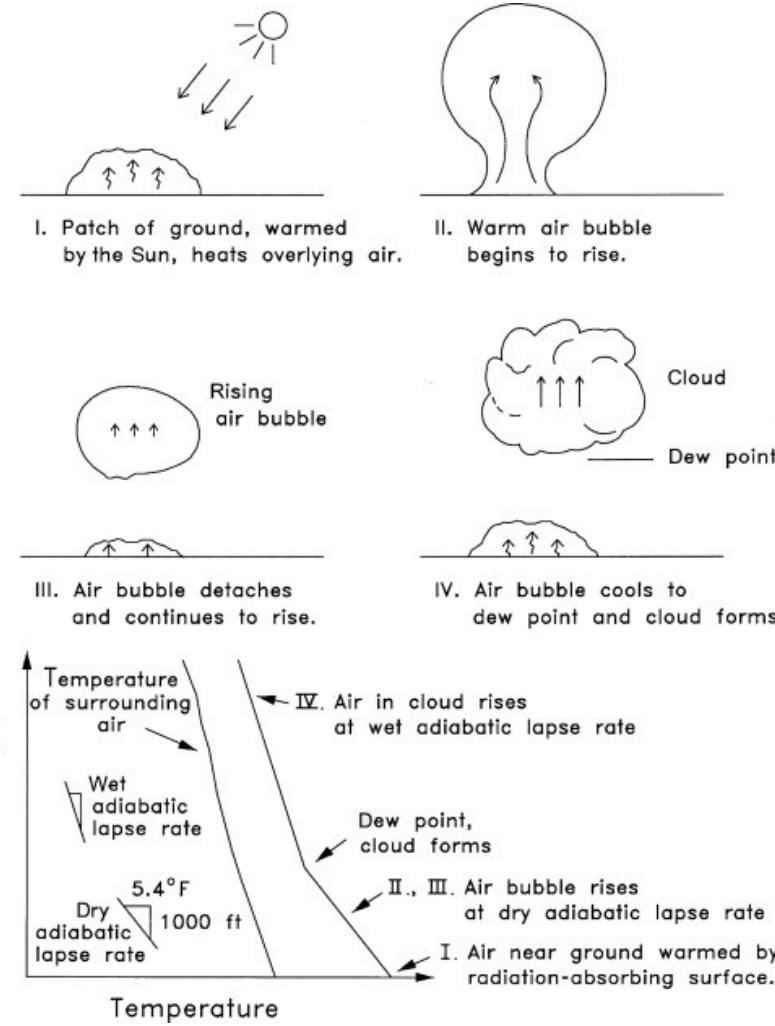
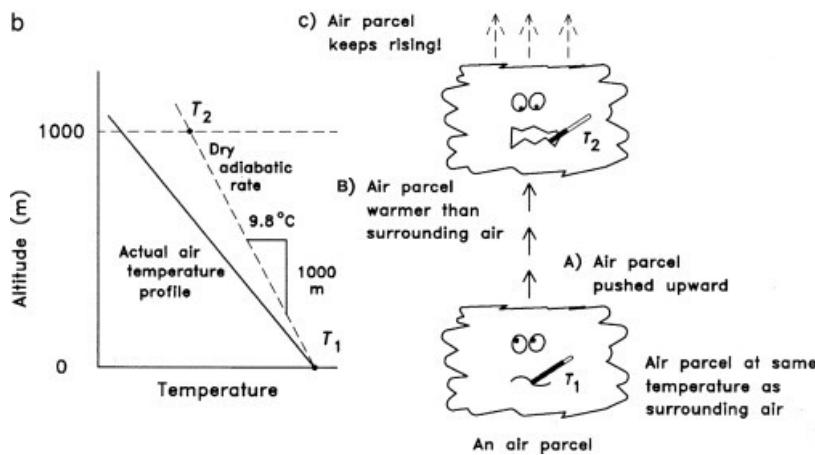
Mean tropospheric conditions ~-6.5 K/km

Stability of hydrostatic equilibrium and convection

a



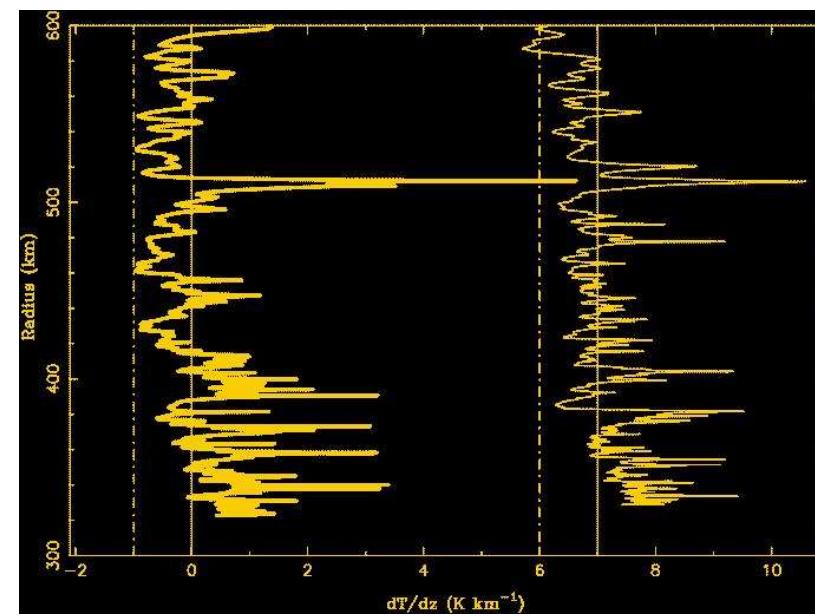
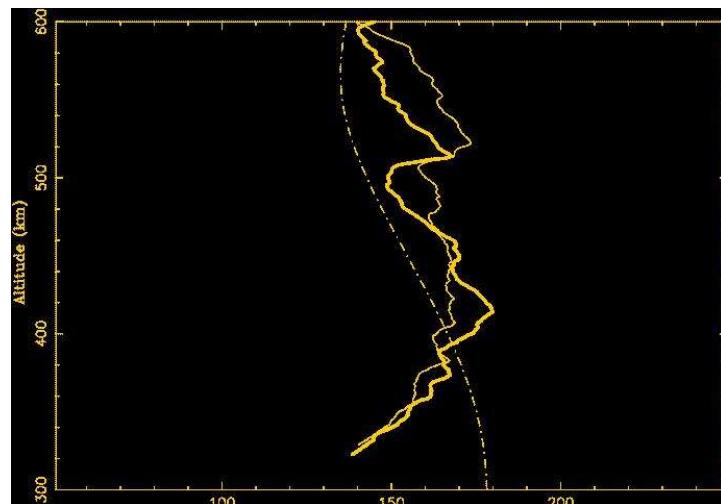
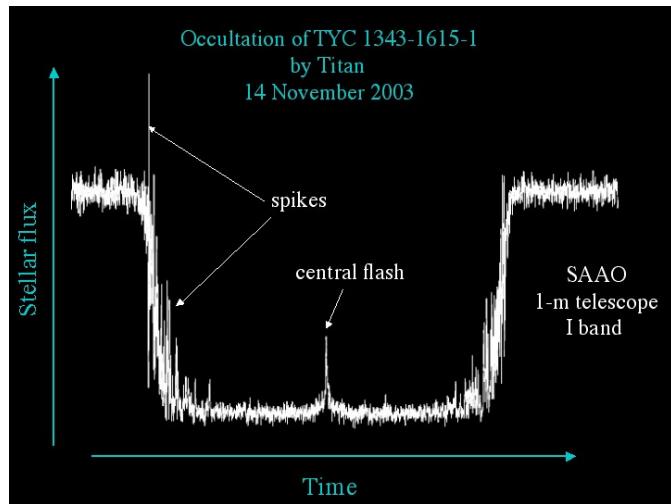
b



Convection

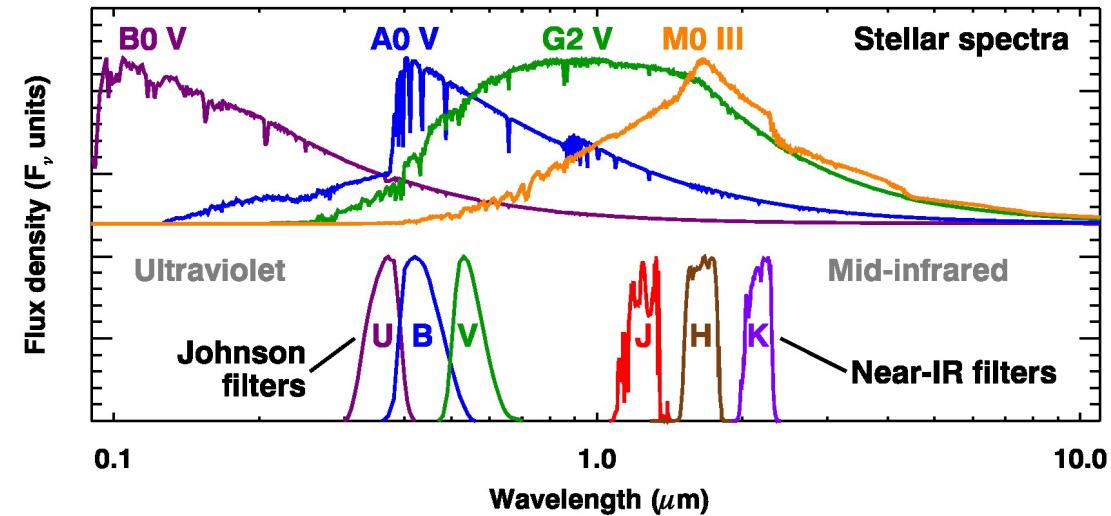
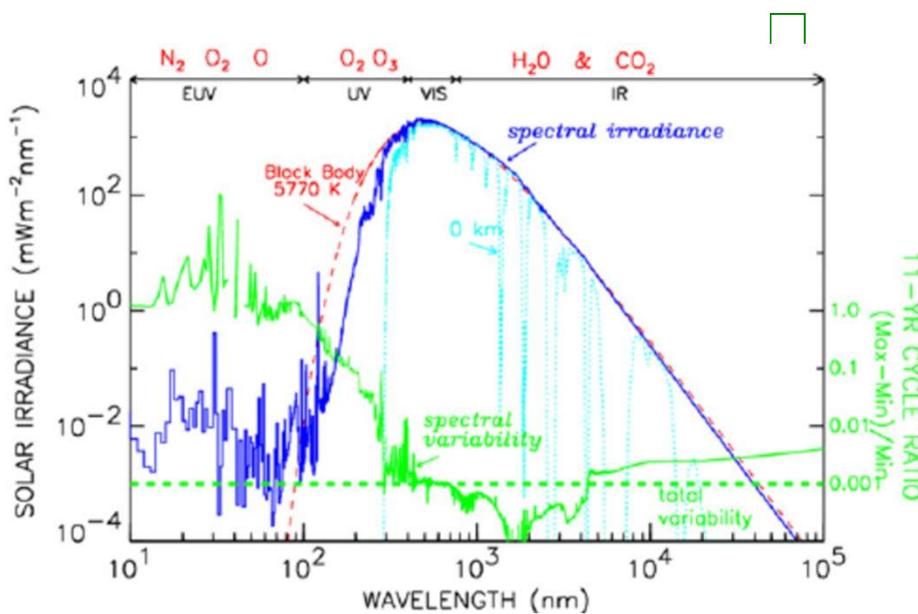
- If $\left| \frac{dT}{dz} \right| < |\Gamma_{ad}|$, an adiabatically ascending airmass, being colder (denser) than the environment, sinks back and oscillate (exercise: what is the period of oscillations?). *Atmosphere stable against convection*
- If $\left| \frac{dT}{dz} \right| > |\Gamma_{ad}|$, (i.e. $dT/dz < \Gamma_{ad}$) an adiabatically ascending airmass, being warmer (less dense) than the environment, keeps rising : *atmosphere unstable against convection*
 - Convection drives upwards energy flux $\Phi_K = -K\rho C_p^* \left(\frac{dT}{dz} - \Gamma_{ad} \right)$ that brings back the gradient to near-adiabatic
 - Except near planetary surfaces, atmospheric profiles are not super-adiabatic

Example: temperature gradients from stellar occultations



Effects of solar / stellar irradiation on planetary atmospheres

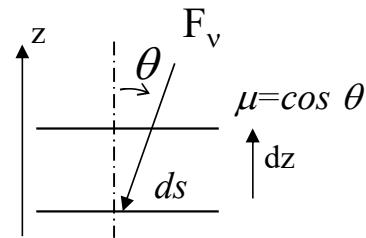
- Solar spectrum
 - Near-blackbody in Vis/IR
 - Strong departure from BB in UV
 - strong emission lines, e.g. Ly α at 121.6 nm
 - strong variability in UV/EUV, up to $\sim 100\%$ below 150 nm



Absorption of solar/stellar radiation in an atmosphere

- Definitions / assumptions
 - Spectral flux density (or spectral irradiance)
 - Atmosphere = 1 molecule, with absorption cross section σ_a at frequency ν (cm^2), and concentration $n(z)$ (cm^{-3})
- Upon traversing at atmospheric layer dz , under incidence angle (θ), F_ν changes by dF_ν

$$F_\nu \text{ (e.g. erg cm}^{-2} \text{ Hz}^{-1} \text{ s}^{-1}\text{)}$$



$$\frac{dF_\nu}{F_\nu} = -k_\nu ds = \sigma_a n(z) \frac{dz}{\mu} = -\frac{d\tau}{\mu}$$

where k_ν = absorption coefficient (cm^{-1}), ds = absorption path in layer = $-dz / \cos(\theta) = dz / \mu$

- At altitude z within atmosphere

$$\tau = \int_z^\infty \sigma_a n(z') dz' \quad \tau \text{ atmospheric opacity at frequency } \nu \text{ integrated from top to } z$$

$$F_\nu = F_{\nu,0} \exp \left(-\frac{1}{\mu} \int_z^\infty \sigma_a n(z') dz' \right) = F_{\nu,0} \exp(-\tau / \mu)$$

Calculation of the absorption rate

Further simplifying hypotheses

- Isothermal atmosphere, absorption cross section σ_a independent of altitude
- Isothermal atmosphere implies $\tau = n_0 \sigma_a H = n \sigma_a H \exp(- (z-z_0)/H)$, hence $d\tau/dz = - \tau / H$

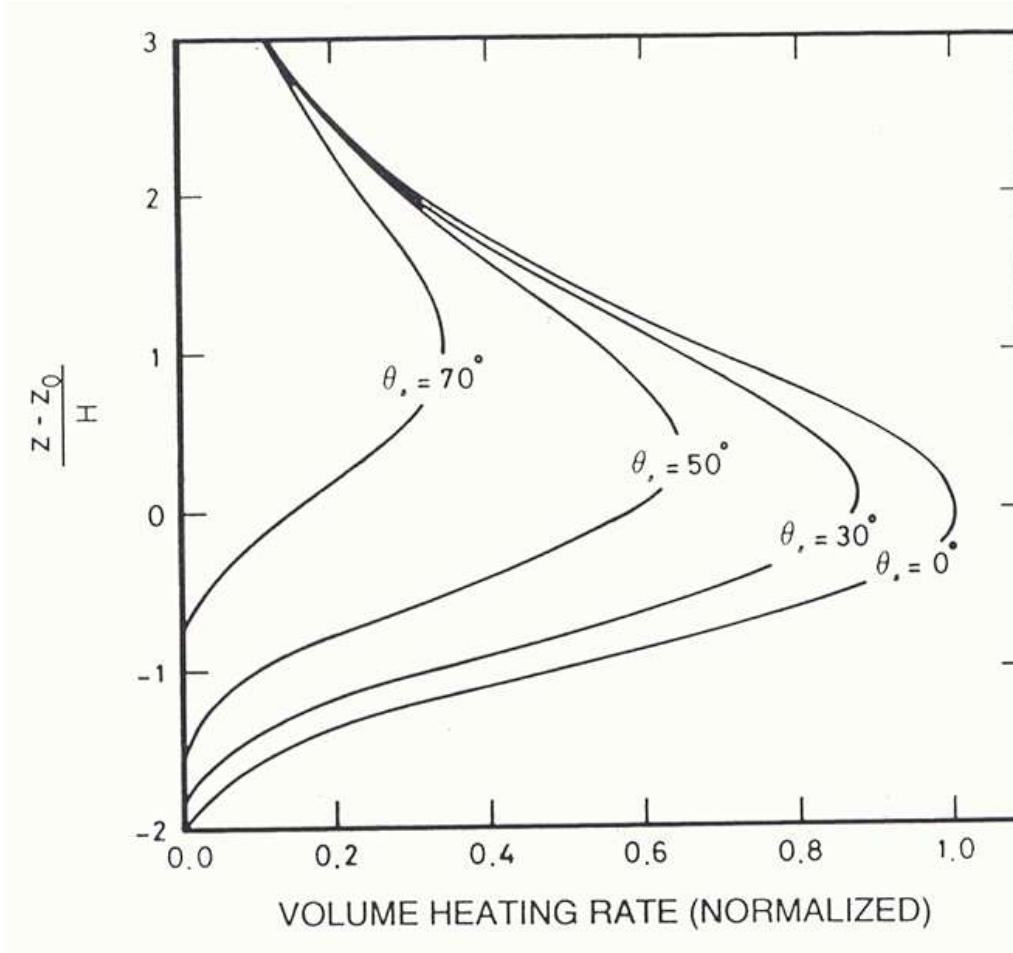
Energy absorption rate

- Defined by $r(z) = - dF_v / ds = \mu dF_v / dz$ (e.g., in $\text{erg cm}^{-3} \text{Hz}^{-1}$)
- Calculation (whiteboard) gives: $r(z) = F_{v,0} \tau / H \exp(- \tau / \mu)$

$$r(z) = F_{v,0} \sigma_a n_0 \exp \left[-\frac{z-z_0}{H} - \frac{1}{\mu} \sigma_a H n_0 \exp(-\frac{z-z_0}{H}) \right]$$

- In reality, should integrate over all frequencies, and account for opacities of all molecules
- **Chapman layer [Exercise]**
 - Absorption maximum at τ / μ (absorption) = 1, dependent on incidence angle, independent on incident flux
 - Value at peak is proportional to $\mu = \cos \theta$ (and of course to insolation)
 - Width of Chapman layer $\sim H$

Chapman layer



Effects of absorbed solar/stellar radiation

- Infrared
 - Absorption by molecules in their vibrational bands, producing heating in « moderately deep » (0.1-100 mbar) part of atmospheres: *stratospheres*
- UV
 - Absorption in upper atmospheres (μ bar range and higher; exception: ozone in Earth stratosphere)
 - Absorption can cause heating, but also photodissociation and photoionization
 - Heating rate < absorption rate (heating efficiency $\varepsilon < 1$)
 - Heating (FUV, EUV) \rightarrow *thermospheres*
 - Photo-dissociation
 - $A + h\nu \rightarrow B + C$ (A = molecule, B, C = radicals) \rightarrow initiation of ***photo-chemistry***
 - Involves only photons more energetic than bond energy of molecule A; role of Ly α
 - Photo-ionization
 - $A + h\nu \rightarrow A^+ + e^- \rightarrow$ production of *ionospheres*

Atmospheric photochemistry

- Photolysis rates

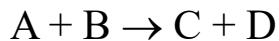
- $A + h\nu \rightarrow B + C$
 - Loss rate for A = Production rate for B and C

$$\frac{dn_B}{dt} = -\frac{dn_A}{dt} = n_A \int_{\nu_0}^{\infty} \sigma_{\nu} \mu F_{\nu} d\nu = J n_A$$

$\hbar\nu_0$ = Bond energy of A = threshold for photolysis
 F_{ν} = photon flux per frequency (ph Hz⁻¹) at altitude z
 σ_{ν} = photodissociation cross section
 J = photodissociation coefficient (sec⁻¹)
(altitude - dependent)

- Chemical reactions

- Two-body



k = reaction rate (cm³ sec⁻¹)

$$\frac{dn_A}{dt} = \frac{dn_B}{dt} = -\frac{dn_C}{dt} = -\frac{dn_D}{dt} = -k n_A n_B$$

- Three-body



$$\frac{dn_A}{dt} = \frac{dn_B}{dt} = -k n_A n_B [M]$$

k = reaction rate (cm⁶ sec⁻¹)

→ Chemical production ($P_i(z)$) and loss ($L_i(z)$) terms (in cm⁻³ s⁻¹) for each species i .

If locally $P_i = L_i$, photochemical equilibrium. True if reactions proceed faster than transport.

