Planetary Atmospheres

Structure

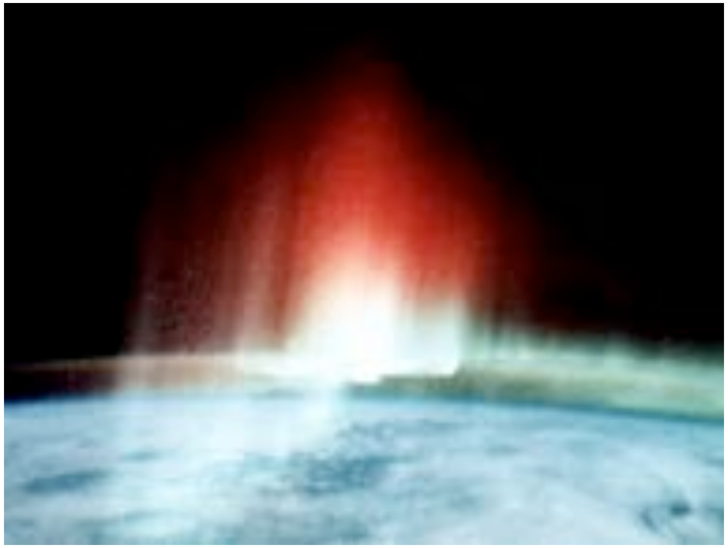
Composition

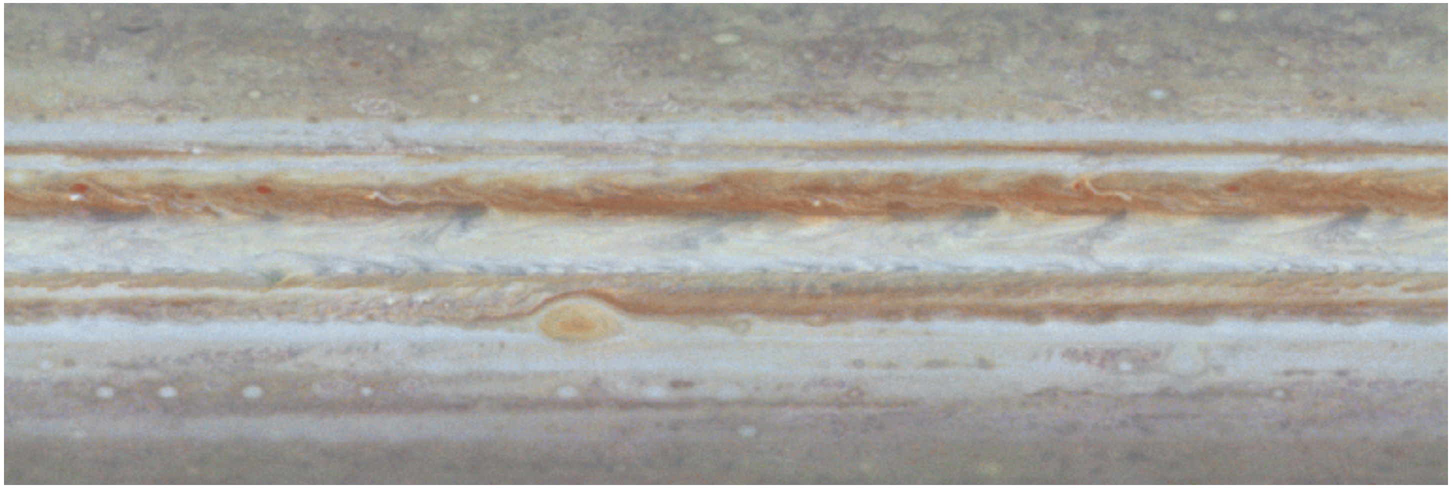
Clouds

Photochemistry

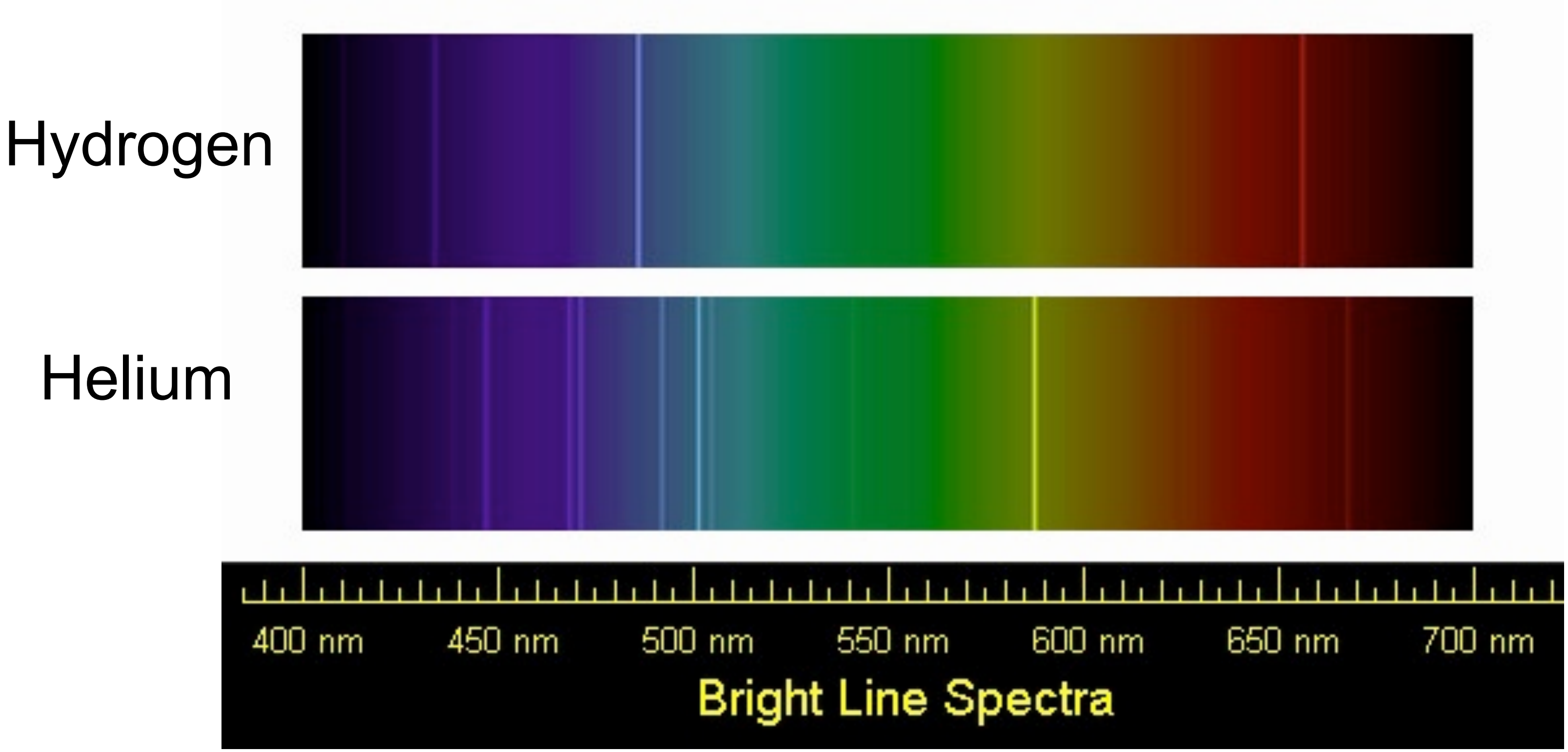
Meteorology

Atmospheric Escape





Spectra



Each element/molecule has its own spectral ‘fingerprint’ that can be observed in either emission or absorption depending on its temperature relative to the light source.

Cooler ⇒ Then wavelengths will be absorbed and appear dark in the spectrum.

Spectra: Sources

In observing spectral emission/absorption features in a planetary atmosphere, one must consider the primary sources of the continuum spectra.