Photochemistry with transport

- Eddy and molecular diffusion transport

$$\Phi_i^d = - D_i n_i^e \frac{d(n_i/n_i^e)}{dz} = - D_i n_i \left(\frac{1}{n_i} \frac{dn_i}{dz} + \frac{1}{H_i} + \frac{1}{T} \frac{dT}{dz} \right) \rightarrow \text{Tends towards diffusion equilibrium}$$

$$\Phi_i^K = - K n \frac{d(n_i/n)}{dz} = - K n_i \left(\frac{1}{n_i} \frac{dn_i}{dz} + \frac{1}{H} + \frac{1}{T} \frac{dT}{dz} \right) \rightarrow \text{Tends toward uniform mixing ratio } (q_i = n_i / n = \text{cte})$$

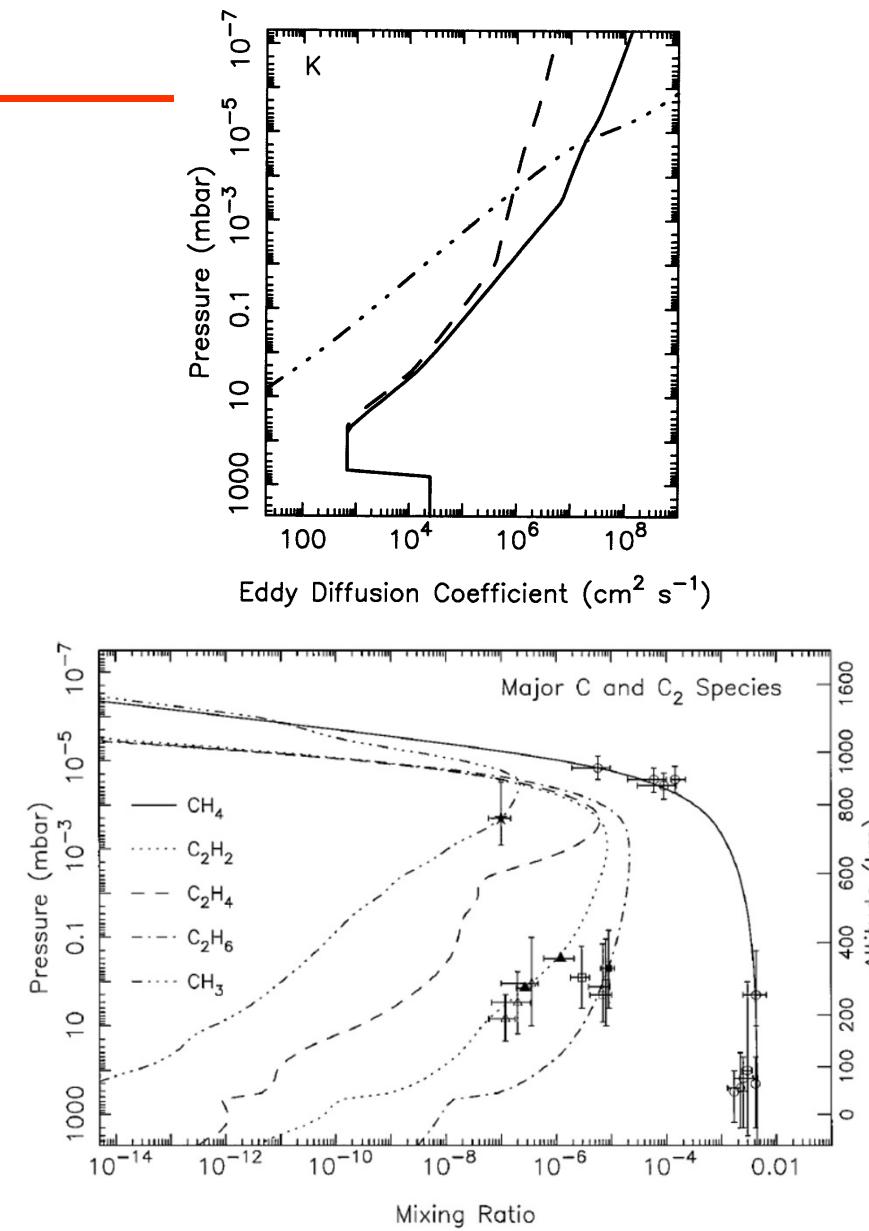
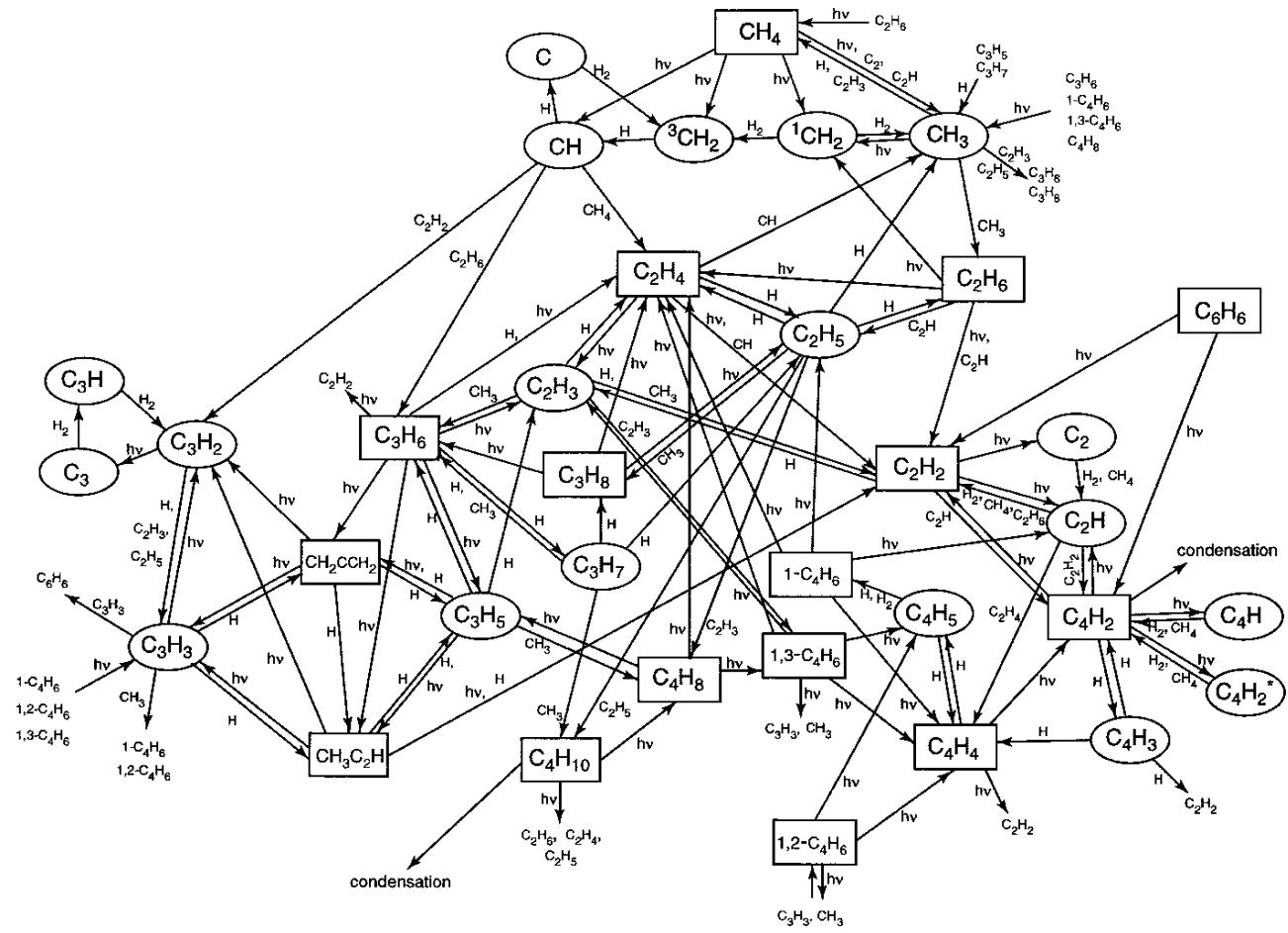
- Transport equation with chemical sources (« continuity equation »)
 - Gradient of (eddy+molecular) flux = « Net » chemical production

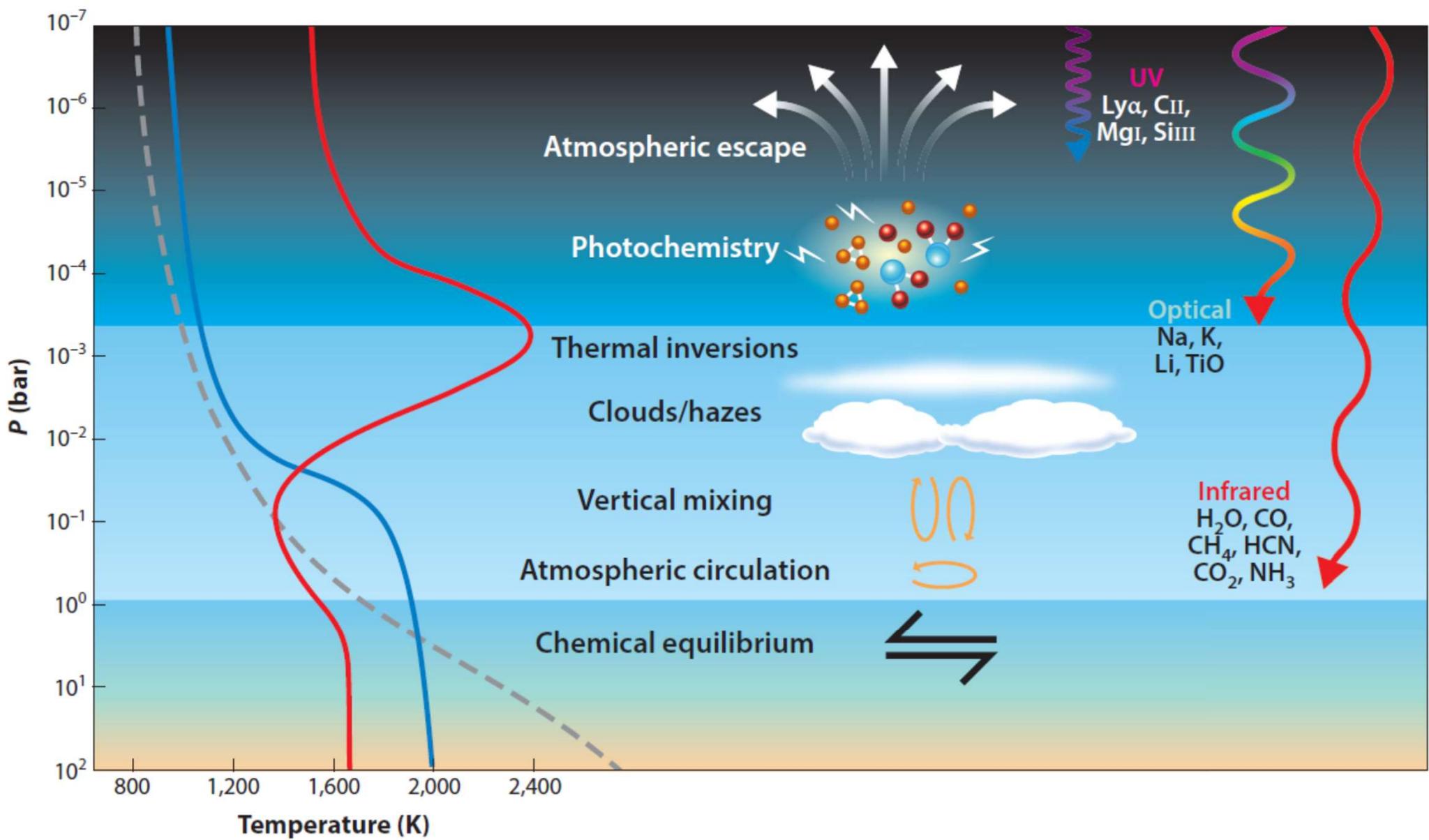
$$\Phi_i = \Phi_i^d + \Phi_i^K$$

$$\frac{d\Phi_i}{dz} = P_i - L_i$$

P_i = chemical production rate; L_i = chemical loss rate

Example: Saturn's methane photochemistry





Chemistry in deep atmospheres

- Thermochemical equilibrium

- In deep / hot atmospheres, chemical equilibria occur according to the law of mass action



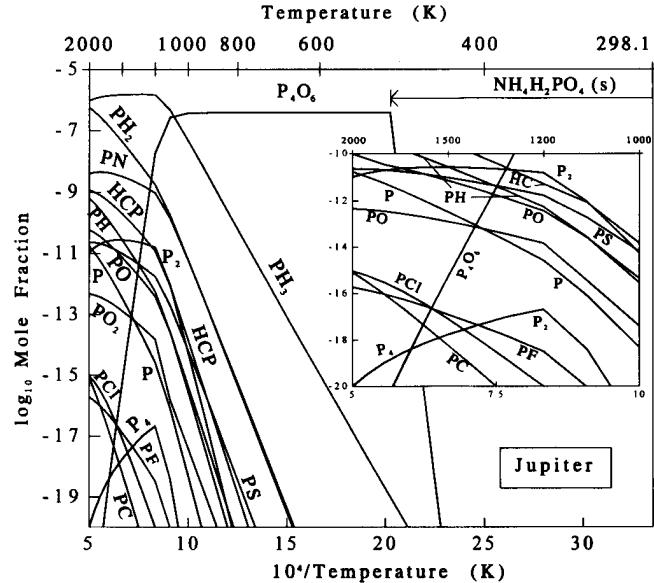
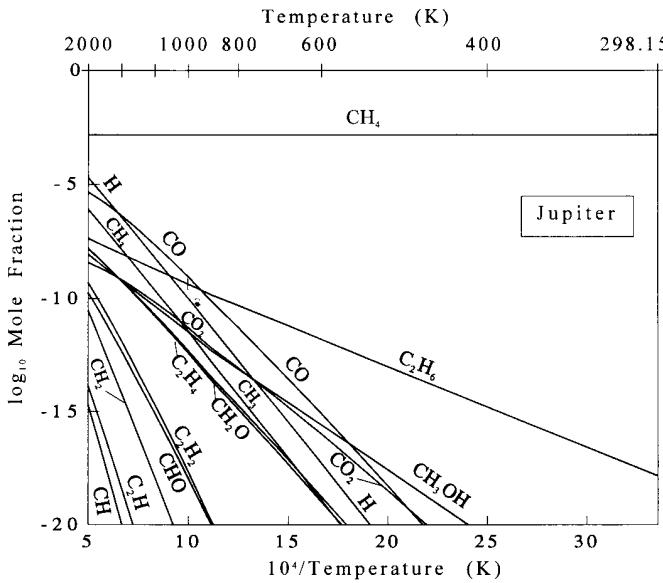
$$K_p(T) = \frac{p_C^\gamma p_D^\delta}{p_A^\alpha p_B^\beta} = e^{-\frac{\Delta G_0(T)}{RT}}$$

- However, reactions are not instantaneous. Typically, reaction rates vary as $\exp(-E_a/kT)$, where E_a is activation energy (Arrhenius law)
 - Equilibria are reached whenever reactions are faster than other processes, especially transport. Typically at $T \geq 500-1500$ K, but very species-dependent.

- Application to Giant Planets

- Thermochemical **equilibrium** predicts that, throughout atmosphere: C in mostly in form of CH₄ ; N as NH₃ ; O as H₂O ; S as H₂S
 - Indeed, in the Solar System Giant Planets, this is observed
 - Jupiter CH₄/H₂ ~ 2 10⁻³ ; H₂O ~ 10⁻⁵ ; NH₃ ~ 2 10⁻⁴ while CO ~ 10⁻⁹ and N₂ is undetected.
 - However, some predictions are not matched...

Chemistry in deep atmospheres



← Thermochemical equilibrium calculations

Yet, species like CO and PH₃ are observed in Jupiter's upper troposphere (~300 K) with non-negligible abundances (10^{-9} and 10^{-6}) ...

- Interpretation : chemical time constants (t_{chem}) increase dramatically as T decreases since ($k \sim e^{-E_a/kT}$). Chemical reactions are ‘frozen’
- Deep atmosphere is mixed by convection with time constant $t_{\text{mix}} \sim H^2/K$
- The observed chemical composition in the upper troposphere is not representative of equilibrium at this level, but of the level where $t_{\text{chem}} \sim t_{\text{mix}}$ (“quenching level”). Species are transported vertically faster than they can be destroyed chemically.
- Evidence also for such *disequilibrium chemistry* in exoplanets

Atmospheric escape: (1) thermal

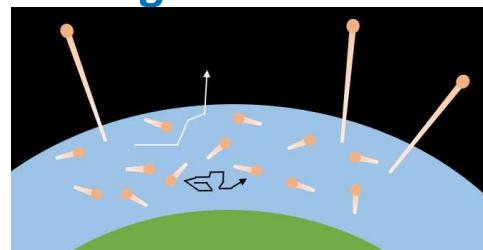
- **Jeans escape:** Jeans parameter $\lambda_{esc} = \text{ratio of gravitational to thermal energy}$

$$\lambda_{esc} = \frac{GMm}{kT(R+z)} = \left(\frac{v_e}{v_{th}}\right)^2$$

$$\Phi_{Jeans} = \frac{n_{exo} v_{th}}{2\sqrt{\pi}} (1 + \lambda_{esc}) \exp(-\lambda_{esc})$$

$$\tau_{Jeans} = \frac{N_{exo}}{\Phi_{Jeans}}$$

where Φ is in $\text{mol cm}^{-2} \text{s}^{-1}$ and N_{exo} in mol cm^{-2}

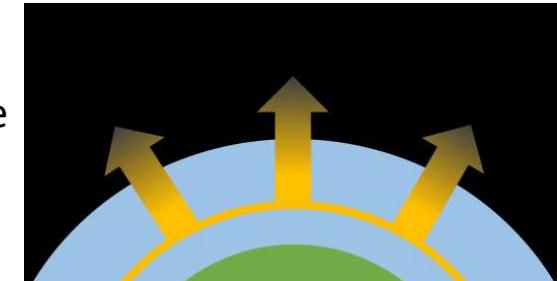


Examples: Earth T ~1000 K
 Atomic H: $\lambda_{esc} = 7$, $\tau_{Jeans} \sim 1$ day
 Molecular H₂: $\lambda_{esc} = 14$, $\tau_{Jeans} \sim 1$ year

Jupiter : H: $\lambda_{esc} = 185$, $\tau_{Jeans} \sim \text{infinite!}$

- **Hydrodynamical escape (= « blow off »)**

- If $\lambda_{esc} < \sim 1.5$, massive escape of light gases entrains other (heavier) gases in global escape
- Energy-limited escape flux
 - All of the absorbed heating energy E_{EUV} ($\text{erg cm}^{-2} \text{s}^{-1}$) is used in escape against gravitational field
 - Likely a large overestimate (x10-100) of actual escape rate
 - Even so, predicts inconsequential escape rates for hot Jupiter ($\sim 10^{12} \text{ g / s}$, i.e a lifetime of ? (exo) years)
- Has been important in early atmospheres of telluric planets (highly irradiated)
- Observed to be present on close-by exoplanets (evidence for extended cloud of H)



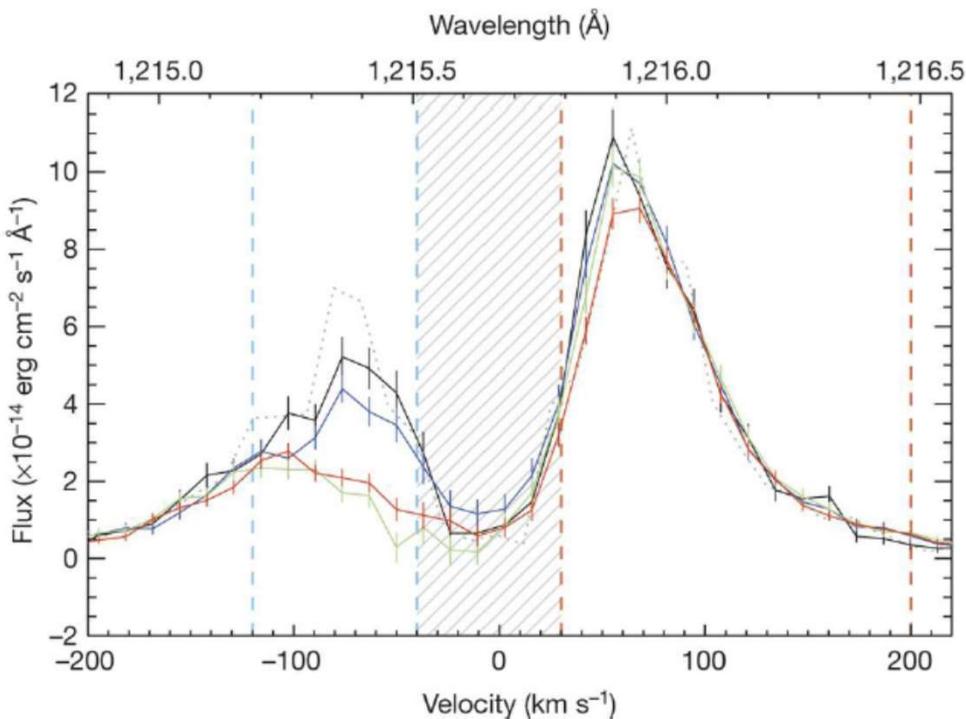
$$F_{esc} (\text{mol/s}) = \frac{E_{EUV}}{U_P} 4\pi R_{EUV}^2$$

$$U_P = \frac{GMm}{R_P} \quad (\text{erg})$$

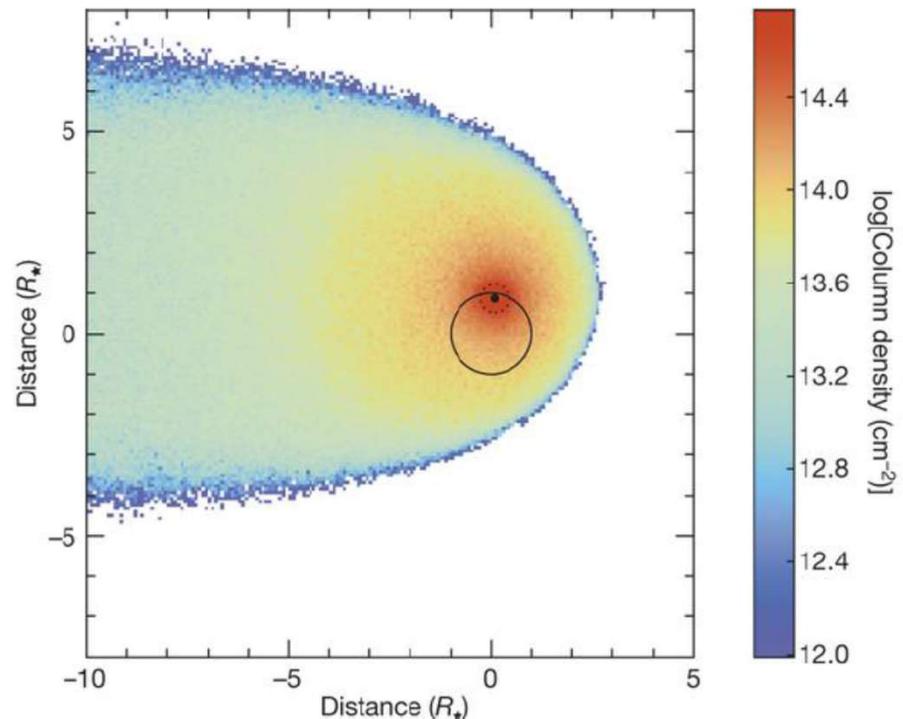
$$F_{esc} (\text{g/s}) = \frac{E_{EUV}}{GM} 4\pi R_{EUV}^2 R_P$$

Direct evidence for current atmospheric escape in exoplanets

GJ436b (hot Neptune, $R = 4 R_E$, $a = 0.028$ au, $T_{eq} \sim 700$ K)

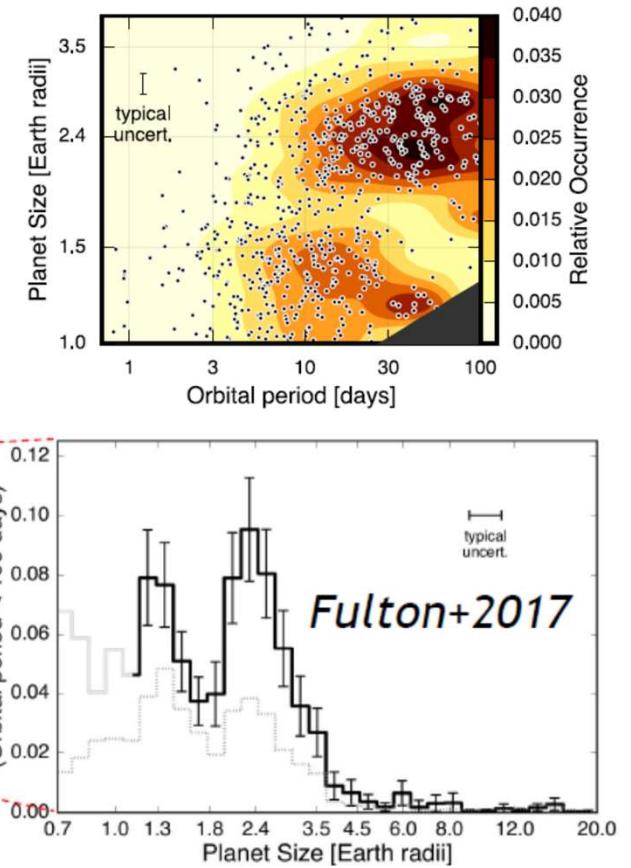
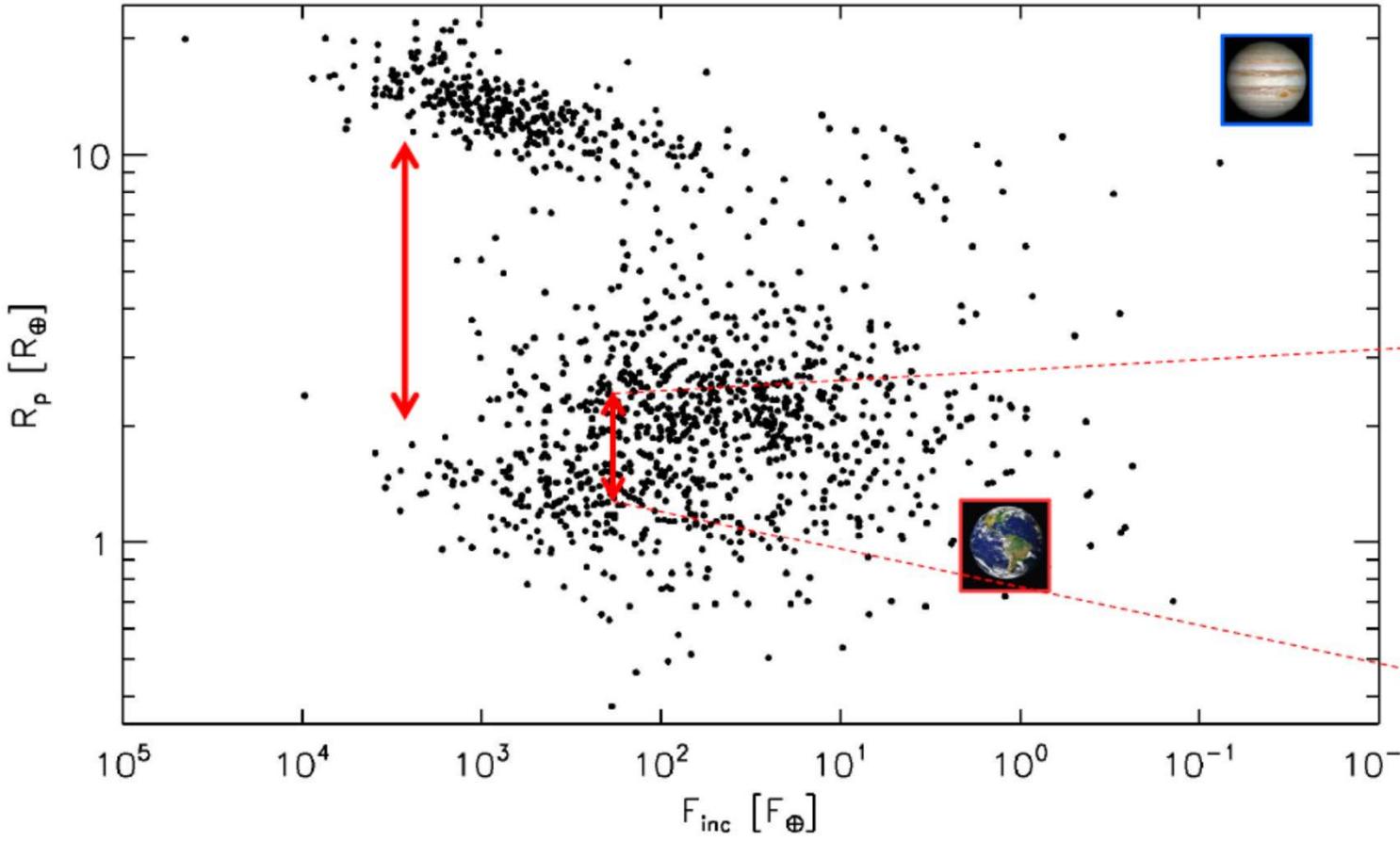


Transit spectroscopy
 Absorption depth $\sim 50\%$ \Leftrightarrow the extended H atmosphere has
 an effective (obscuring)area $\sim 50\%$ of the star !



Ehrenreich+2015
Bourrier+2016

Evidence for atmospheric escape in exoplanets



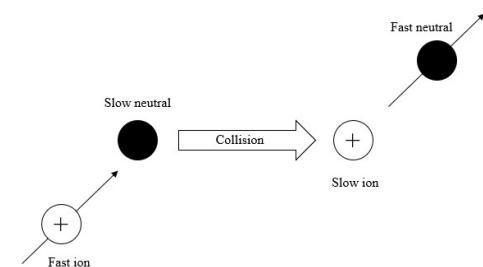
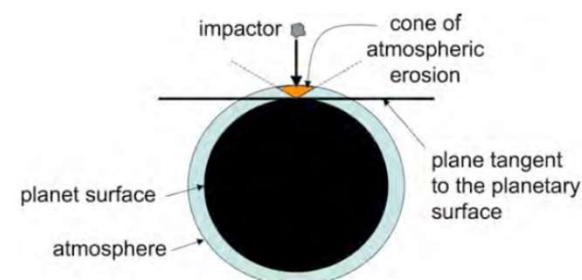
Atmospheric escape: (2) non-thermal

- Non-thermal escape = processes accelerating neutral molecules beyond thermal velocity

- Charge exchange between neutrals and ions : $H^+ \text{ (fast)} + H \text{ (slow)} \rightarrow H \text{ (fast)} + H^+ \text{ (slow)}$
- Photodissociation with extra energy (e.g. $O_2 + h\nu \rightarrow O^* + O^*$)
- Atmospheric sputtering: impact of stellar wind or particles from planetary magnetospheres

- Impact erosion (asteroids, comets)

PLANET	KEY GASES LOST	DOMINANT MECHANISMS
Earth	Hydrogen	Charge exchange, Jeans, polar wind
	Helium	Polar wind, charge exchange
Early Earth	Hydrogen and moderately light gases, including neon	Hydrodynamic escape and drag
Venus	Hydrogen, helium	Charge exchange, sputtering
Early Venus	Hydrogen and moderately light gases, including oxygen	Hydrodynamic escape and drag
Mars	Hydrogen	Jeans
	Carbon, oxygen, nitrogen, argon	Sputtering, photochemical
Early Mars	All gases	Impact erosion
	Hydrogen and many heavier gases, including carbon dioxide	Hydrodynamic escape and drag



H escape from Earth



Methods of investigation of planetary atmospheres (Solar System)

Imaging/spectroscopy in solar reflected range (visible, near-IR)

- Imaging: cloud and haze characterization
 - morphologies, evolution, wind measurement
- Spectroscopy of « clear » atmospheres
 - Measure of gas column densities

$$F_\nu = F_{\nu,0} \exp \left[-\tau_\nu \left(\frac{1}{\mu_1} + \frac{1}{\mu_2} \right) \right] A_{surf}$$

- No vertical information

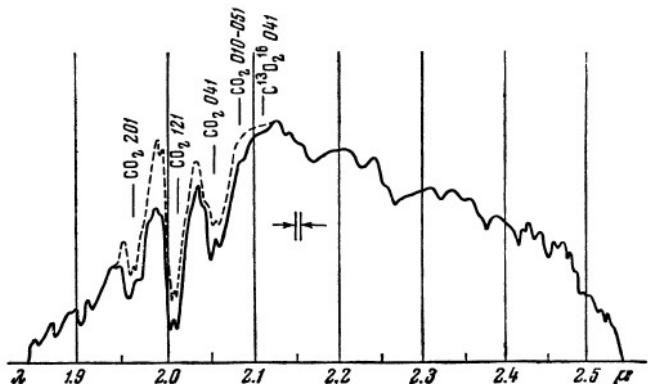
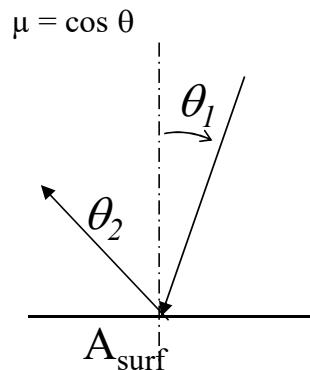
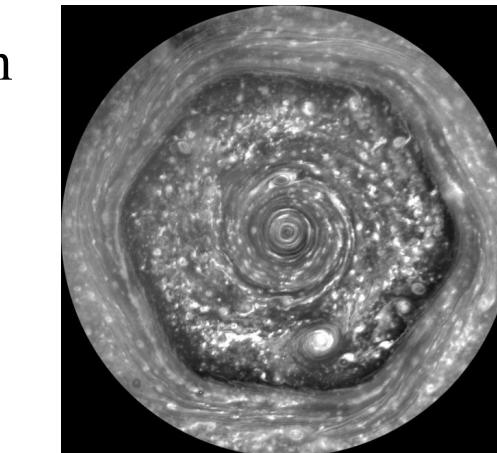


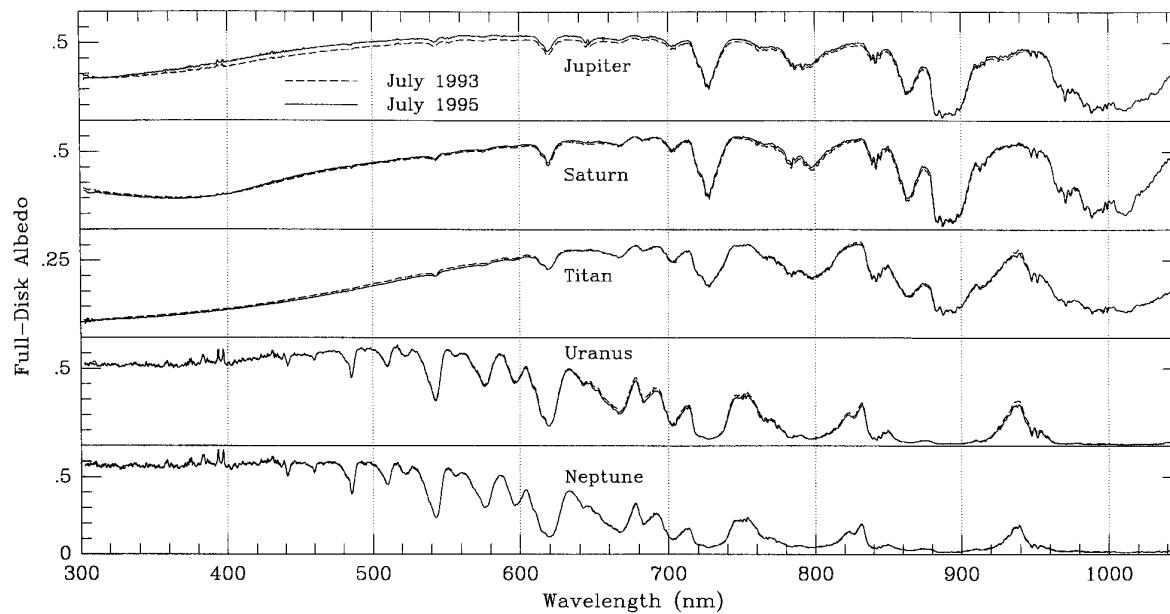
Fig. 2. Martian spectrum in 1.85-2.55 μ region. Grating spectrometer, PbS with CO₂ cooling, mean of three recordings of March 5, 1963, Z = 34-44°, slit 1.5 mm ($\Delta\lambda = 60$ Å), rate 7.3 Å/sec, $\tau = 4.4$ sec. Solar spectrum shown by dashed line.



First measurement of Mars surface pressure (1963)

Spectroscopy in solar reflected component: hazy atmospheres

- Competition between
 - Backscattering on hazes / clouds (scattering coefficient σ_ν)
 - Absorption between top of atmosphere and the scattering layer (absorption coefficient k_ν)
- Penetration level (« probed level ») depends on relative values of k_ν and σ_ν , specifically on $\omega_0 = \sigma_\nu/(k_\nu + \sigma_\nu)$ = single scattering albedo
 - Strong scattering / weak absorption $\rightarrow \omega_0 \sim 1 \rightarrow$ deep sounding
 - Weak scattering / strong absorption $\rightarrow \omega_0 \ll 1 \rightarrow$ shallow sounding
- Often difficult to determine accurately; requires observing a gas with known vertical profile
- In practice, solar light usually does not penetrate deeper than ~ 1 bar
- Information on cloud/haze properties (vertical distribution, particle size)
- Information on column-integrated gases down to scattering level
 - No vertical information
 - Molecular bands are always in absorption in reflected component

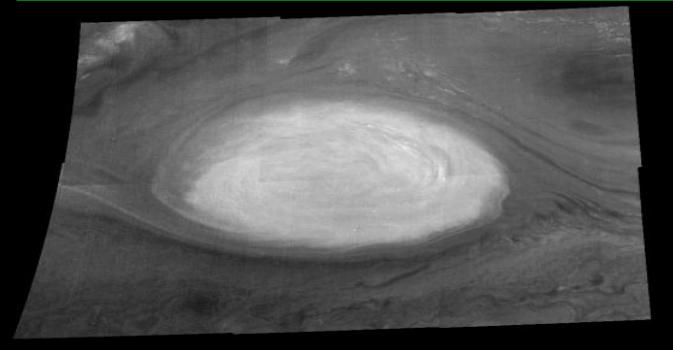
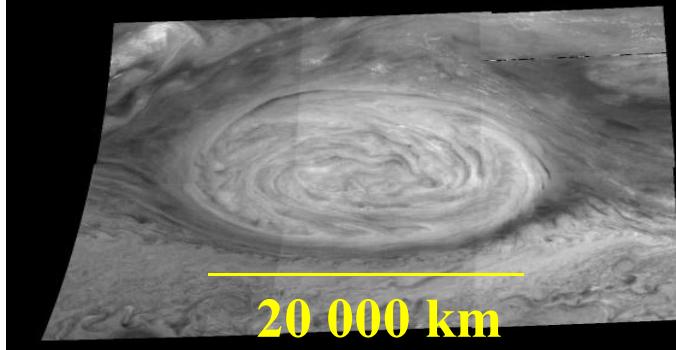


$$\lambda = 732 \text{ nm}$$

Weak CH₄ absorption → « deep »

$$\lambda = 886 \text{ nm}$$

Strong CH₄ absorption → shallow



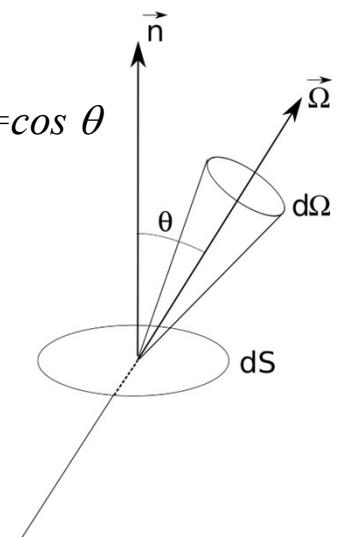
At both λ (especially 886 nm) Jupiter Great Red Spot is brighter than its surroundings.... What do you infer ?

Spectroscopy in thermal range

- Thermal component = emission of atmospheric layers
 - Each layer, as seen from an outside observer
 - Emits its own, temperature-dependent radiation, usually as a blackbody
 - Partially absorbs radiation from layers below it
 - Definitions
 - **Specific intensity (or spectral radiance)** I_ν at frequency ν (unit = $\text{W m}^{-2} \text{sr}^{-1} \text{Hz}^{-1}$) such as : Power emitted by elemental surface dS in solid angle $d\Omega$, under normal angle θ in frequency interval $d\nu$:
- $dP = I_\nu \cos \theta dS d\Omega d\nu$ $d\Omega = 2 \pi \sin \theta d\theta$; Ω (sphere) = 4π
- I_ν is homogeneous to a Planck function
 - Mean spectral intensity J_ν $J_\nu = \frac{\int I_\nu d\Omega}{\int d\Omega}$
 - Spectral flux $F_\nu = \int I_\nu \cos \theta d\Omega = 2 \pi \int_0^\pi I_\nu \cos \theta \sin \theta d\theta = 2 \pi \int_{-1}^{+1} I_\nu \mu d\mu$
 - Total emitted flux in half-space per unit area for a *blackbody* (Stefan-Boltzmann law)

$$P = \int_0^\infty I_\nu d\nu \int \cos \theta d\Omega = 2 \pi \int_0^\infty I_\nu d\nu \int_0^{\pi/2} \cos \theta \sin \theta d\theta$$

$$P = \pi \int_0^\infty B_\nu(\nu, T) d\nu = \sigma T^4$$



Radiative transfer in thermal range

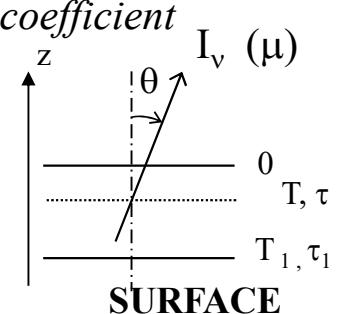
- General radiative transfer equation $dI_\nu = - I_\nu \alpha_\nu ds + j_\nu ds$

α_ν = extinction coefficient ; $\alpha_\nu = k_\nu + \sigma_\nu$, where k_ν = absorption coefficient and σ_ν = scattering coefficient
 j_ν = emission coefficient; $ds = \cos \theta dz = \mu dz$

Defining: source function $S_\nu = j_\nu / \alpha_\nu$, i.e. $j_\nu = \alpha_\nu S_\nu$

$$d\tau_\nu = -(k_\nu + \sigma_\nu) dz = -\alpha_\nu dz, \quad \text{defining} \quad \tau_\nu = \int_z^\infty (k_\nu + \sigma_\nu) dz$$

$$\mu \frac{dI_\nu}{d\tau_\nu} = I_\nu - S_\nu$$



- If no scattering, $\alpha_\nu = k_\nu$ and if the medium is in local thermodynamical equilibrium (LTE),

then $S_\nu = B_\nu = \frac{2h\nu^3/c^2}{\exp(h\nu/kT)-1}$ (Planck function)

- Solution $I_\nu(\mu, 0) = \frac{1}{\mu} \int_0^{\tau_1} B_\nu(T) e^{-\tau/\mu} d\tau + B_\nu(T_1) e^{-\tau_1/\mu}$ where T_1 = surface temperature, τ_1 = opacity at surface

- Outgoing radiation bears information (vertically-resolved) on

- Atmospheric composition (through τ)
- Atmospheric temperature structure (mostly through T in Planck function)

Thermal spectroscopy

- Without surface term (« semi-infinite atmosphere »)
- Define brightness temperature T_b as: $I_\nu(\mu,0) = B_\nu(T_b)$
- Radiation originates mostly from level where $\tau_\nu = \mu = \cos \theta$ ($\tau_\nu = 1$ for vertical viewing) and the measured radiance is \sim equal to the Planck function at the temperature at this level

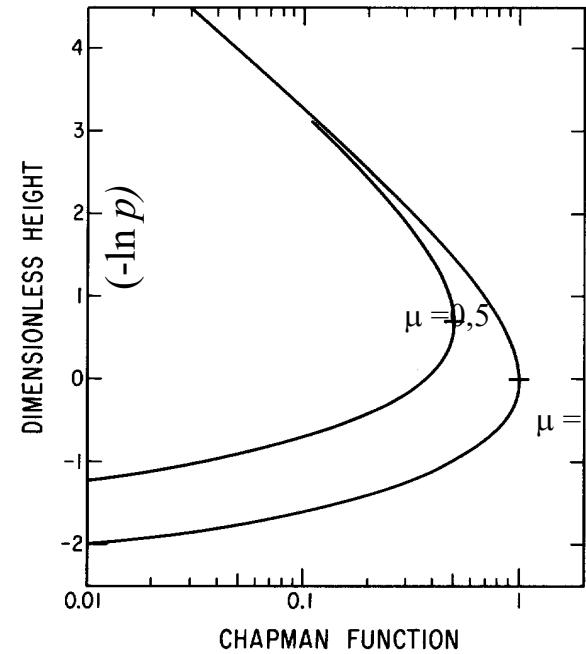
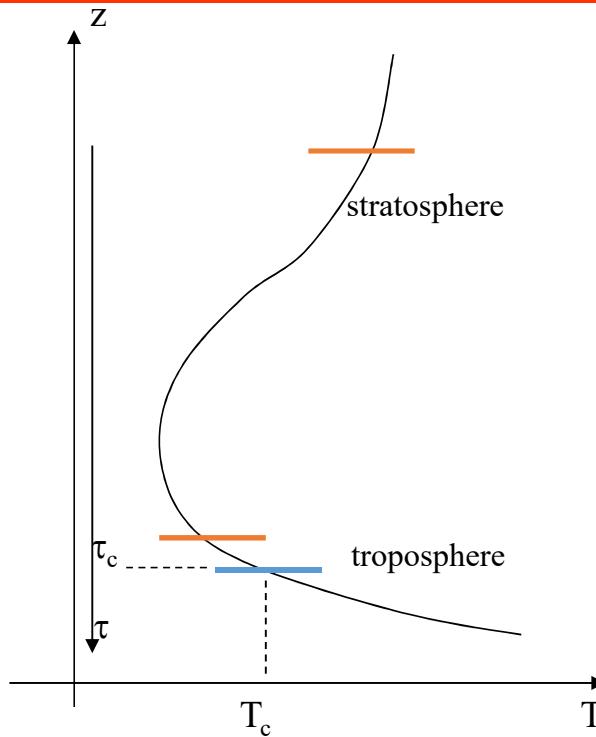
$$T_b(\nu) \sim T[(\tau_\nu = 1)]$$