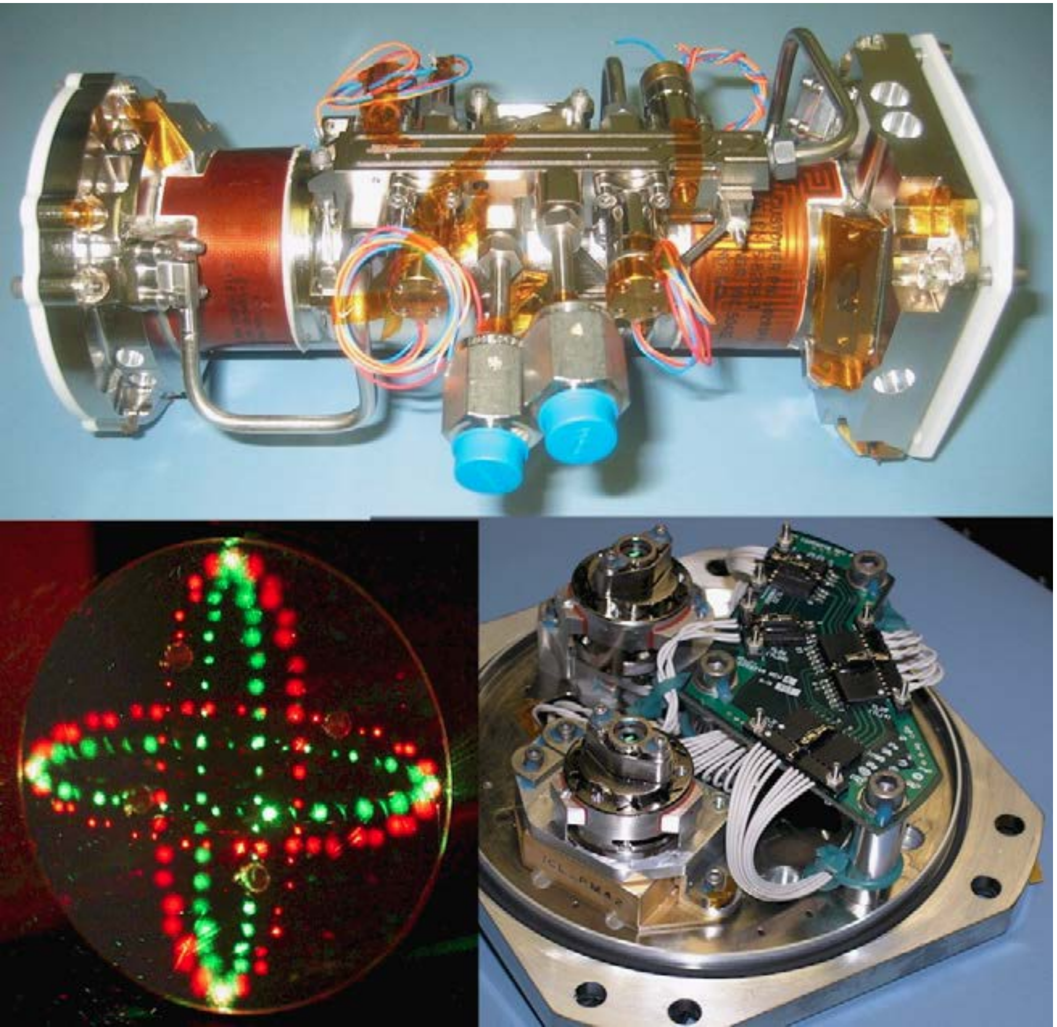
Reflected sunlight:
Generally in the UV, visible and near-infrared wavelengths

Example: Uranus and Neptune appear green/blue due to the presence of methane in their atmospheres. Methane absorbs the red part of the visible spectrum, causing mostly green/blue light to be reflected.

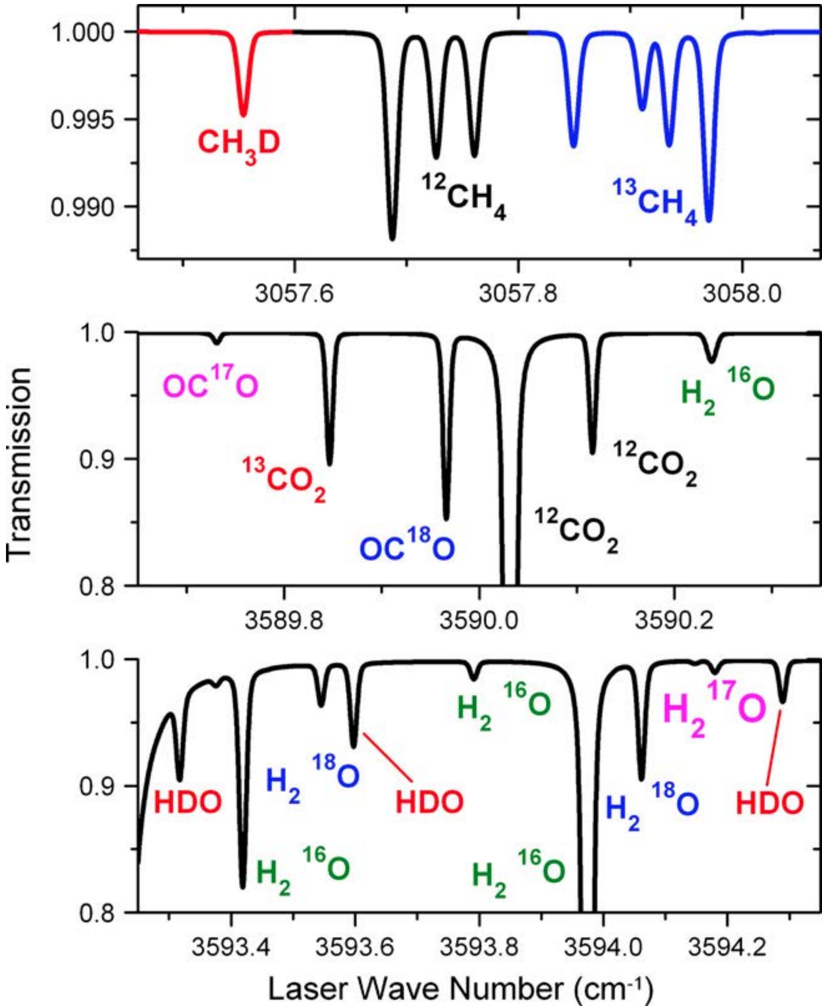
Thermal radiation:
From the ‘surface’ or deeper atmospheric layers of the planet, generally peaks in the infrared and radio wavelengths due to the temperature of the ‘surface’ generating a black body radiation curve.

Ground-based spectra: Fraunhofer + Planetary + Telluric

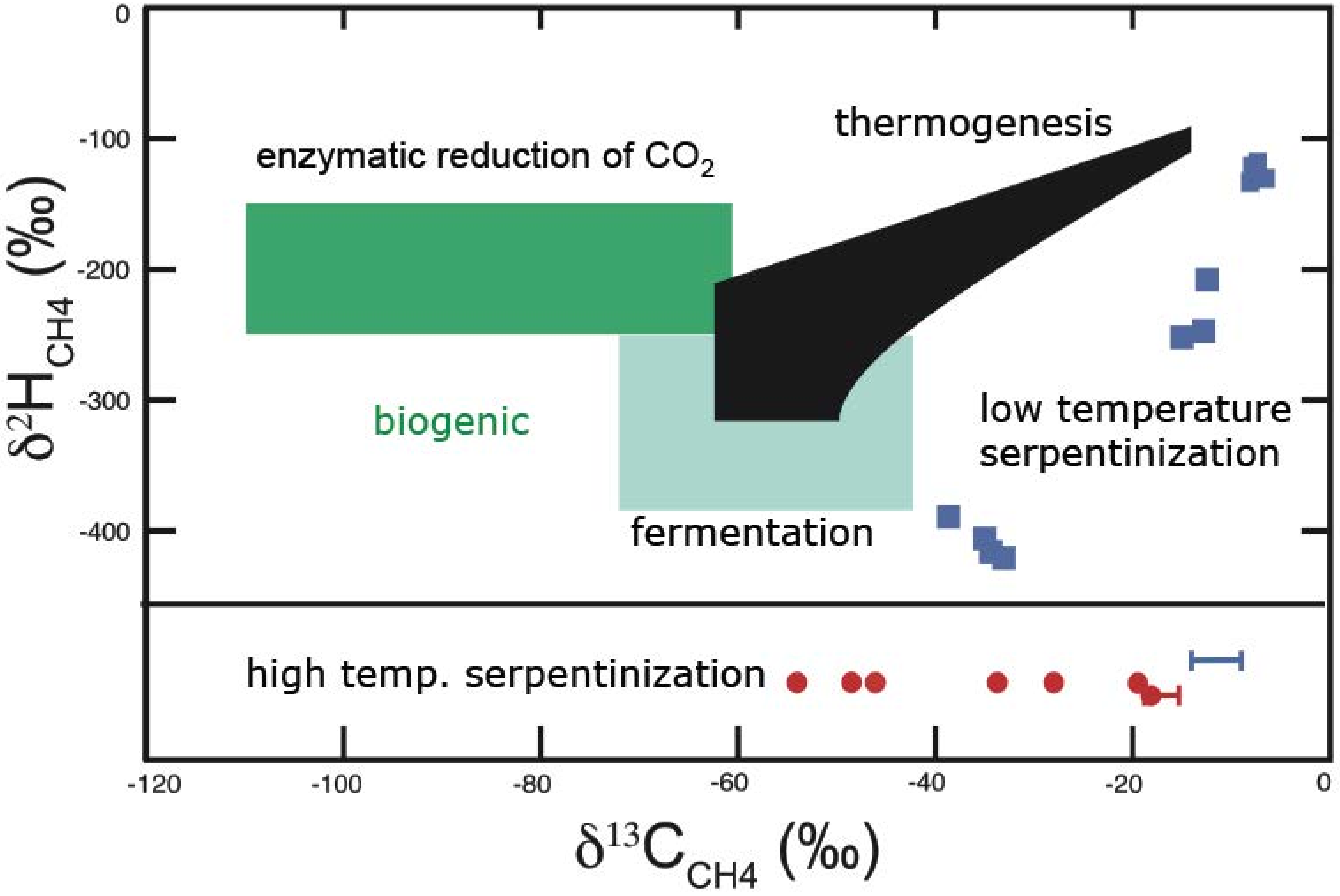
…or just take your own light source and instrument to the planet!