$$I_\nu(\mu,0) = \frac{1}{\mu} \int_0^\infty B_\nu(T) e^{-\tau/\mu} d\tau$$

$$\tau_\nu = \int_z^\infty k_\nu dz = \int_z^\infty \sigma_a n(z') dz'$$

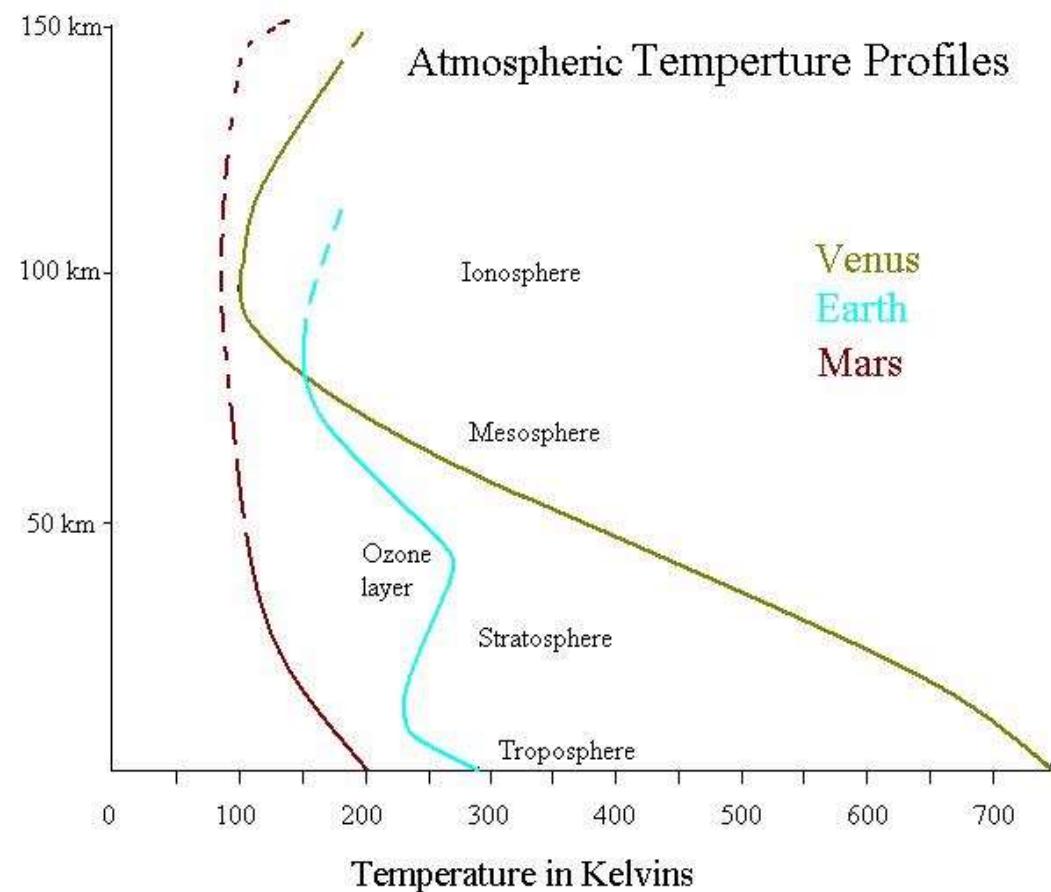
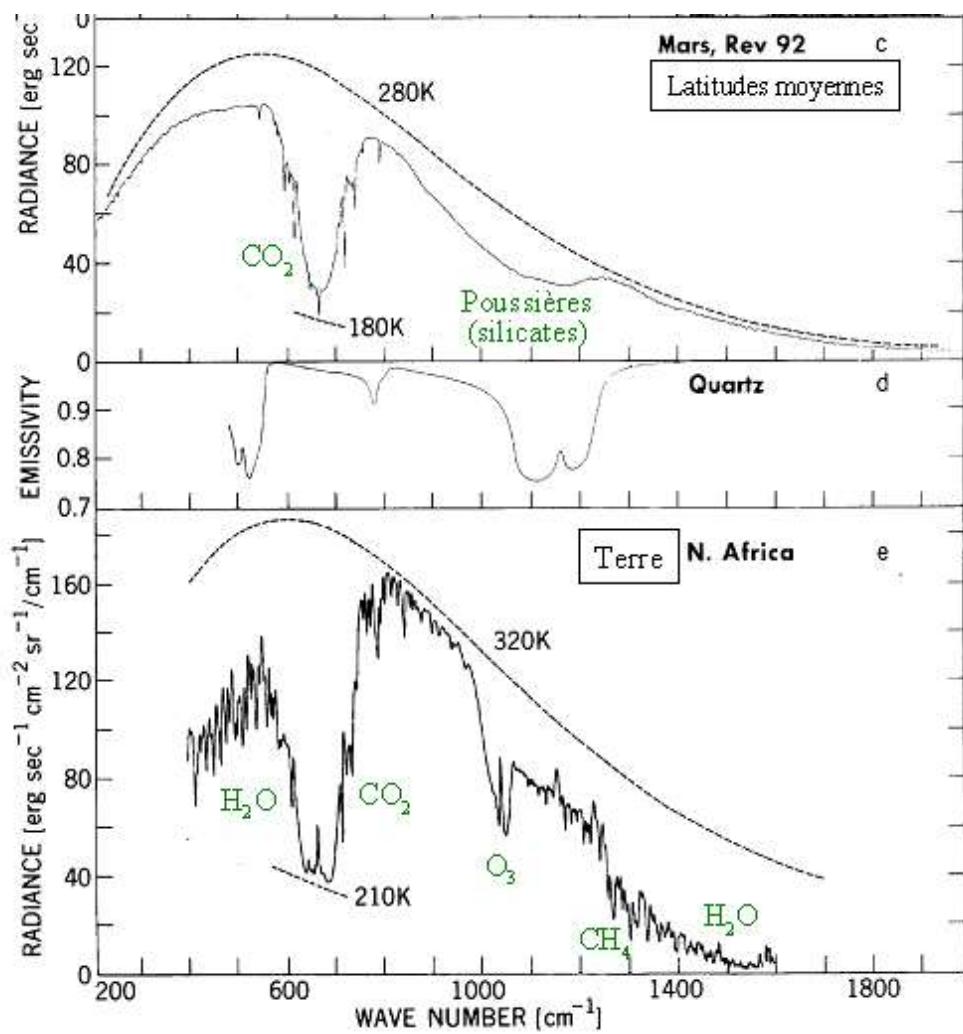
- Molecular opacities vary with $\nu \rightarrow$ the altitude of $\tau_\nu = 1$ level varies with $\nu \rightarrow$ vertical sounding
 - If atmospheric $T(z)$ profile is known, one can determine the altitude z_0 where $T_b(\nu) = T(z_0)$. Then, $\tau_\nu = 1$ occurs at z_0 , and one can determine the column abundance of the absorber above this level. Different $\nu \rightarrow$ different $z_0 \rightarrow$ vertical profile of abundance : **compositional sounding**
 - If abundance profiles are known, one can calculate τ_ν and thus the $z_0(\nu)$ level at which $\tau_\nu = 1$. The measured $T_b(\nu)$ can then be identified to $T(z_0)$: **thermal sounding**
 - In thermal component, lines/bands can appear in absorption or in emission, depending on whether the probed level in the line/band is cooler or warmer than in the adjacent frequencies
- In reality the « $\tau = 1$ level » is a Chapman layer with finite width (~ 2 scale heights around $\tau = 1$)

Spectral lines in thermal radiation

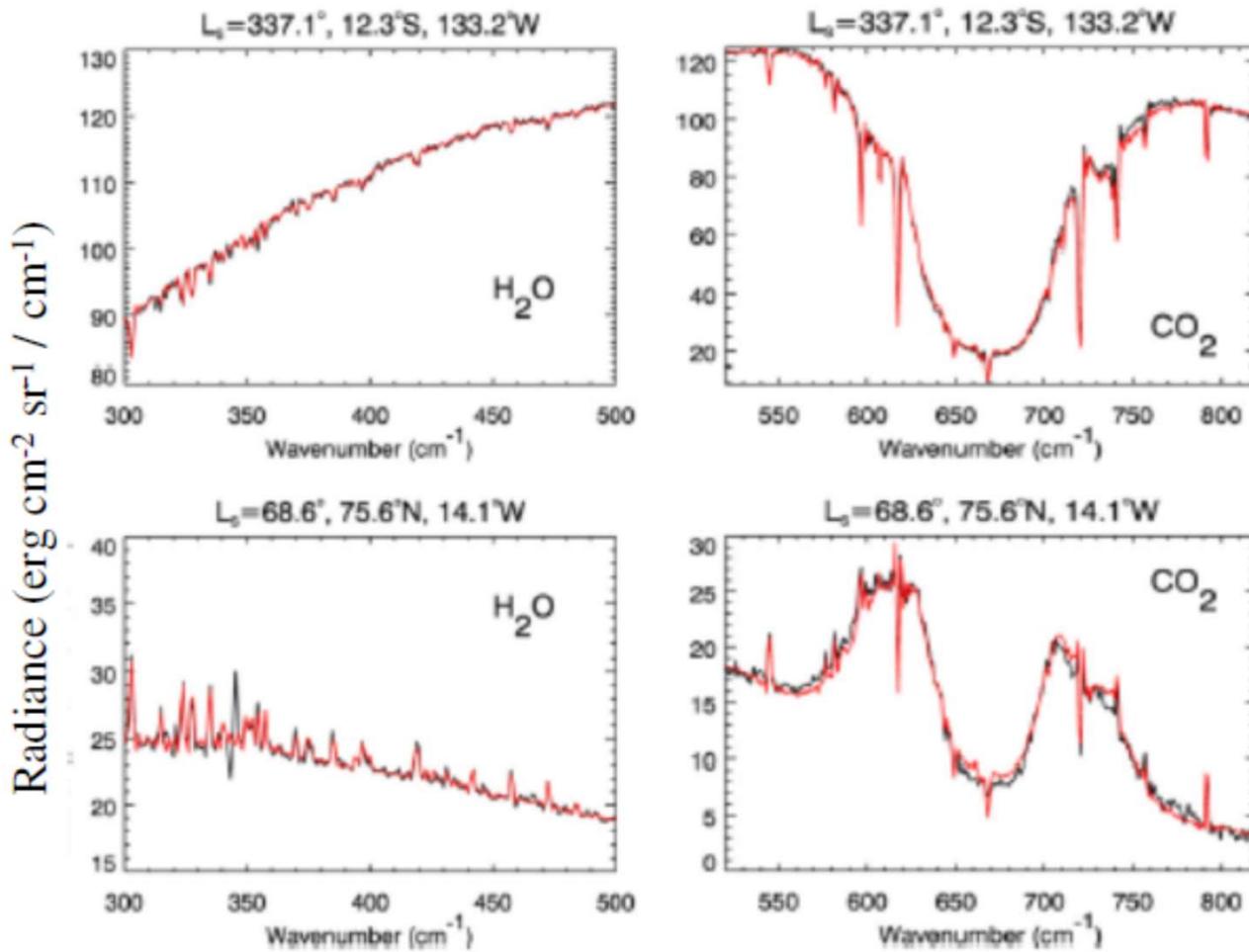


- Let ν_c be a « continuum » frequency, i.e. a frequency outside a molecular absorption
- At $\nu \neq \nu_c$ (but close), $\tau_\nu(z) \nearrow$, hence the $\tau_\nu = 1$ level moves upwards (ALWAYS)
 - If $T < T_c$, the radiance decreases \Rightarrow spectral feature in *absorption*
 - If $T > T_c$ (large line opacity, with $\tau_\nu = 1$ reached in stratosphere, the radiance decreases \Rightarrow spectral feature in *emission*)

Example: thermal spectra of Mars and Earth

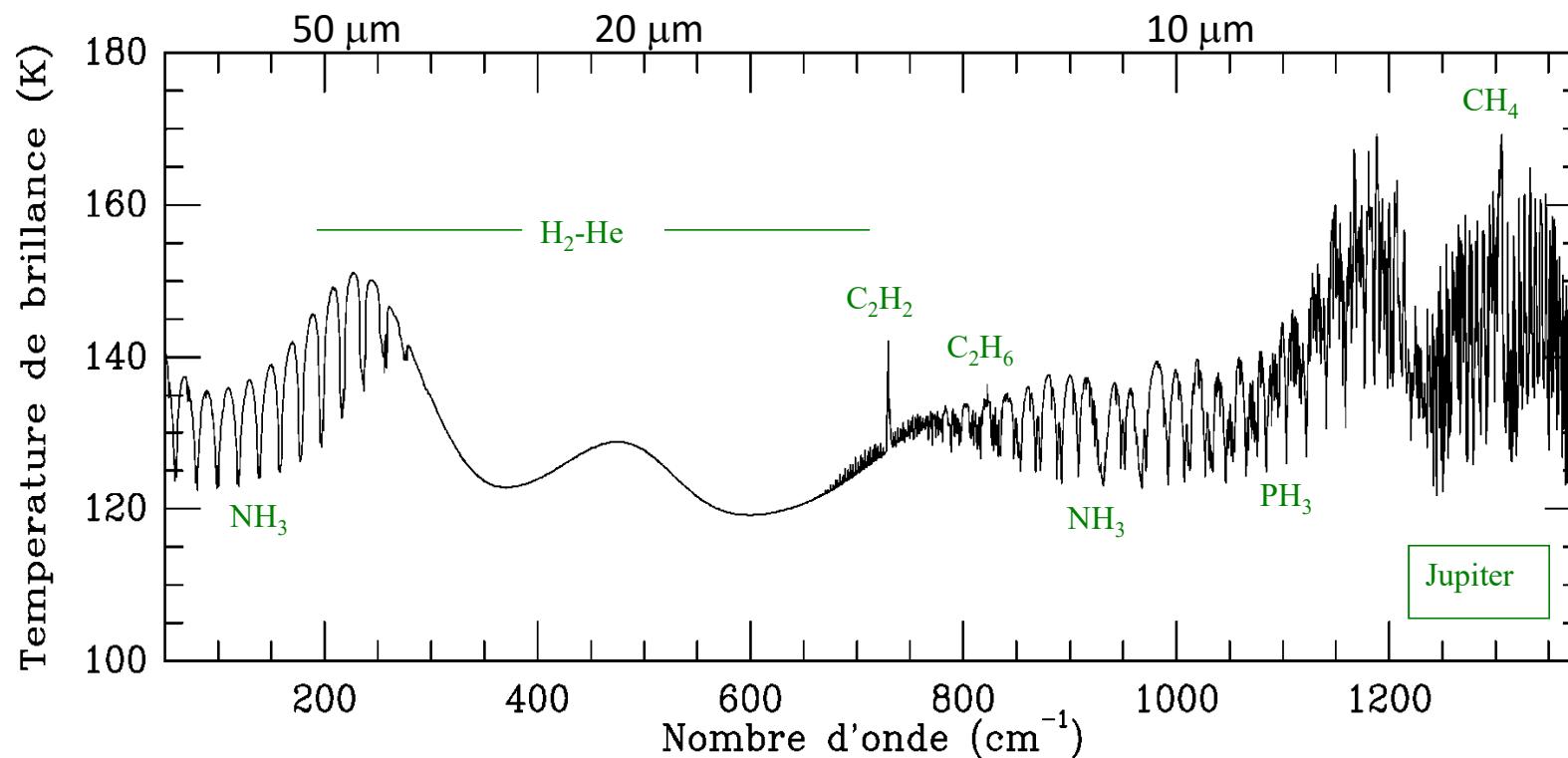


More Mars spectra for your inspection...



What can we do from this Jupiter spectrum?

Note: the abundance of CH₄ in Jupiter is known independently



Occultations

- Principle

- Measure of extinction of signal when a source of light (Sun, star, radio signal from a spacecraft) disappears behind the limb of a planet with atmosphere. Extinction can be due to:

- Refraction (visible, radio)

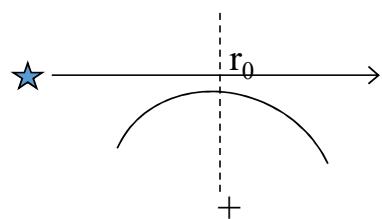
Extinction → atmospheric refractivity profile (if composition is known)

→ Profiles of concentration $n(z)$, pressure $p(z)$, and temperature $T(z)$

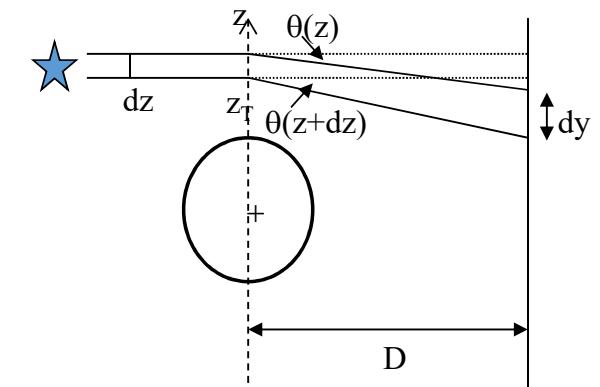
- Gas absorption (UV)

$$I = I_* e^{-\tau_s}$$

$$\tau_s \sim \sigma_i n_i \sqrt{2\pi r_0 H}$$



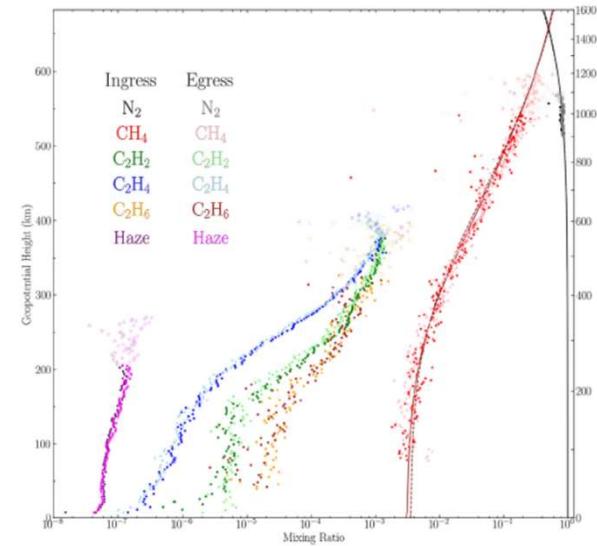
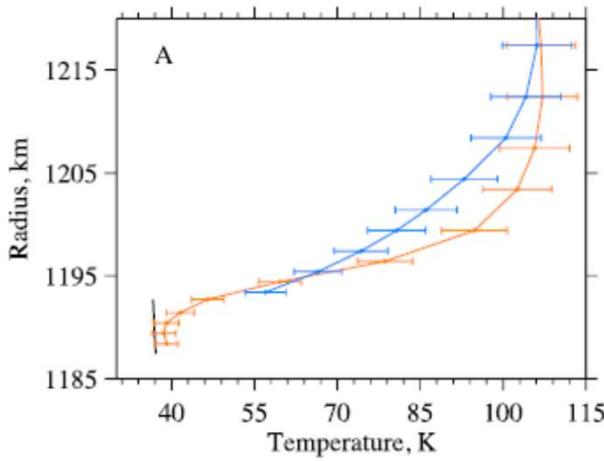
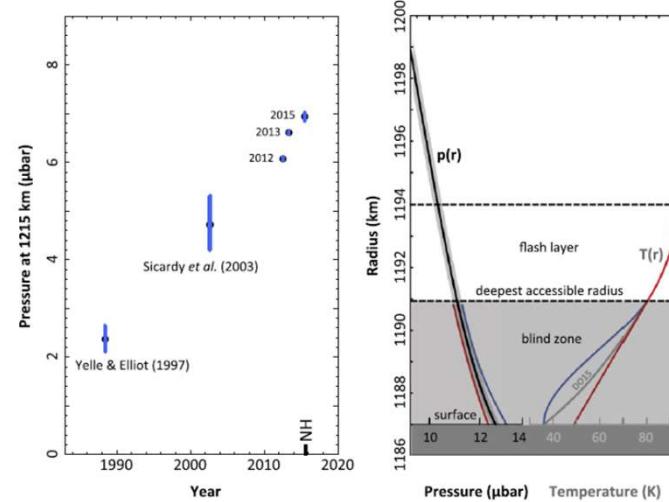
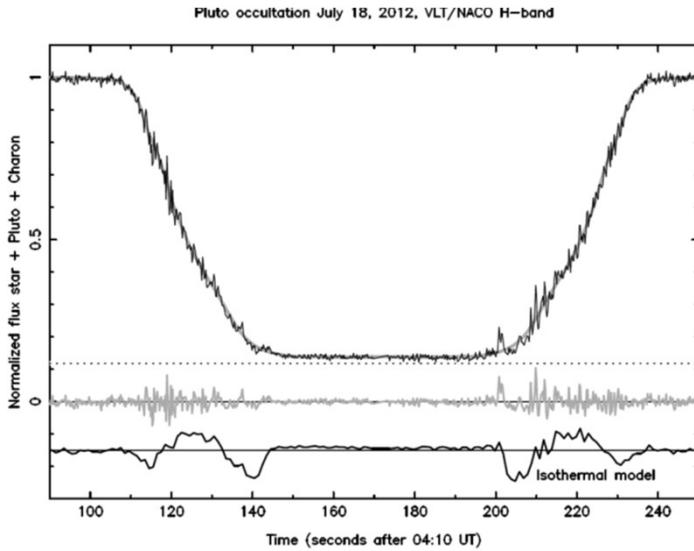
(horizontal viewing → increase in path → in sensitivity)



- Temperature and composition information

- Probed levels: from ~bar to ~nanobar (1 mbar – 1nbar in UV, 0.1-100 μ bar in optical, 1 mbar-1 bar in radio)
 - Exquisite spatial and vertical resolution
 - Limited occurrence; can probe only planetary limbs

Probing Pluto's atmosphere ($p=12 \mu\text{bar}$) from visible, radio and UV occultations

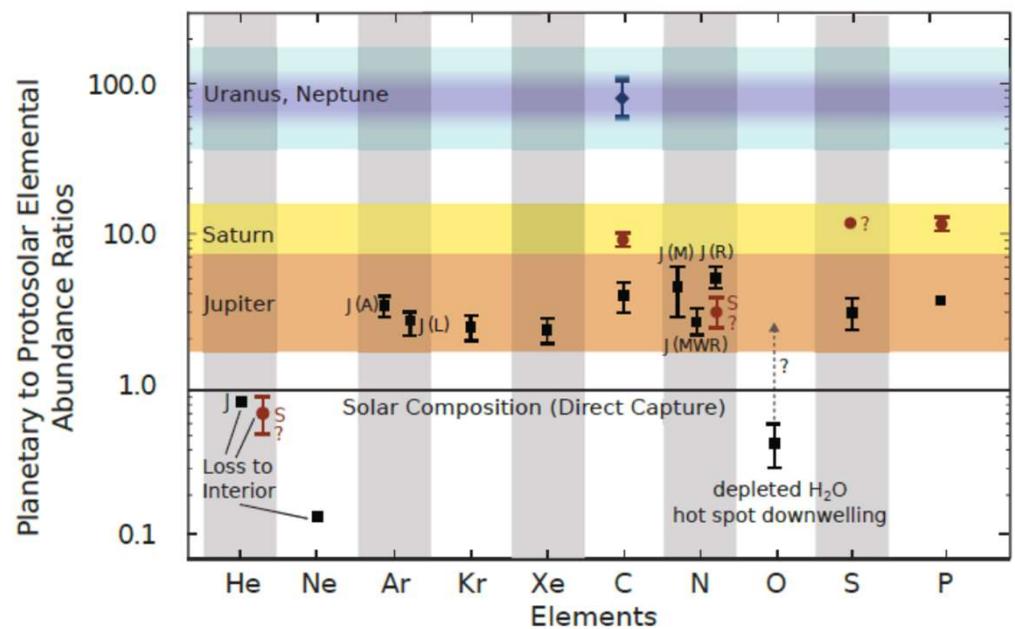
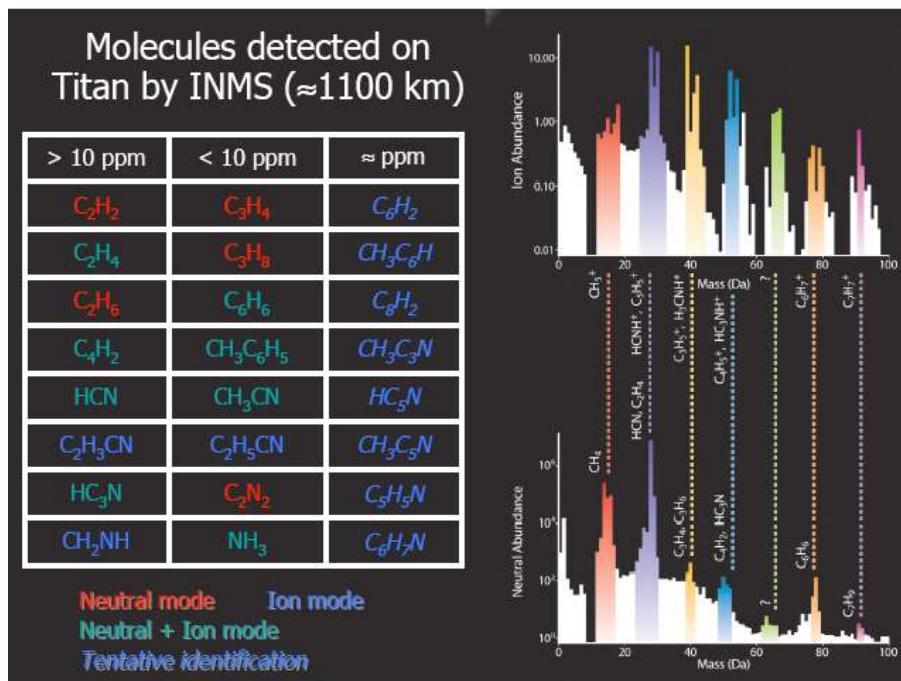


Stellar occultations from Earth

Radio and UV occultations from New Horizons probe (2015)

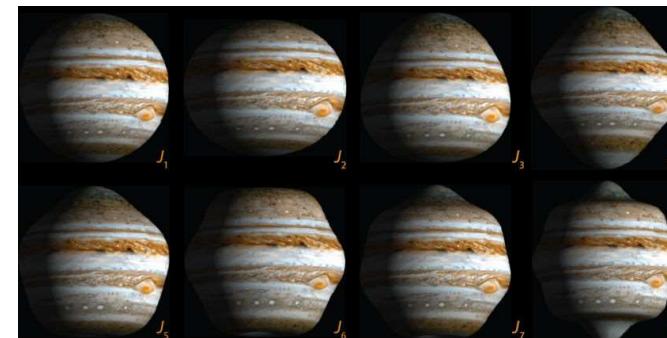
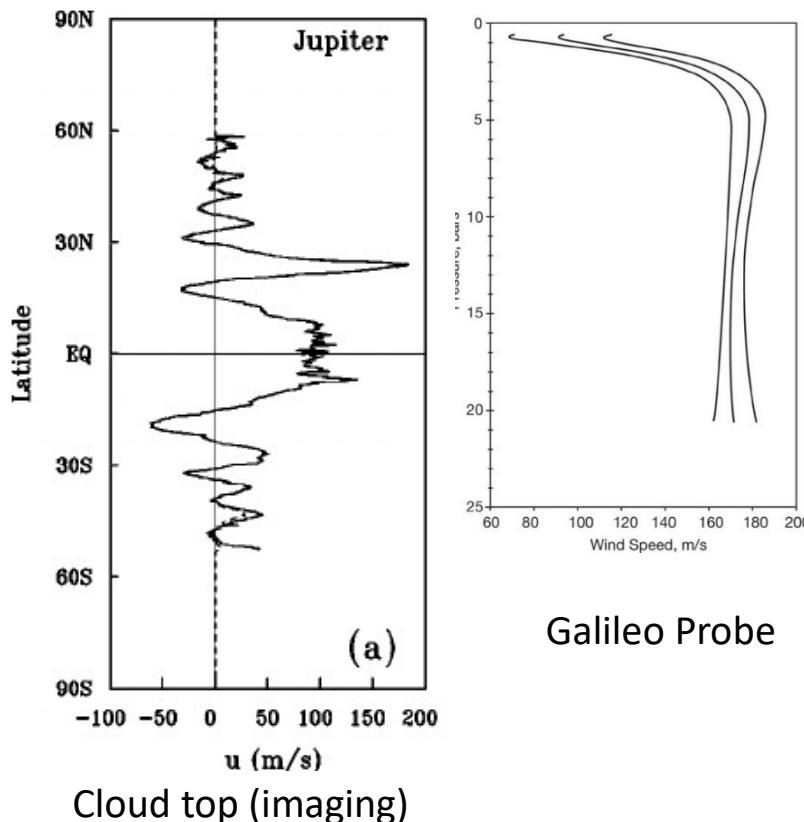
Probing solar system atmospheres: other techniques

- In situ measurements (descent probes...), many instruments, in part.
 - Atmospheric / cloud instrument, measuring directly P, T, cloud properties
 - Mass spectrometer – extremely useful for composition, for species not observable for spectroscopy
 - Noble gases (He, Ar)
 - Very low abundance gases



Probing solar system atmospheres: other techniques

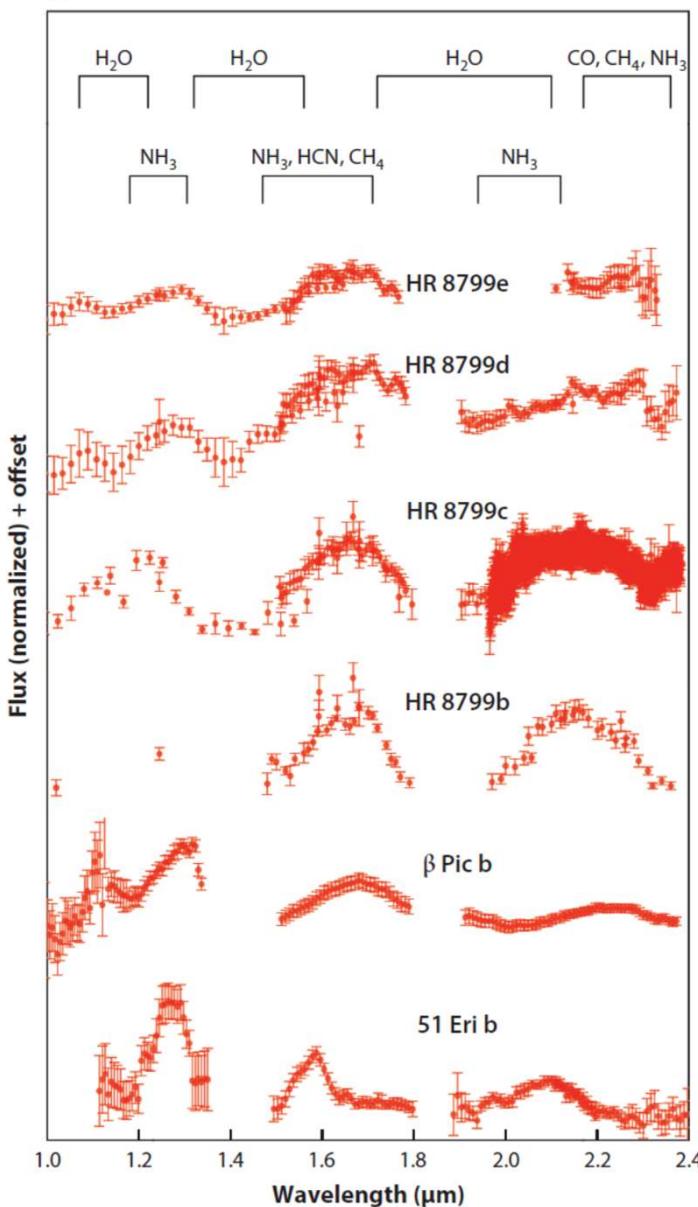
- Wind measurements below cloud levels from spacecrafts
 - Tracking of descent probes (Galileo, Huygens...)
 - Gravity field measurements (JUNO, Cassini Grand Finale)



Winds, if deep enough, impact mass distribution within planet hence structure of gravitational field.

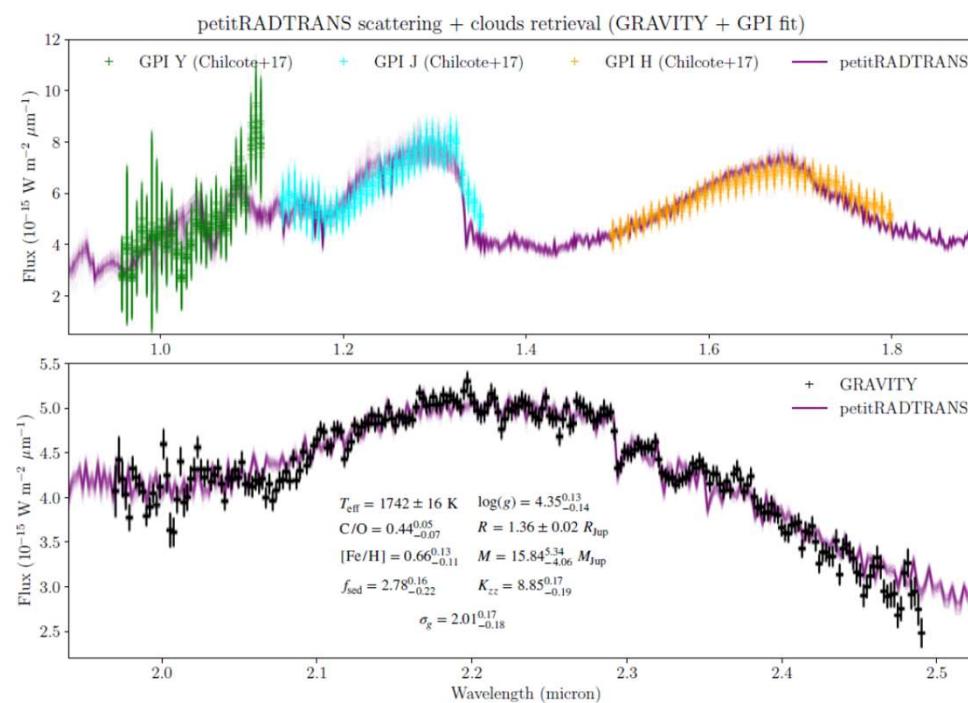
Result: winds in Jupiter (Saturn) extend ~ 3000 km (~ 9000 km) below cloud tops, corresponding to a layer as thick as 4 % (15 %) of the planet radius. Evidence of the internal energy source as motor of deep circulation

Methods of investigation of planetary atmospheres (Extrasolar Planets)

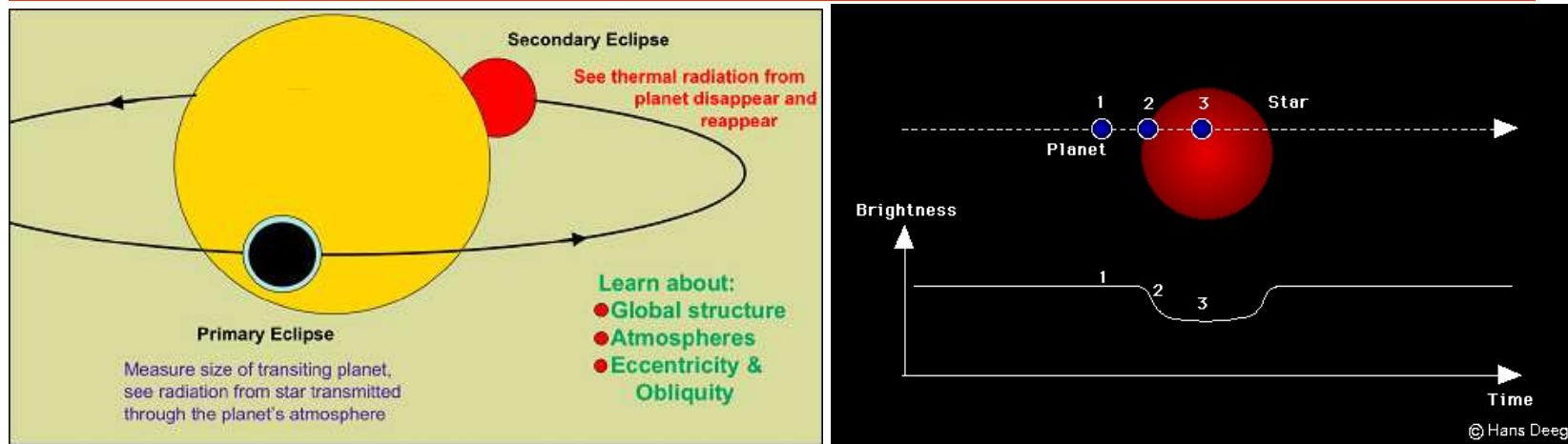
b

« Ordinary » spectroscopy of exoplanetary atmospheres

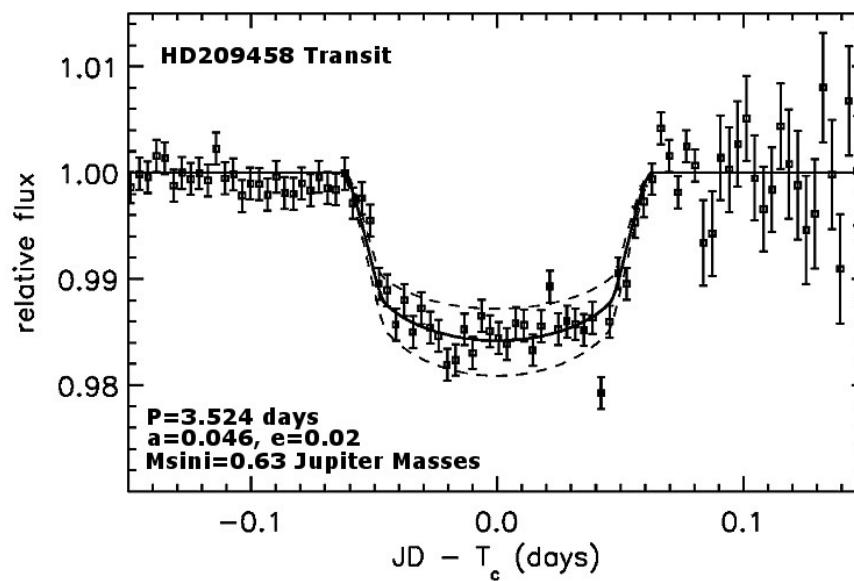
Young and hot objects, sufficiently distant from star
Direct spectroscopy, but requires high sensitivity



Exoplanetary atmospheres: Transit spectroscopy and phase curves

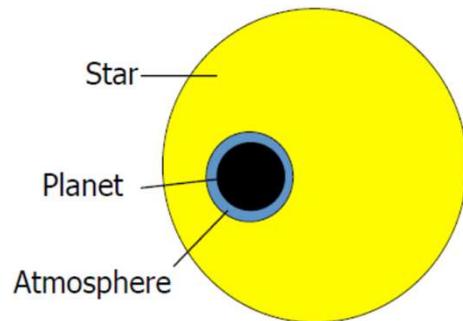


$$\delta = (R_P/R_\star)^2$$



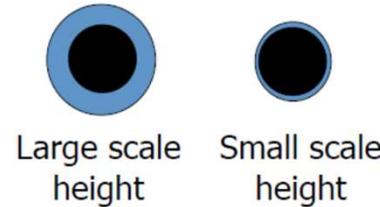
Idea: do this
spectroscopically
to see if the planet « size »
changes with wavelength

Atmospheric sounding of transiting exoplanet



Scale Height

$$H = \frac{kT}{g\mu}$$



Atmospheric opacity in horizontal viewing $\tau_\lambda(z) \sim \sigma_\lambda n(z) \sqrt{2\pi R_P H}$

Effective absorption radius
(if σ_λ is independent of z)

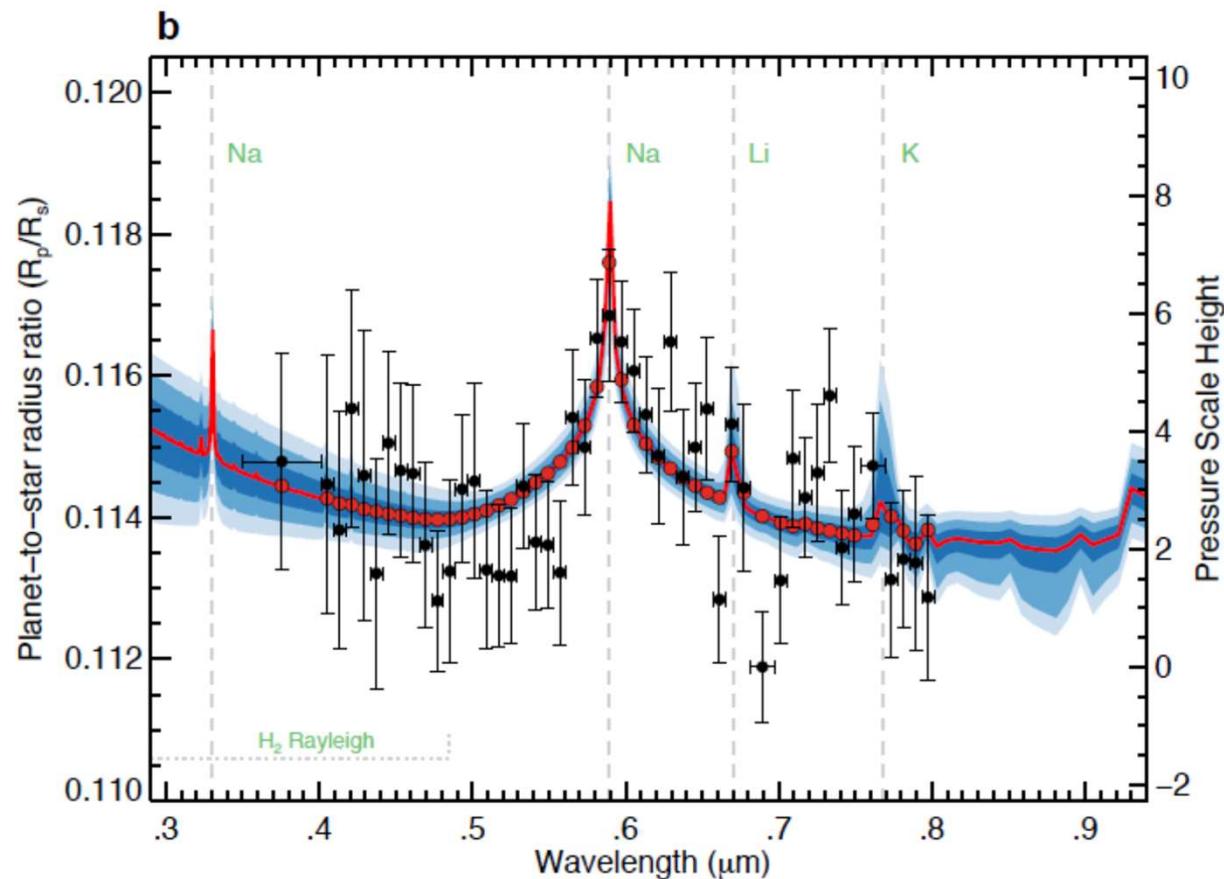
$$R_{eff(\lambda)} \sim R_P + z (\tau_\lambda = 1) \quad z (\tau_\lambda = 1) \sim H \ln (\sigma_\lambda n_0 \sqrt{2\pi R_P H}) \sim \alpha H$$

Differential absorption depth $\Delta\delta = (R_{eff}/R_\star)^2 - (R_P/R_\star)^2 \sim \frac{2R_P \alpha H}{R_\star^2} = 2 \alpha \delta \frac{H}{R_P}$ **note proportionality to H**

Exercise: calculate effect for a hot Jupiter (1300 K), and an Earth

Note also that $\frac{dz}{d\lambda} \approx H \frac{d \ln \sigma}{d\lambda}$, so if σ_λ can be calculated (e.g. opacity due to haze scattering), measuring $z(\lambda)$ yields H

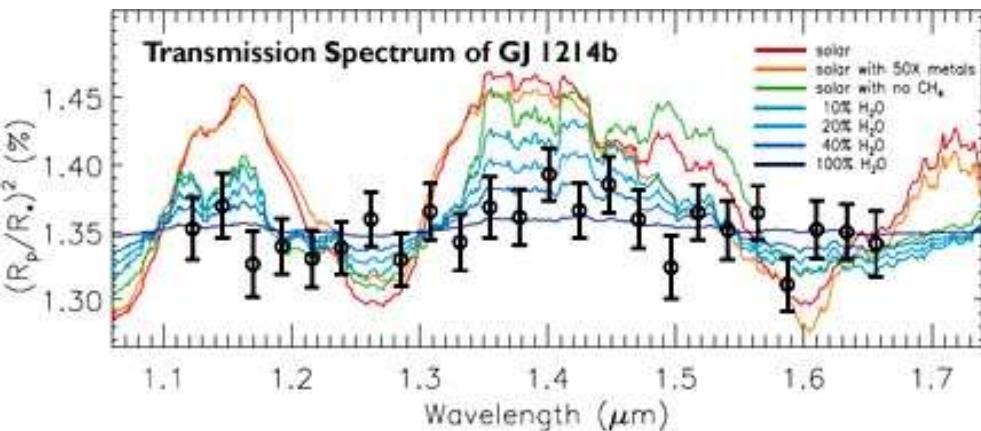
Atmospheric sounding of transiting exoplanet



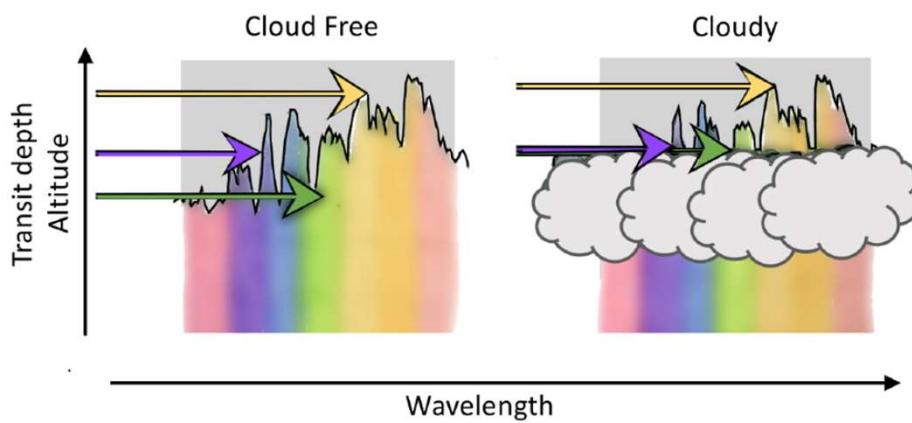
WASP-96b: evidence for haze scattering and Na emission

Detected species: H₂O, CH₄, CO₂, CO, TiO, VO, H, He, Fe, Ca, Na, Mg, other atoms and ions

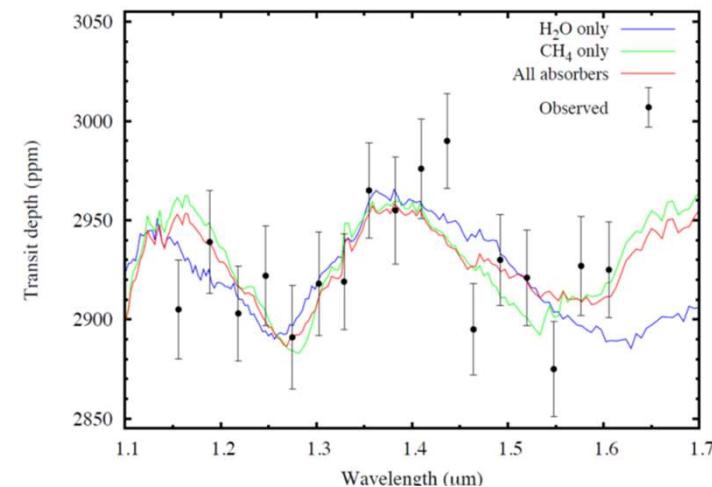
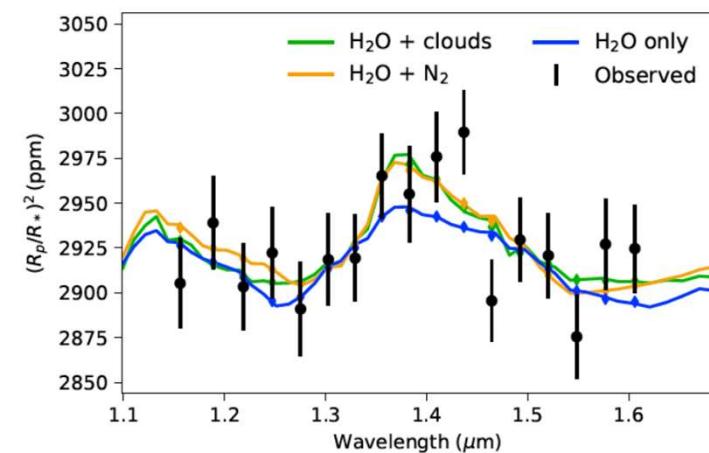
Dangers / ambiguities



Super Earth ($6.5 M_E$). No signature \rightarrow small scale height, heavy gases?
BEWARE OF CLOUDS!



K2-18b: a super Earth ($9 M_E$) in habitable zone of an M star
BEWARE of ambiguities at low resolution!



K2-18b: methane, not water !