Webster & Mahaffy (2011)



Distinguishing CH_4 sources via isotopes



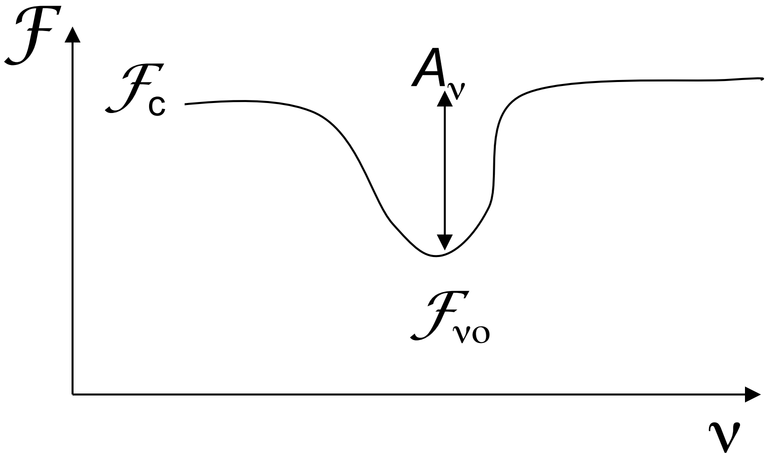
Atreya et al. (Methane Workshop)

Spectra: Line Profiles

Spectral lines have a finite shape that tells about various properties in the planetary atmosphere we are observing.

Note that the absorption features are often not 'blacked out' lines in the spectrum, but decreases in intensity with a measurable depth and width.

Absorption Depth:



$$A_\nu \equiv \frac{\mathcal{F}_c - \mathcal{F}_{\nu_0}}{\mathcal{F}_c}$$

$$A_\nu = 1 - e^{-\tau_\nu}$$

Spectra: Line Profiles

Absorption Width: Equivalent Width

$$EW = \int_0^\infty A_\nu \, d\nu = \int_0^\infty \left(1 - e^{-\tau_\nu}\right) d\nu$$

The optical depth at the center of the line is determined by the extinction coefficient and the column density (N_c):

$$\tau_{\nu o} = \int_0^L N \alpha_{\nu o} \, dl = N_c \alpha_{\nu o}$$

Therefore, for $\tau_\nu << 1$:

$$EW \approx \int_0^\infty \tau_\nu \, d\nu = N_c \alpha_{\nu o} \int_0^\infty \Phi_\nu \, d\nu$$

Where the line shape:

$$\Phi_\nu \equiv \frac{\alpha_\nu}{\alpha_{\nu_0}}$$

Spectra: Line Profiles

Equivalent Width

$$EW \approx N_c \alpha_{\nu o} \int_0^\infty \Phi_\nu \, d\nu$$

Increases linearly with N_c while $\tau_\nu << 1$, but as τ_ν increases the line profile saturates, causing the EW to become proportional to $(N_c)^{1/2}$

If this behavior (the “curve of growth”) is understood, then the abundance of a gas can be determined from the observed EW.

Spectra: Line Profiles

What controls the line shape?

Lorentz Line Shape: Shape due to finite lifetime of excited states

Doppler Broadening: Due to relative motion along the line of sight. Can be used to infer atmospheric wind speeds as well as temperature based on the Maxwellian distribution function.

Pressure/Collisional Broadening: Due to collisions between molecules slightly perturbing the energy levels of electron states (I.e. photons with λ_{ul} +/- $\delta\lambda$ can cause excitation/de-excitation)

* Remember that τ_{ν} will determine what altitude you are 'probing' for a given wavelength (or frequency).

Compositions of Terrestrial Atmospheres

	Earth	Venus	Mars	Titan
Pressure	1 bar	92 bar	0.006 bar	1.5 bar
N_2	77%	3.5%	2.7%	98.4%
O_2	21%	-	-	-
H_2O	1%	0.01%	0.006%	-
Ar	0.93%	0.007%	1.6%	0.004%
CO_2	0.035%	96%	95%	~1ppb
CH_4	1.7ppm	-	?	1.6%
^{40}Ar	6.6×10^{16} kg	1.4×10^{16} kg	4.5×10^{14} kg	3.5×10^{14} kg
H/D	3000	63	1100	3600
$^{14}N/^{15}N$	272	273	170	183

Isotopes are useful for inferring outgassing and atmos. loss

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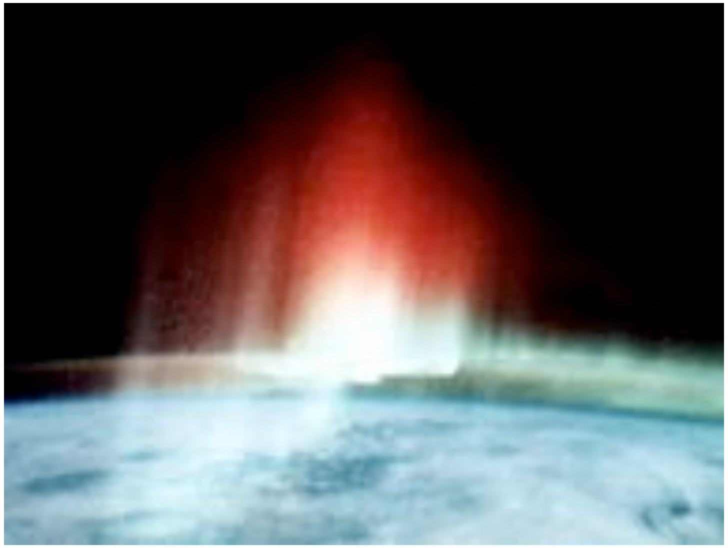
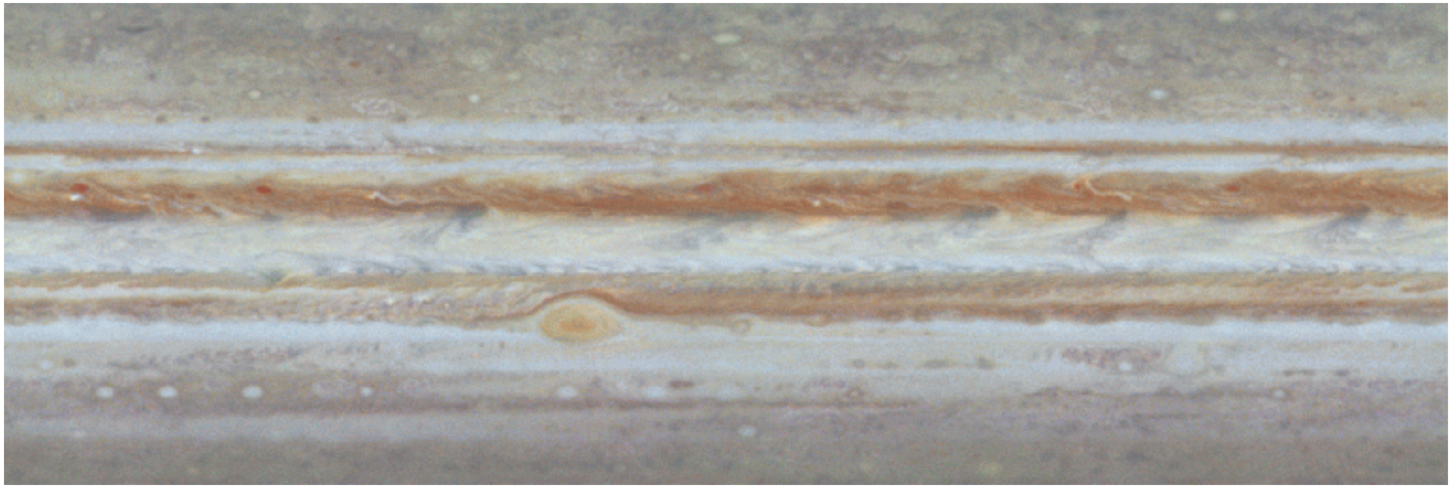
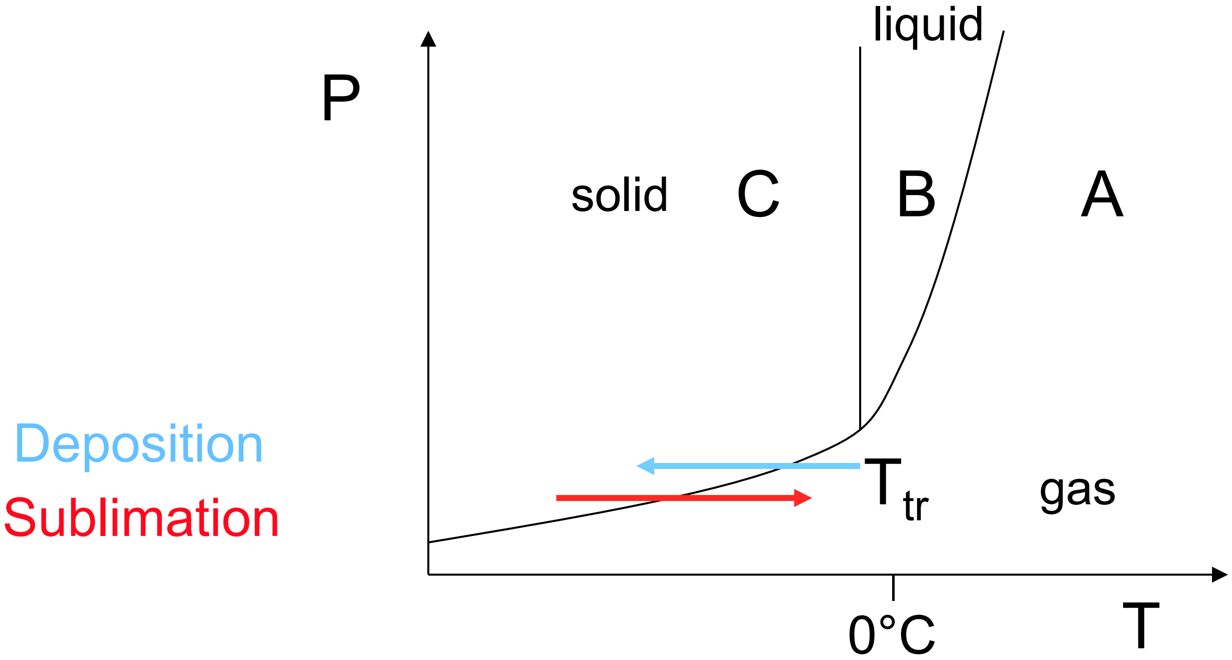