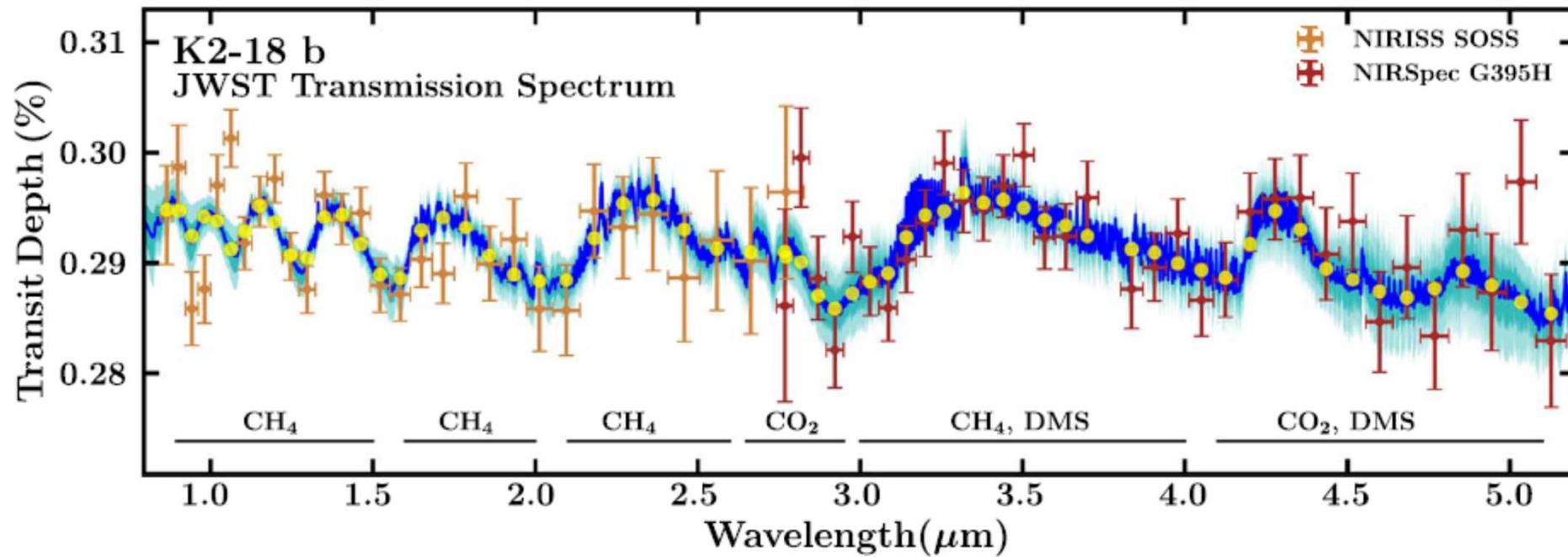


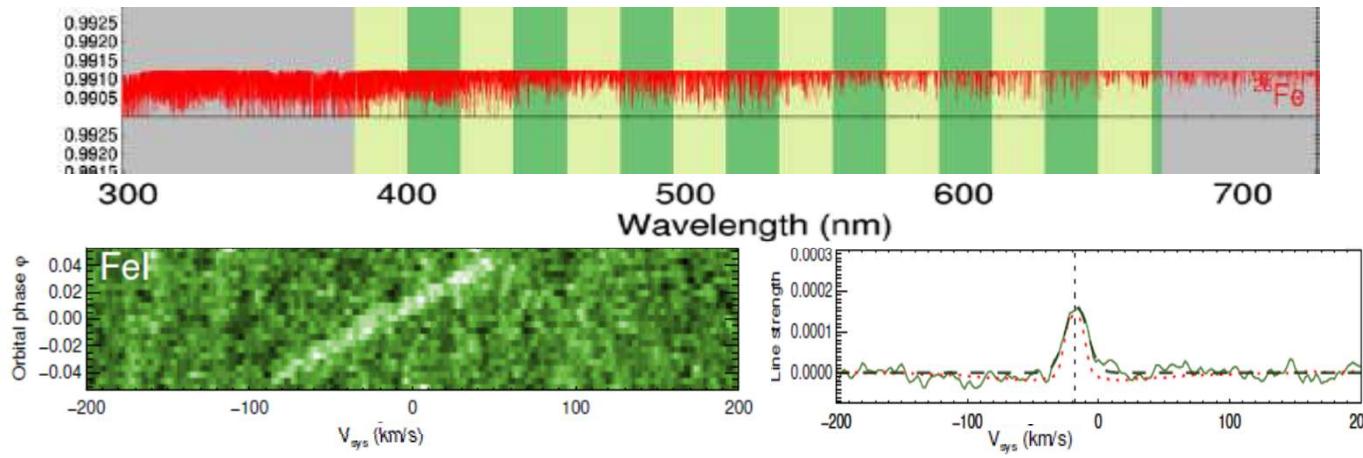
Figure 3. The transmission spectrum of K2-18 b. The observed JWST spectrum and retrieved model fits are shown for the one-offset retrieval case discussed in Section 3. The data in orange show our NIRISS spectrum between 0.9 and 2.8 μm , and those in dark red show our NIRSpec G395H spectrum between 2.8 and 5.2 μm . The spectra are binned to $R \approx 25$ and $R \approx 55$ for NIRISS and NIRSpec, respectively, for visual clarity. The retrievals are conducted on the native resolution spectra, resulting in the best-fit reduced $\chi^2_{\nu} = 1.080$. The NIRSpec spectrum is vertically offset by -41 ppm, corresponding to the median retrieved offset in the one-offset retrieval case. The blue curve shows the median retrieved model spectrum, while the medium- and lighter-blue contours denote the 1σ and 2σ intervals, respectively. Yellow points correspond to the median spectrum binned to match the observations. The prominent molecules responsible for the features in different spectral regions are labeled.

Exoplanet spectroscopy: synchronous detection

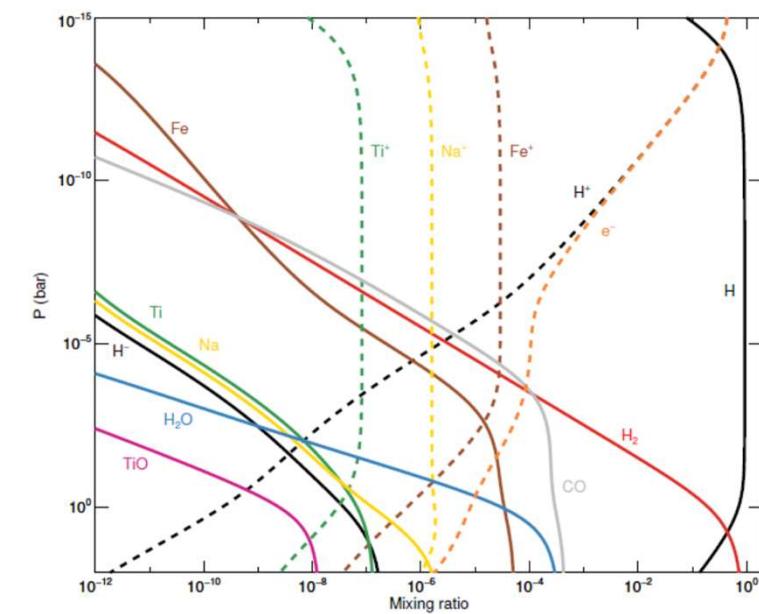
- Transit spectroscopy is triply difficult
 - (a) planet/star surface contrast can be weak ($\delta \sim 1\%$ for a Jupiter around solar type star; 0.01% for Earth around Sun; 0.2% for SuperEarth ($2x R_{\text{Earth}}$) around M star ($0.4 R_{\text{sun}}$)).
 - Atmospheric differential may be weak ($\Delta\delta / \delta \sim 1\%$)
 - Lines from atmosphere may be weak if target molecule/atom is weakly abundant
 - So, a simple ratio « in-transit » / « out-transit » shows nothing at first glance.
- But
 - (a) the target species may have *many* (thousands) lines
 - (b) if the planet orbit is well known, its line-of-sight velocity across the transit is predictable.
- Method
 - Compute a synthetic spectrum (« template ») of the expected spectrum of each species
 - Split the transit data in time bins, as a function of ‘phase’
 - For each bin, correlate in velocity the template with the observed spectrum
 - If the correlation tracks the planet velocity, bingo!
 - Resumming the lines in the planet rest frame will show the « total » line.

Exoplanet spectroscopy: synchronous detection

- Fantastically successful!
- A new emerging population: the ultra-hot Jupiters ($T_{eq} > 3000$ K)
 - Most molecules are thermally dissociated in atoms and ions
Fe, Na, Mg, Ca, Fe+, Ti+, Cr, Sc, Y... detected...

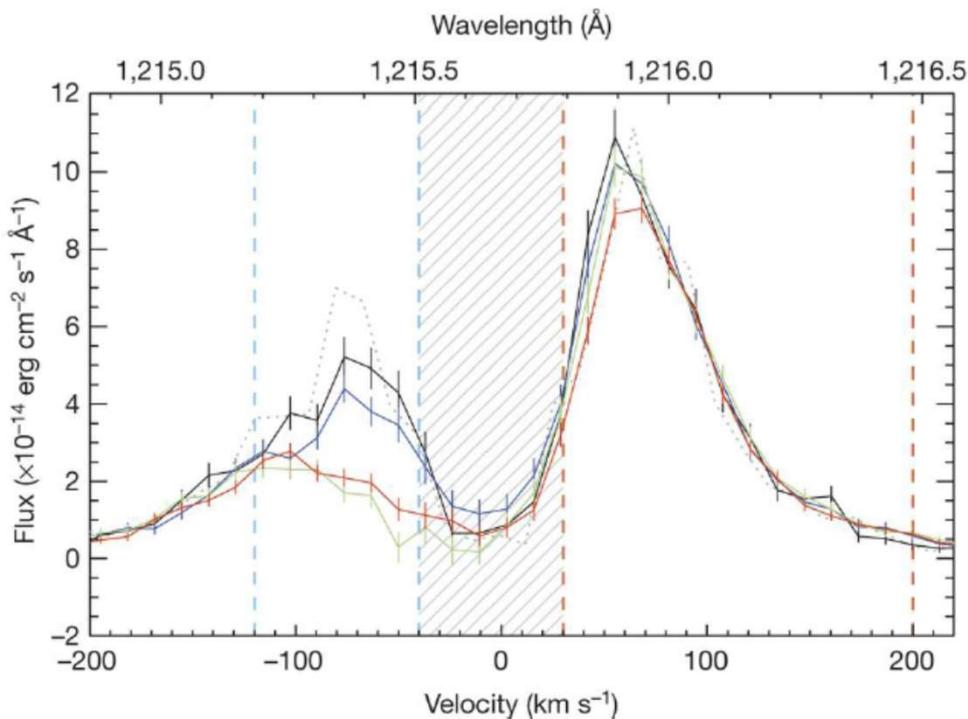


Kelt 9-b, at 0.03 au from A star, $T_{eq} = 4050$ K!!

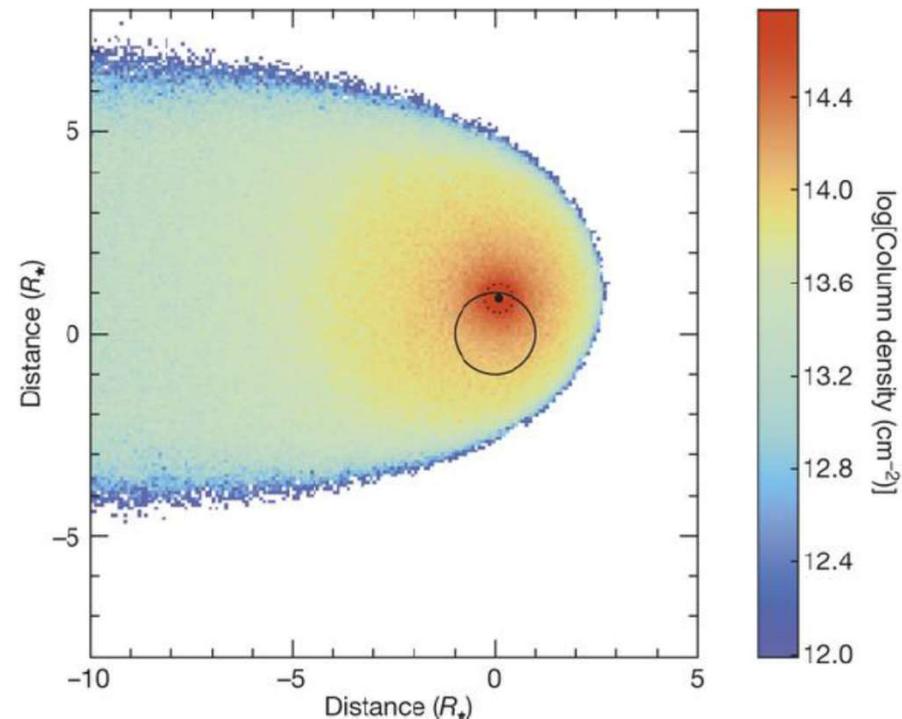


Transit spectroscopy: direct evidence for current atmospheric escape in exoplanets

GJ436b (hot Neptune, $R = 4 R_E$, $a = 0.028$ au, $T_{eq} \sim 700$ K)

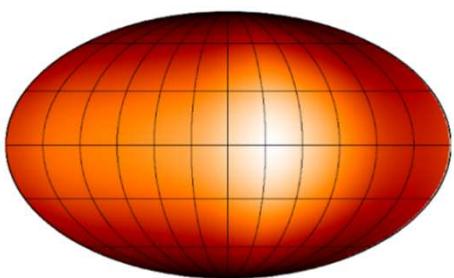
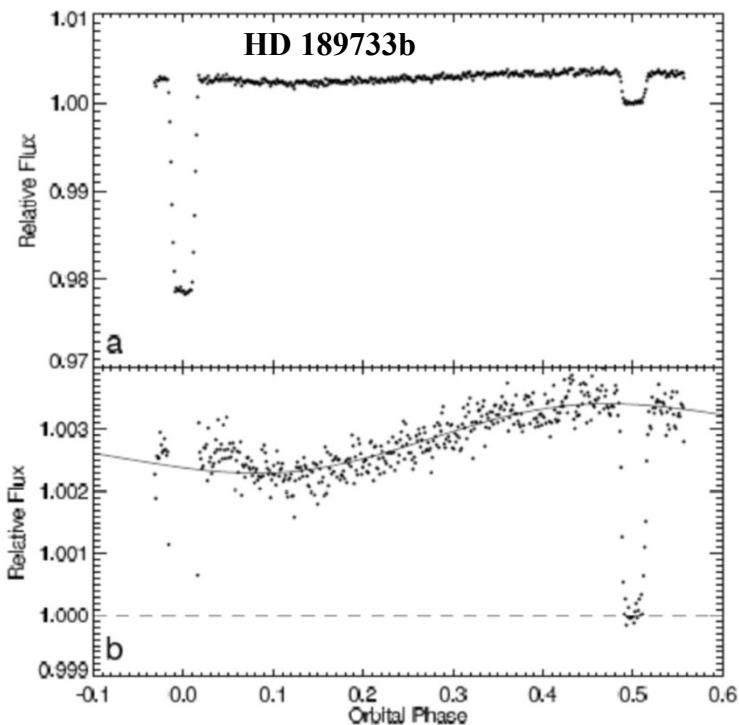


Transit spectroscopy
 Absorption depth $\sim 50\%$ \Leftrightarrow the extended H atmosphere has
 an effective (obscuring) area $\sim 50\%$ of the star !

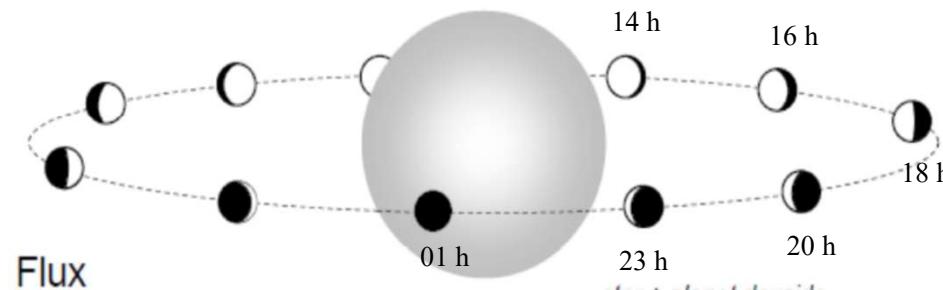


*Ehrenreich+2015
 Bourrier+2016*

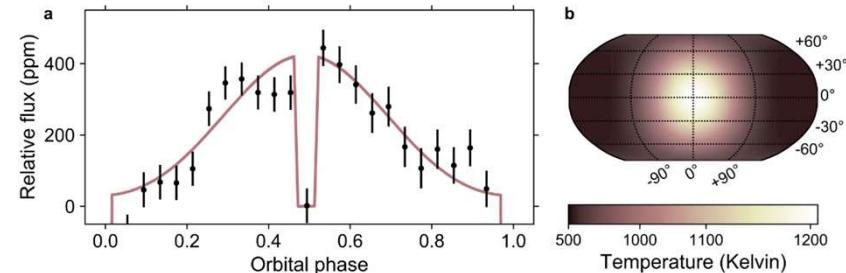
Exoplanet thermal emission: phase curves and temperature maps



- ΔT (day-night) ~ 240 K (only)
-Peak thermal flux NOT at secondary eclipse, but before
 \rightarrow Afternoon hours hotter than noon
 \rightarrow Evidence for day-to-night winds

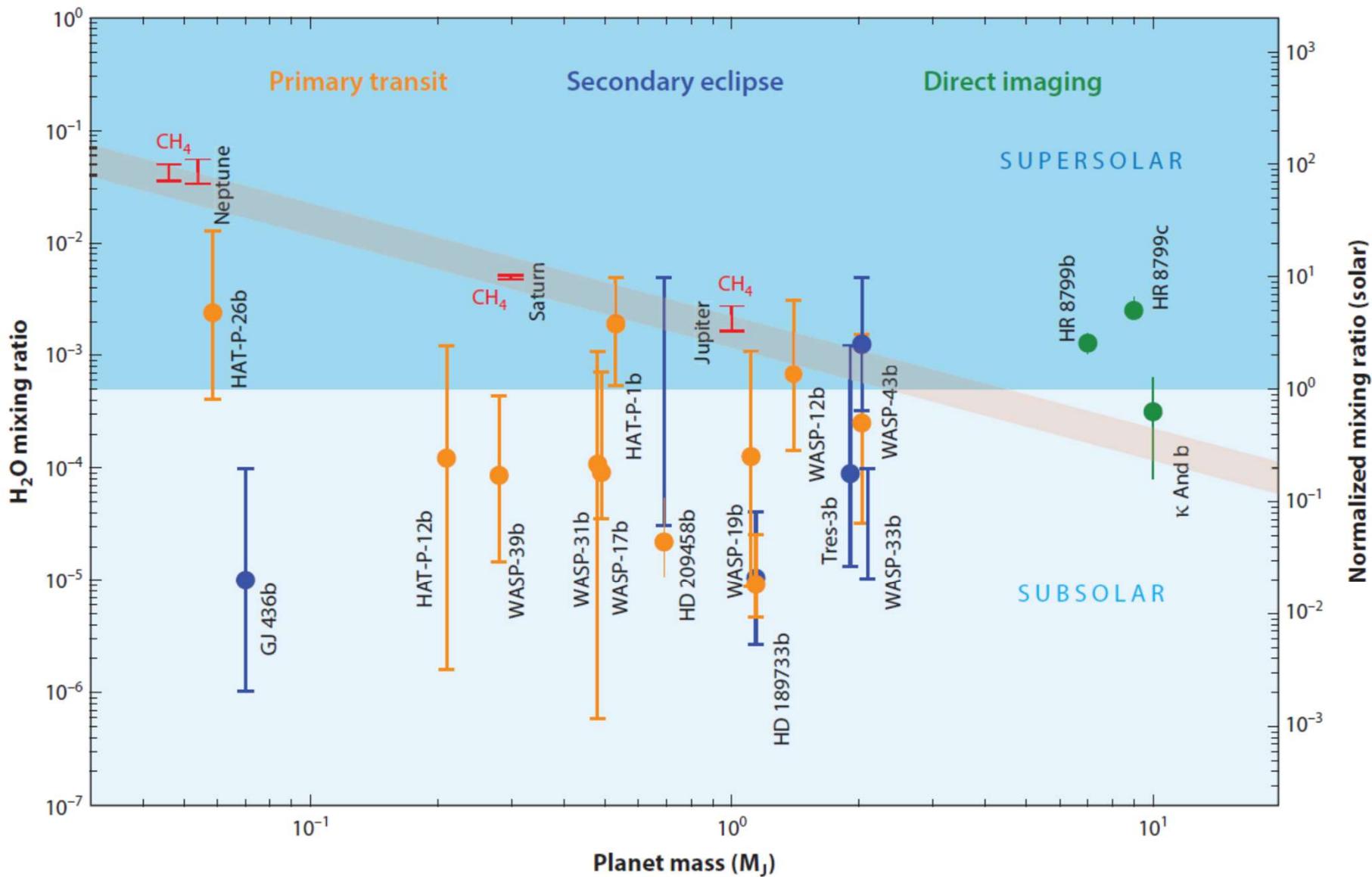


TESS discovery LHS 3844b is the first terrestrial exoplanet accessible for phase curve observations



- $T_{\text{day}} = 1080 \pm 40$ K
- T_{night} is consistent with 0 K
- No hotspot offset
- $R_p = 1.3 R_\oplus$
- $P = 11$ hours
- $T_{\text{eq}} = 1000$ K

Kreidberg et al. 2019, Nature



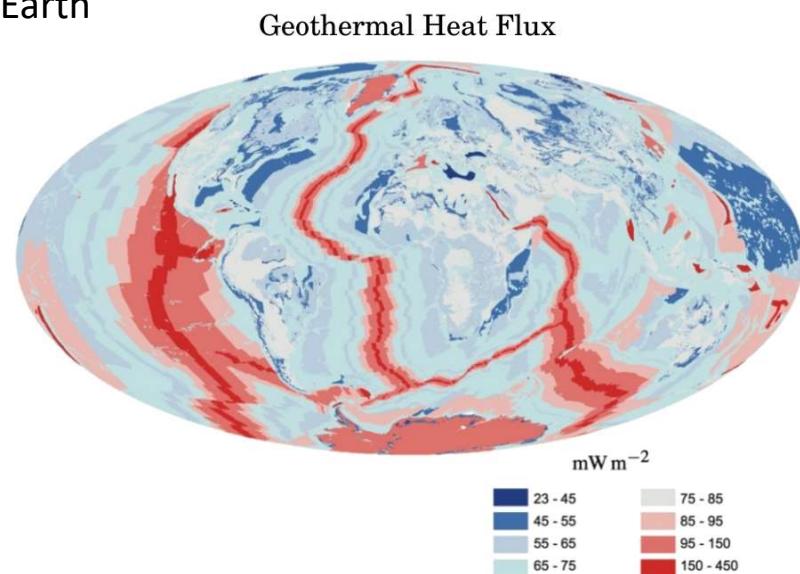
Energy budget and vertical structure in planetary atmospheres

Energy sources, sinks, and exchanges in an atmosphere

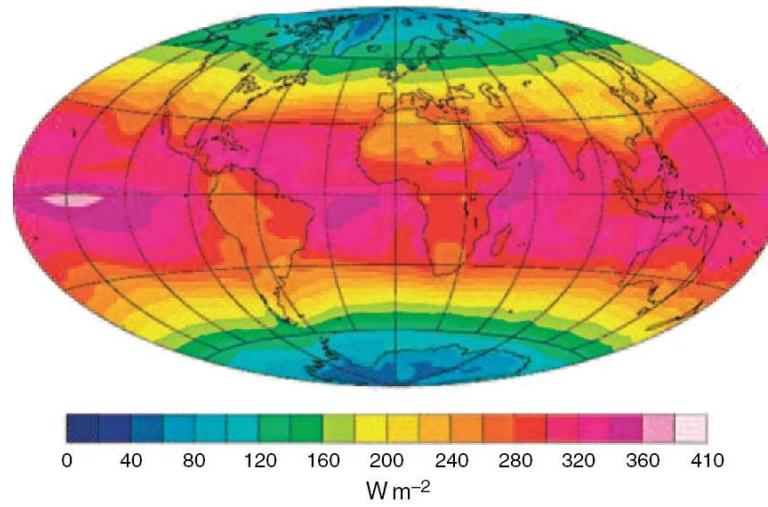
- Energy sources
 - Solar/stellar radiation (UV, visible, near-IR; $\lambda < \sim 4 \text{ } \mu\text{m}$ in solar system) – absorbed at surface and/or within atmospheres
 - Internal heat
 - Giant planets – due to cooling/contraction. Comparable to solar heating (cf exercise)
 - Volcanism (negligible on Earth, not on Io)
 - Non-radiative energy sources in atmosphere (e.g. impact of energetic particles, solar wind on upper atmospheres)
- Energy sinks
 - Thermal radiation from atmosphere towards space and surface (thermal infrared [$\lambda \geq 5 \text{ } \mu\text{m}$])
- Energy exchanges between atmospheric layers
 - Radiation (absorption of radiation by other layers)
 - Convection (low-amplitude, macroscopic-scale, vertical motions)
 - Thermal conduction (i.e. exchanges by molecular collisions; operates on distances \sim mean free path)
- **An atmosphere's thermal structure reflects its thermal budget**
- Radiative exchanges are usually dominant

Volcanism on Earth and Io

Earth

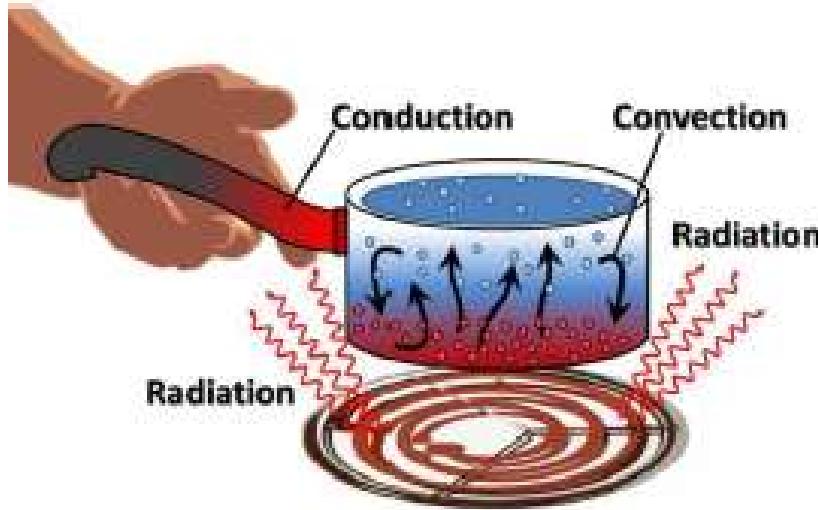


Absorbed Solar Radiation Flux



Io – volcanism due to tidal heating by Jupiter is about 2.5 W m^{-2} ; while absorbed solar flux is $\sim 8 \text{ W m}^{-2}$...

Radiative, convective, and conductive exchanges



TERRE

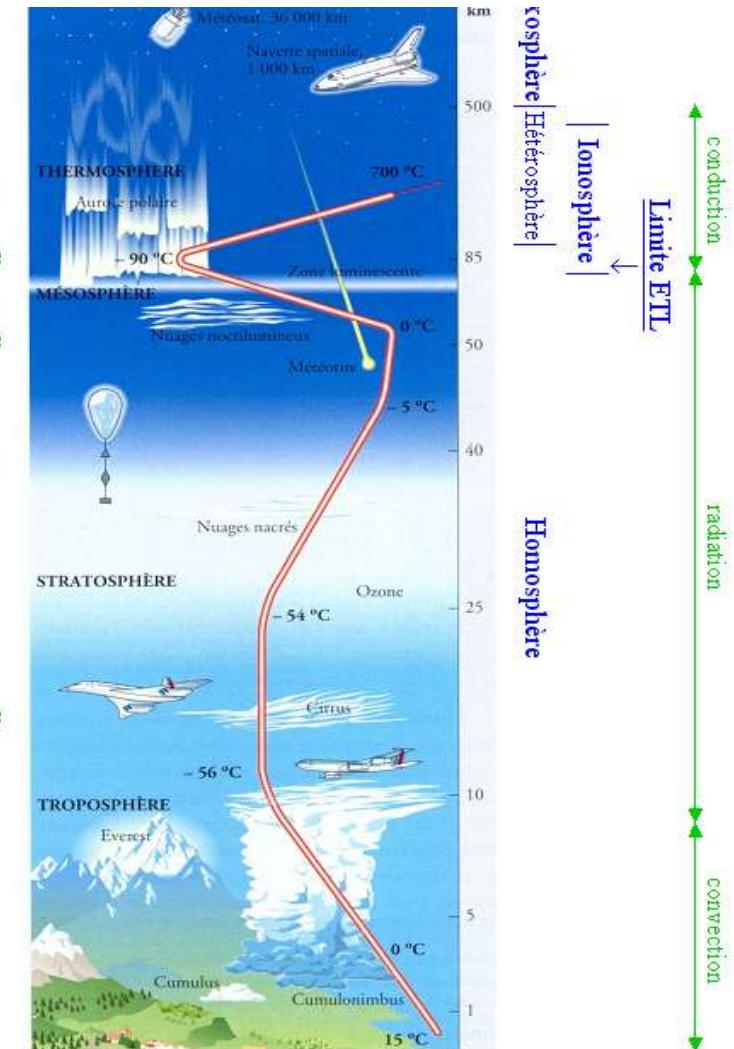
Thermosphère

Mésopause
Mésosphère
Stratopause

Stratosphère

Tropopause

Troposphère



In planetary atmospheres

- Radiation: efficient everywhere except at too low pressures ($\sim 1-100$ nbar)
- Convection: present only if atmospheric gradient is unstable (steeper than adiabatic)
- Thermal conduction: important only at low pressures (< 100 nbar), where mean free paths are long, and radiation exchanges become unefficient

Heat equation

- General formulation

$$\rho C_p^* \frac{\partial T}{\partial t} = Q(z) - \frac{\partial \Phi}{\partial z}$$

- $Q(z)$ is the local heating source per volume (e.g. W m^{-3}).
- $\Phi(z)$ is the energy flux (W m^{-2}), which includes

- **conduction flux:** $\Phi_c = -K_c \frac{\partial T}{\partial z}$ $K_c = \text{thermal conductivity coefficient } (\text{W m}^{-1} \text{ K}^{-1})$

- **convection flux** $\Phi_K = -K \rho C_p^* \left(\frac{dT}{dz} - \Gamma_{ad} \right)$ **only if** $dT/dz < \Gamma_{ad}$, i.e if $\left| \frac{dT}{dz} \right| > \left| \Gamma_{ad} \right|$
 (super-adiabatic = unstable atmosphere). Upward flux.

K : eddy diffusion coefficient ($\text{m}^2 \text{ s}^{-1}$)

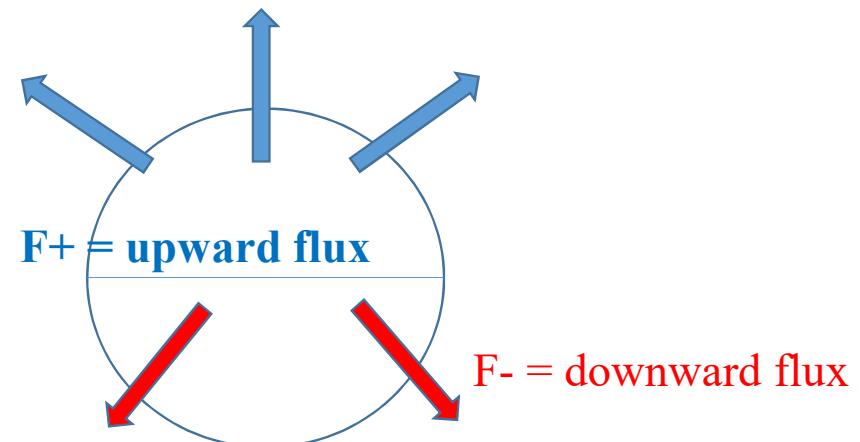
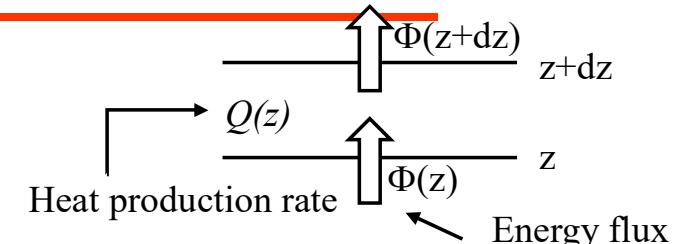
C_p^* : specific heat capacity ($\text{J kg}^{-1} \text{K}^{-1}$)

- Radiative flux = solar flux + thermal flux**

$$\Phi_{rad} = \Phi_{sol} + \Phi_{th}$$

$$F_v = \int I_v \cos(\theta) d\Omega$$

$$\Phi_{rad} = F+ - F-$$

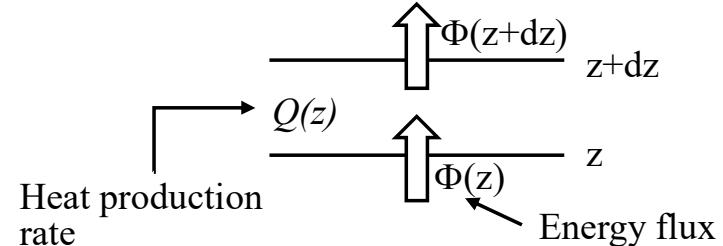


Radiative equilibrium

$$\rho C_p^* \frac{\partial T}{\partial t} = Q(z) - \frac{\partial \Phi}{\partial z}$$

- Steady state: $dT/dt = 0$
- No local heating terms: $Q(z) = 0$.
- Convective and conductive fluxes can be neglected

$$\Phi(z) = \Phi_{\text{rad}}(z)$$

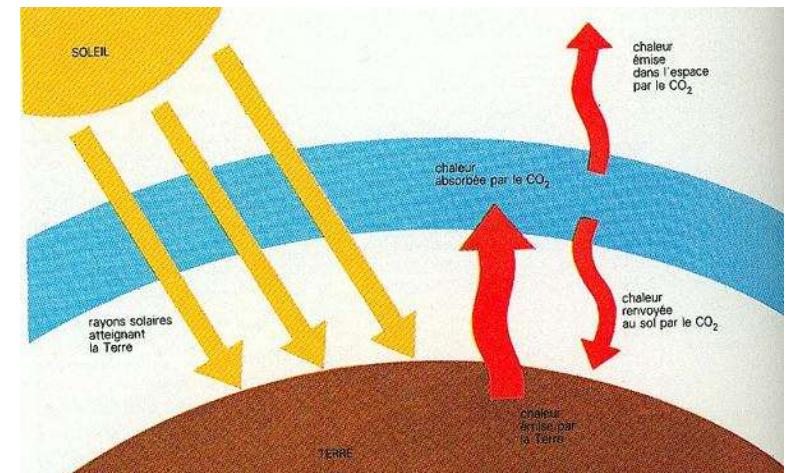


- This implies: $\Phi_{\text{rad}}(z) = \text{cte} = \Phi_{\text{sol},+}(z) + \Phi_{\text{th}+}(z) - [\Phi_{\text{sol},-}(z) + \Phi_{\text{th},-}]$

Tropospheres and greenhouse effect

- « Idealized » troposphere
 - No absorption of solar radiation within atmosphere (zero opacity at solar wavelengths)
 - Solar radiation is absorbed only at surface
 - In contrast, the atmosphere has a non-zero opacity for thermal radiation (τ), therefore, it can be heated by the surface (or, for giant planets, by the internal source).
 - Heat absorbed by atmosphere is re-emitted as upward and downward thermal radiation → additional heating term for the surface : *greenhouse effect*
- Expression of radiative equilibrium $\Phi_{\text{sol}}(z) = \Phi_{\text{rad}}(z) + \Phi_{\text{th}}(z)$
 - No absorption of solar radiation → $\Phi_{\text{sol}}(z) = \text{constant}$
 - D'où $\Phi_{\text{th}}(z) = \Phi_{\text{th},+}(z) - \Phi_{\text{th},-}(z) = \text{constant}$
 - The constant is $= \sigma T_{\text{eff}}^4$.
 - If no internal heat:

$$\Phi_{\text{th},+}(z) - \Phi_{\text{th},-}(z) = \sigma T_{\text{eq}}^4 = \text{absorbed solar flux } F_0$$



A qualitative approach to greenhouse effect

- The outgoing thermal radiation (« OLR » = « outgoing longward radiation ») is equal to the absorbed solar flux (for a planet with no internal source): $OLR = F_{abs}$

$$\sigma T_{eq}^4 = (1 - A_b) F_{sol} / 4$$

- If atmosphere transparent to thermal radiation, radiation comes from surface and $T_{surf} = T_{eq}$
- If atmosphere has opacity, the surface must emit more radiation to maintain the same OLR, hence $T_{surf} > T_{eq}$.

Earth

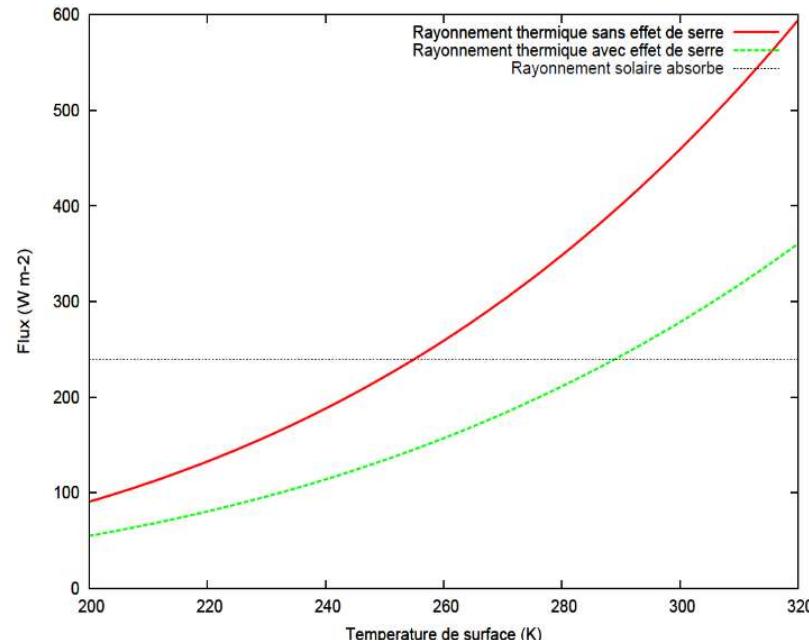
$$F_{sol} = 1368 \text{ W m}^{-2}$$

$$A_B = 0.31$$

$$\rightarrow F_{abs} = 235 \text{ Wm}^{-2}$$

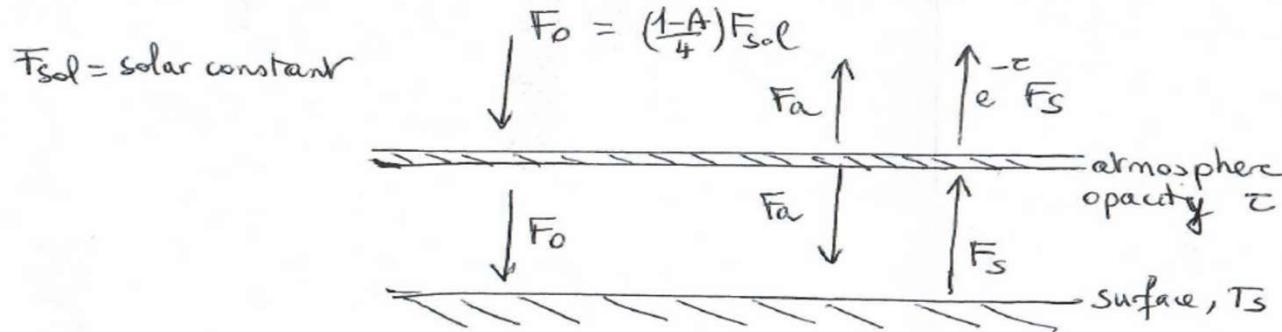
No greenhouse: $T_{surf} = 255 \text{ K}$

With greenhouse: $T_{surf} = 286 \text{ K}$



A first quantitative model of greenhouse effect

One layer atmosphere – entirely transparent to solar radiation
-- with opacity τ for thermal radiation



- Radiative equilibrium: $\phi_{\text{th}, \oplus} - \phi_{\text{th}, \ominus} = F_0 = \sigma T_{\text{eq}}^4$

- Top of atmosphere

$$\phi_{\text{th}, \oplus} = F_a + e^{-\tau} F_s$$

$$\phi_{\text{th}, \ominus} = 0$$

- Surface

$$\phi_{\text{th}, \oplus} = F_a$$

$$\phi_{\text{th}, \ominus} = F_0$$

$$F_0 = F_a + e^{-\tau} F_s$$

$$F_0 = F_s - F_a$$

$$\} \Rightarrow 2F_0 = F_s (1 + e^{-\tau})$$

$$2\sigma T_{\text{eq}}^4 = \sigma T_s^4 (1 + e^{-\tau})$$

$$\Rightarrow T_s = \left(\frac{2}{1 + e^{-\tau}} \right)^{1/4} T_{\text{eq}}$$

$T_s > T_{\text{eq}} \rightarrow$ greenhouse, but $T_s \max = 2^{1/4} T_{\text{eq}} \approx 1.19 T_{\text{eq}}$

could work for the Earth ($T_{\text{eq}} = 255 \text{ K}$, $T_s = 288 \text{ K}$)

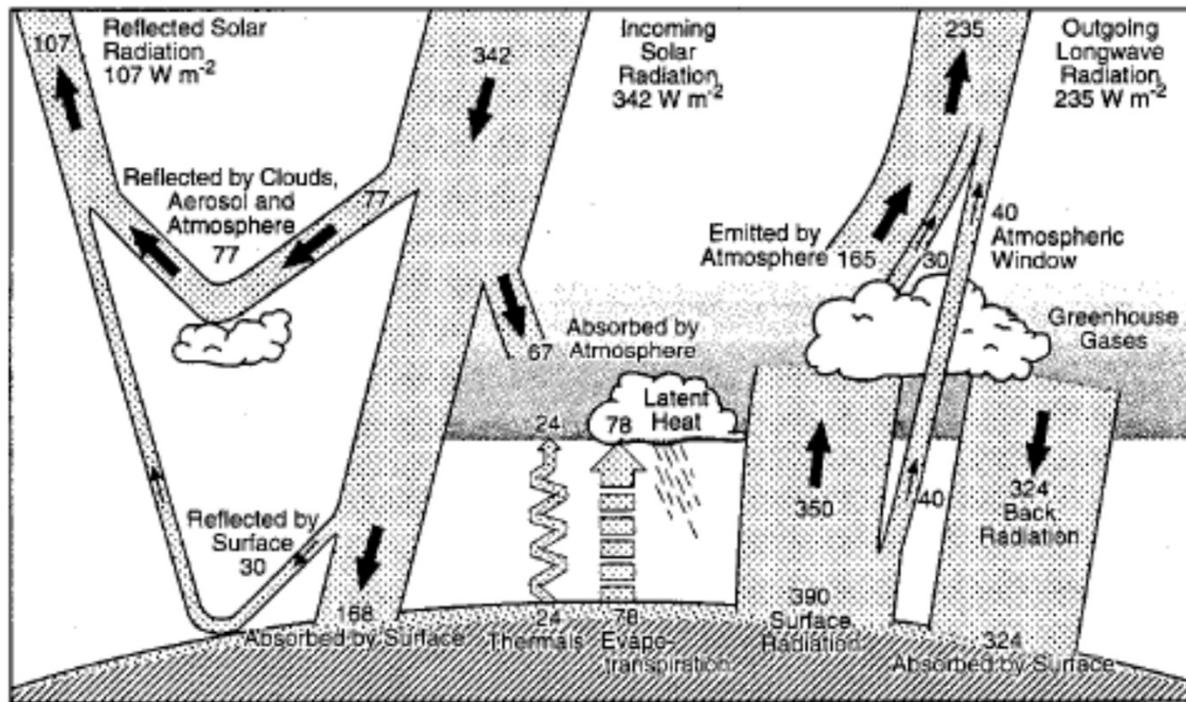
but not for Venus ($T_{\text{eq}} = 231 \text{ K}$, $T_s = 733 \text{ K}!$)

Greenhouse effect in a multi-level atmosphere

- $\Phi_{\text{th},+}(z) - \Phi_{\text{th},-}(z) = \sigma T_{\text{eq}}^4$
- Assume the opacity τ in the thermal range is independent of wavelength
- One can show that the atmospheric temperature profile is defined by $T^4(z) = T_{\text{eq}}^4 \left(\frac{1}{2} + \frac{3}{4} \tau(z) \right)$
 - $T(z)$ increases downwards (since $\tau(z)$ does): *troposphere. Atmosphere is heated « from below ».*
 - $T(z)$ exceeds T_{eq} when τ becomes $> 2/3$.
- The ground temperature T_{surf} is given by $T_{\text{surf}}^4 = T_{\text{eq}}^4 (1 + \frac{3}{4} \tau(z=0))$ $T_{\text{surf}} \neq T(z=0)$!
 - $T_{\text{surf}} > T_{\text{eq}}$: greenhouse effect, *importance increasing with $\tau_{(z=0)}$*
 - Temperature discontinuity: $T_{\text{surf}} > T(0 \text{ km}) \Leftrightarrow$ infinite temperature gradient !
 - Unstable \rightarrow onset of convection \rightarrow addition of an upward convective flux, reducing the near-surface temperature gradient to a near-adiabatic lapse rate. *Radiative equilibrium cannot hold down to surface.*

		$T_{\text{eff}} (\text{K})$	$T_{\text{surf}} (\text{K})$	Constituants responsables
• Order of magnitude of atmospheric opacity: $\tau = 134, 0.84, 0.13$ for Venus, Earth, Mars	Vénus	231	733	CO_2 , nuages $\text{H}_2\text{SO}_4, \text{H}_2\text{O}$
	Terre	255	288	$\text{H}_2\text{O}, \text{CO}_2, \text{CH}_4, \text{O}_3$
	Mars	210	215	CO_2
	Titan *	82	94	$\text{N}_2\text{-CH}_4\text{-H}_2$

The energy budget of the Earth



- In the lack of any atmospheric gases, the Earth surface would absorb 235 W m^{-2} ($342 - 107$)
- In reality, it absorbs 168 W m^{-2} ($235 - 67$) of solar radiation, and 324 W m^{-2} from the atmosphere, i.e 492 W m^{-2} .
- Out of those 492 W m^{-2} , 390 W m^{-2} are being used to heat the surface, and the rest is used for evaporation (78 W m^{-2}) and convection (24 W m^{-2}).
- The difference: $390 - 235 = 155 \text{ W m}^{-2}$ (« natural » greenhouse) causes the Earth to be at 288 K instead of 255 K
- **The natural greenhouse effect permits to have liquid water on surface**

Atmospheric absorption of solar and thermal radiation

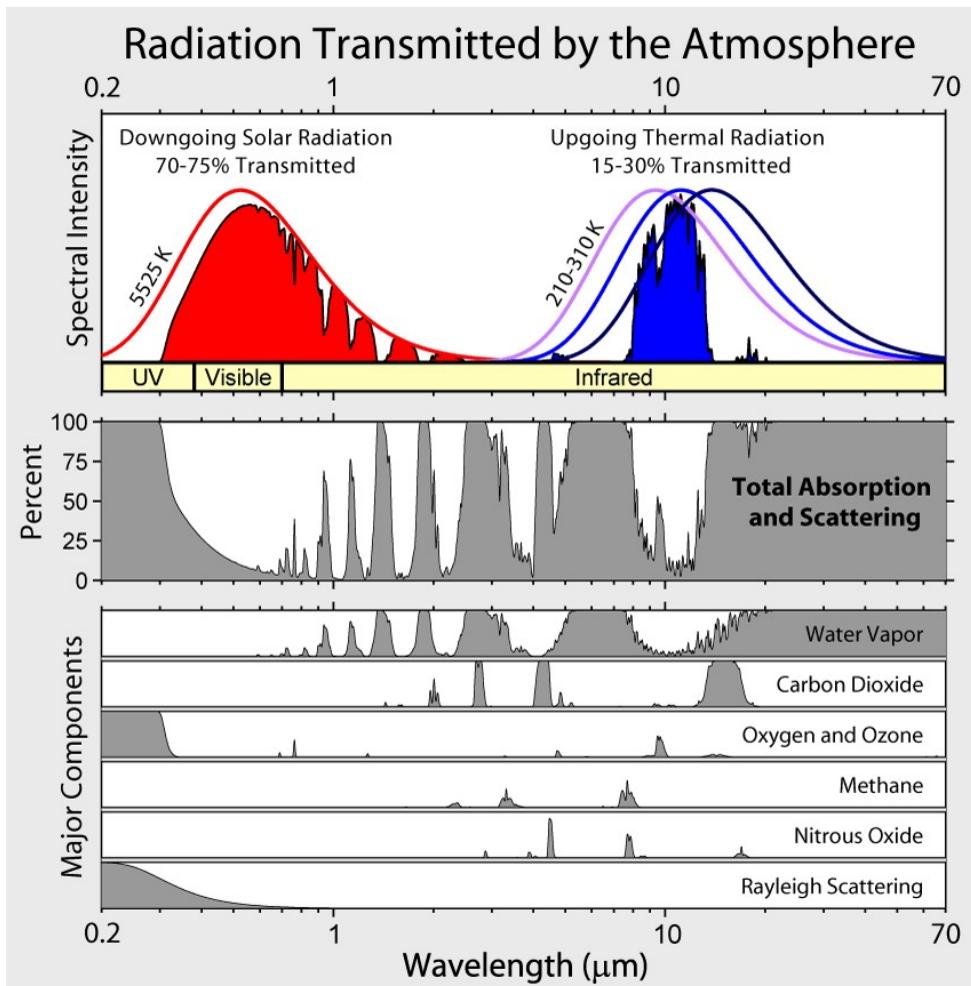


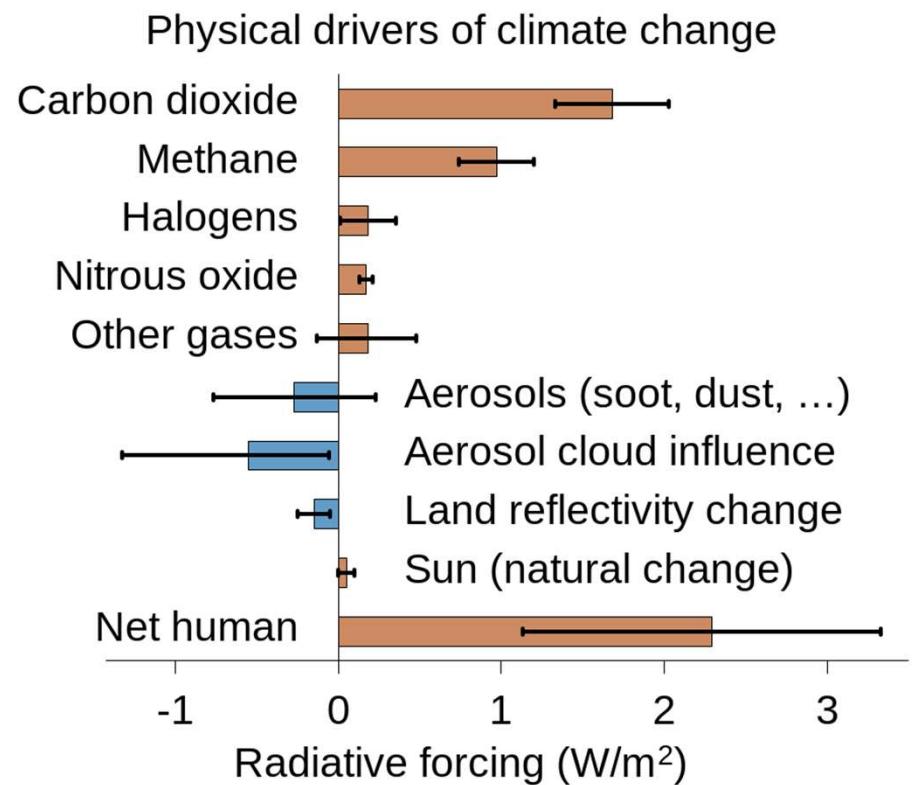
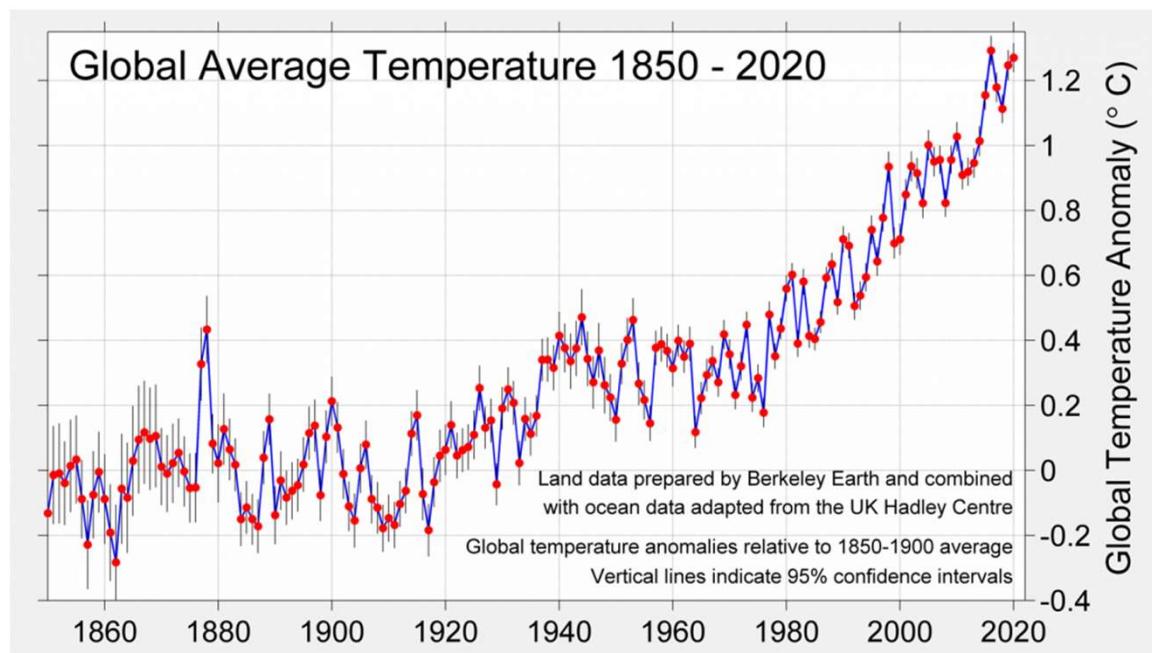
TABLE 3. Clear and cloudy sky radiative forcing (W m^{-2}) and the contribution of individual absorbers to this total. Cloudy sky results are in parentheses.

Gas	Individual contribution	Combined with overlap effects	Percent contribution clear sky
H_2O	71 (49)	75 (51)	60
CO_2	29 (22)	32 (24)	26
Overlap $\text{H}_2\text{O}-\text{CO}_2$	7 (4)		
O_3	8 (7)	10 (7)	8
Overlap with O_3	2		
$\text{CH}_4 + \text{N}_2\text{O} + \text{ovlp}$	8 (4)	8 (4)	6
Total	125 (86)	125 (86)	100

Clouds = 30 W m^{-2}

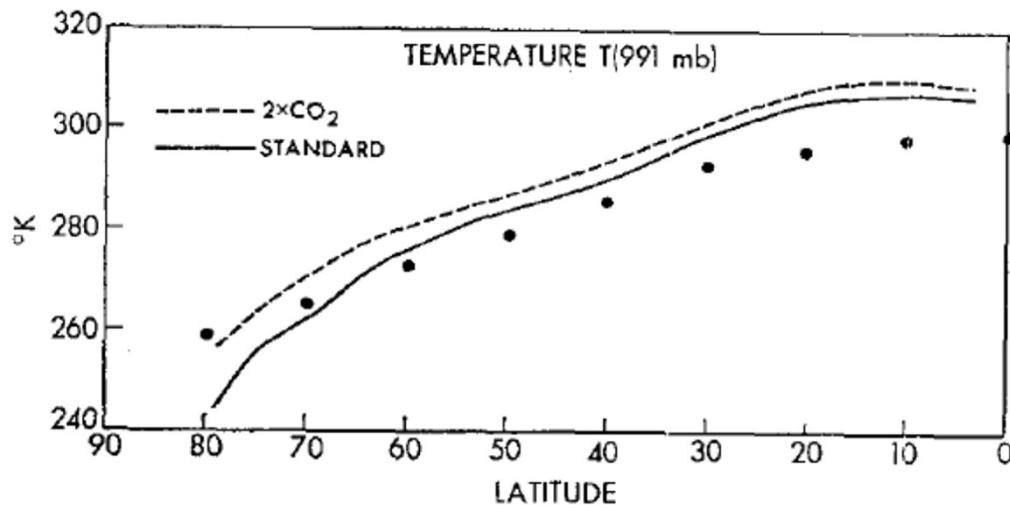
Total = 155 W m^{-2}

Global warming on Earth: effect and cause



Anthropocentric radiative forcing = 2.5 W m^{-2}
to be compared with 155 W m^{-2} natural greenhouse effect

Forty years lost



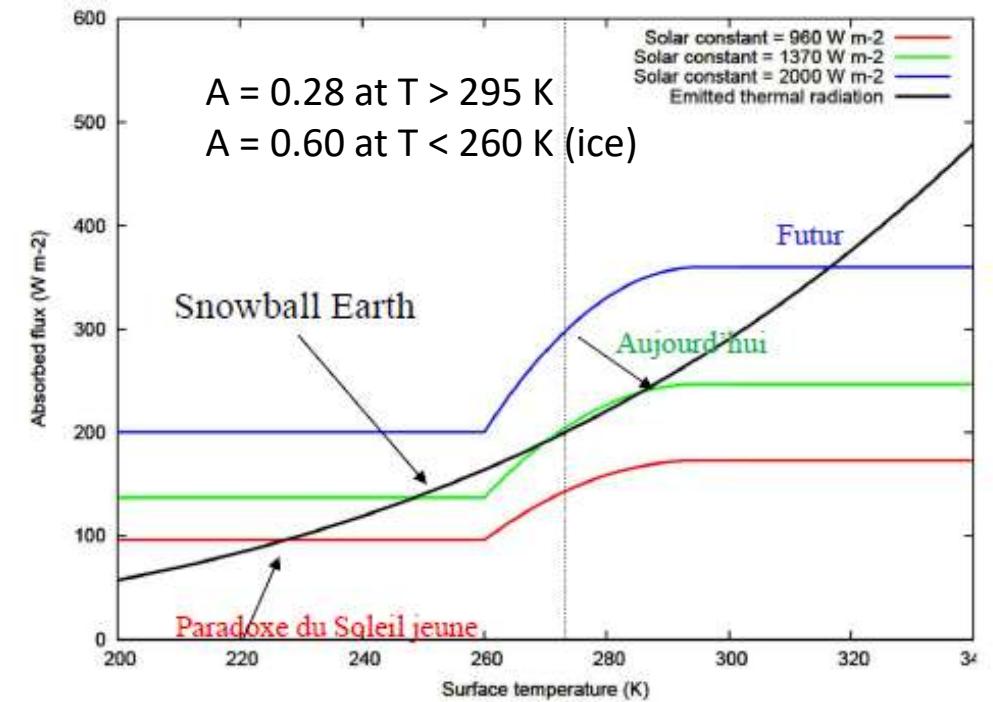
Manabe &
Wetherald (1975)

Charney report
(1979)

We conclude that the predictions of CO₂-induced climate changes made with the various models examined are basically consistent and mutually supporting. The differences in model results are relatively small and may be accounted for by differences in model characteristics and simplifying assumptions. Of course, we can never be sure that some badly estimated or totally overlooked effect may not vitiate our conclusions. We can only say that we have not been able to find such effects. *If* the CO₂ concentration of the atmosphere is indeed doubled and remains so long enough for the atmosphere and the intermediate layers of the ocean to attain approximate thermal equilibrium, our best estimate is that changes in global temperature of the order of 3°C will occur and that these will be accompanied by significant changes in regional climatic patterns.

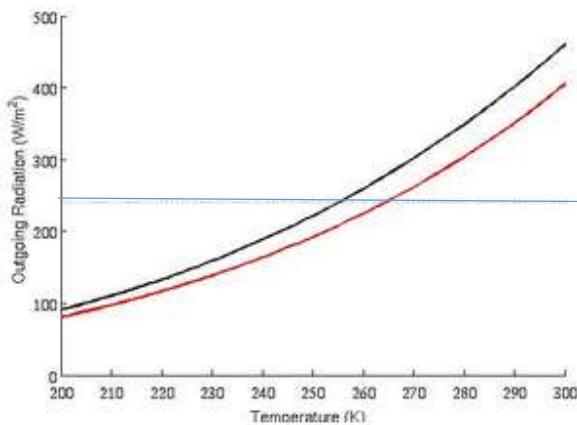
Greenhouse effect and (natural) planetary evolution (1)

- Albedo-temperature couplings
 - If surface albedo depends on temperature, possibility of several equilibrium states !
 - Snowball Earth (2.2 and 0.63 Gyr ago). How did it enter/get out of such glaciation states?)
- Evolution of solar luminosity
 - At 3.8 Gyr ago, Sun was fainter than today by about 25 %. With current atmosphere, the surface temperature would be ~ 230 K instead of 288 K.
 - Yet, there is geological evidence that there was liquid water, i.e. that $T > 273$ K. « Faint young Sun paradox »
 - Solution: higher CO₂ concentrations in the early Earth atmosphere

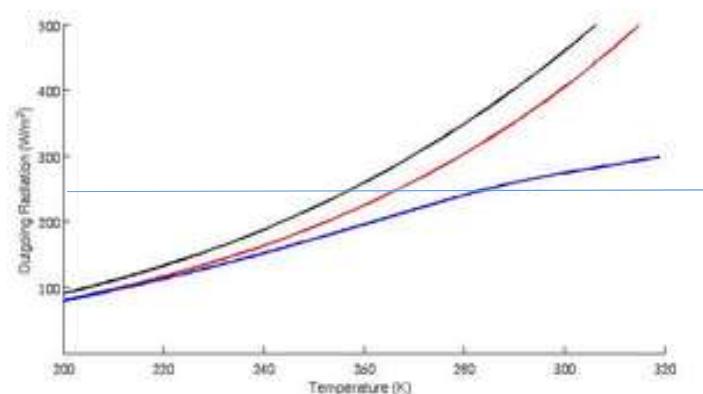


Greenhouse effect and (natural) planetary evolution (2)

- Temperature-composition retroactions : why is Venus dry ?
 - Early Venus was probably humid like Earth: what happened to the water?



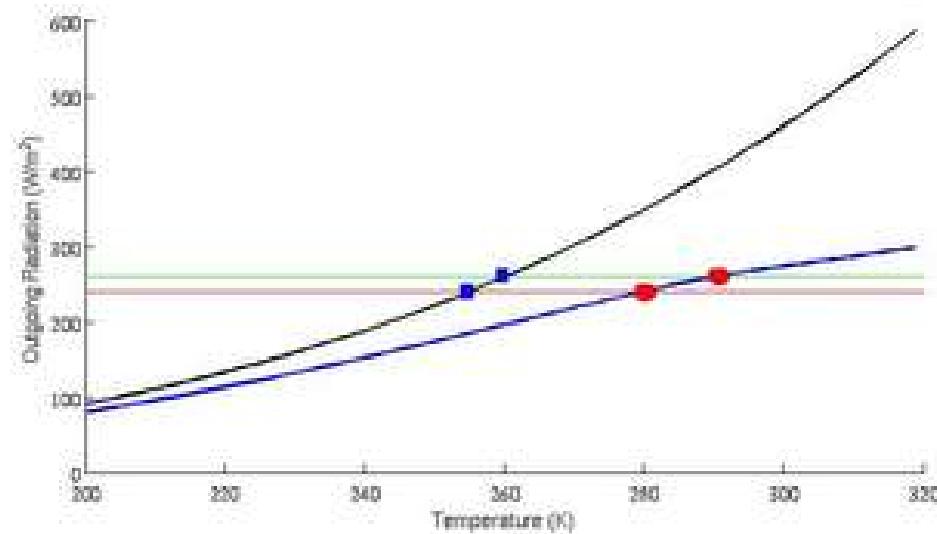
Fixed abundance of greenhouse gas



Greenhouse gas increases with temperature (e.g. H₂O)
→ stabilisation at a higher temperature

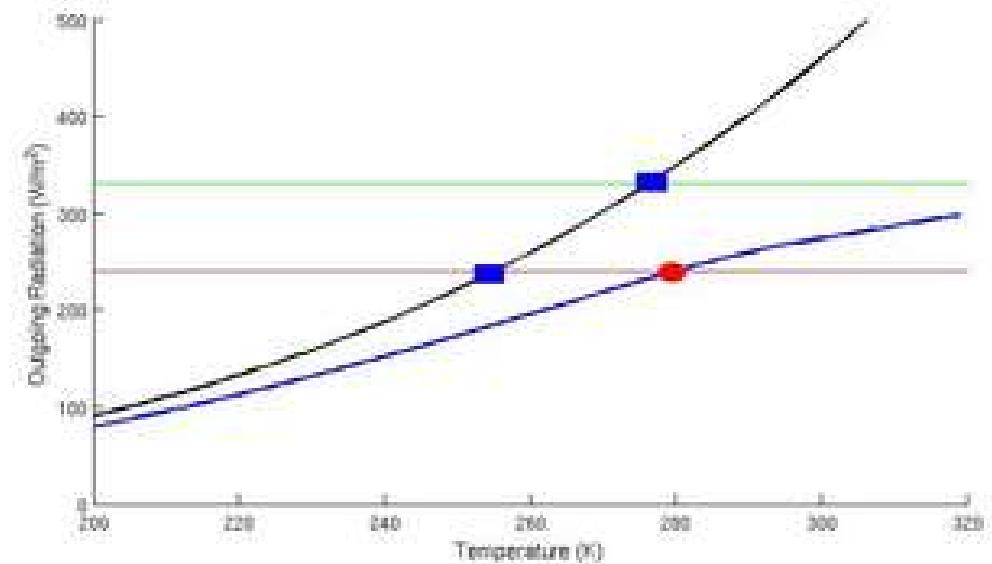
Greenhouse effect and (natural) planetary evolution (3)

- Runaway greenhouse effect



Small increase in solar radiation → stabilization

Has occurred on Venus, predicted to occur on Earth in 1 billion years



Large increase in solar radiation → no stabilization
temperature increase up to $>100^\circ\text{C}$: entire H₂O evaporation : runaway greenhouse effect

What happened on Venus after the loss of water?

On a planet with both liquid water and volcanism:

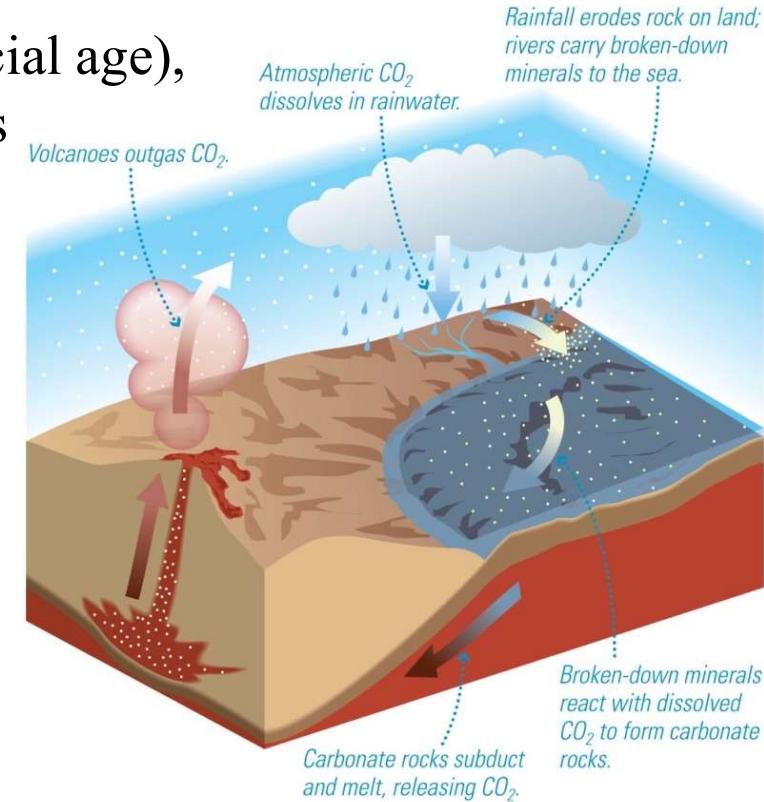
The carbon cycle (a.k.a. the geochemical carbonate cycle): a mechanism stabilizing the climate.

- If $\text{CO}_2 \uparrow$, $T \uparrow$, evaporation \uparrow , rain \uparrow , dissolution of CO_2 in rain \uparrow , hence $\text{CO}_2 \downarrow$
- If $\text{CO}_2 \downarrow$, greenhouse effect \downarrow , $T \downarrow$ (may be up to glacial age), evaporation \downarrow , loss of $\text{CO}_2 \downarrow$, and volcanism replenishes CO_2

Time scale of cycle ~ 0.5 millions of years (***not 50 years...***)

Requires : H_2O and volcanism

Venus: after the loss of H_2O , the carbon cycle was broken and CO_2 accumulated in atmosphere: 90 bar, 730 K



Vertical thermal structure: summary

- Troposphere **convection +radiation**

- Transport of heat flux from the surface (telluric planets) or from interior (giant planets) and absorption by the atmosphere. Non-zero opacity for thermal flux,
 $dT/dz < 0$.

tropopause

- Stratosphere **radiation**

- Direct absorption of solar flux (O_3 on Earth, CH_4 on giant planets, aerosols on Titan)
 - $dT/dz > 0$: the change of sign of dT/dz (« temperature inversion ») that defines the tropopause occurs where the absorption of solar flux dominates over the absorption of thermal radiation.
 - Does not occur on Venus, Mars, because CO_2 absorbs mostly in thermal range.

stratopause

→ Maximum of solar heating

- Mesosphere **radiation**

- Decrease of solar heating
 $dT/dz < 0$

mesopause

→ End of transport par radiation

- Thermosphere **conduction**

- Heat source at high altitude (absorption of solar UV)
 - Transport only by conduction (relatively inefficient) → $dT/dz \gg 0$