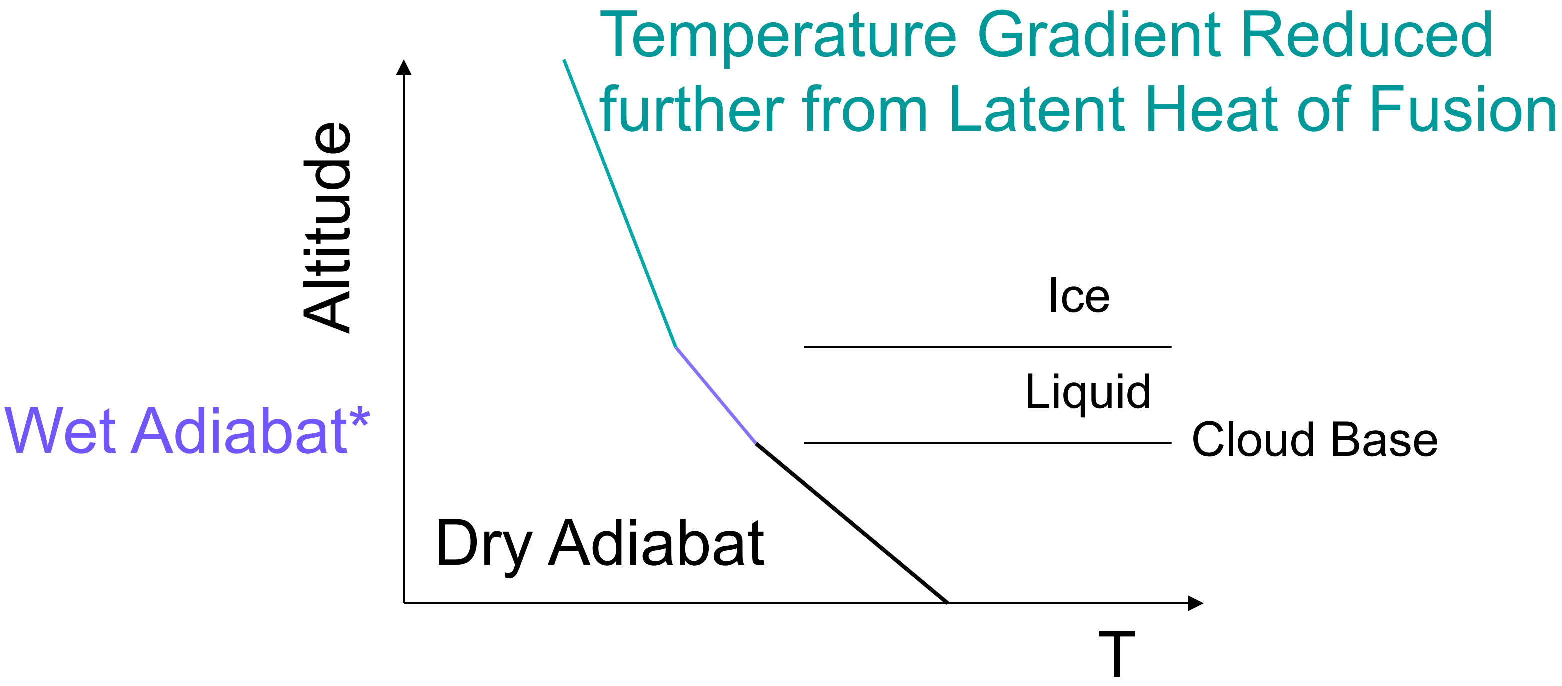
Cloud formation

Saturated Vapor Pressure: Maximum amount of water vapor partial pressure



Cloud formation

The phase change of water has an effect on the temperature structure of the atmosphere as well:



Temperature Gradient Reduced further from Latent Heat of Fusion

Wet Adiabat*

Dry Adiabat

*Thermal gradient reduced due to Latent Heat of Condensation

Clouds

Clouds modify the surface/atmospheric temperature of a planet in the following ways --

1. Decrease amount of incoming sunlight due to Albedo (reflectivity)
2. Heat immediate environment by absorbing solar radiation, thus changing the lapse rate
3. Block outgoing IR radiation can lead to greenhouse warming near the surface
4. Reduce the lapse rate via latent heat release during cloud formation

Cloud formation

Wet Adiabatic Lapse Rate:

$$P = C_L e^{-L_S/\left(R_{gas}T\right)}$$

Saturation Vapor Pressure

$$c_v dT = -PdV - L_S dw_S$$

L_s is *Specific* Latent Heat

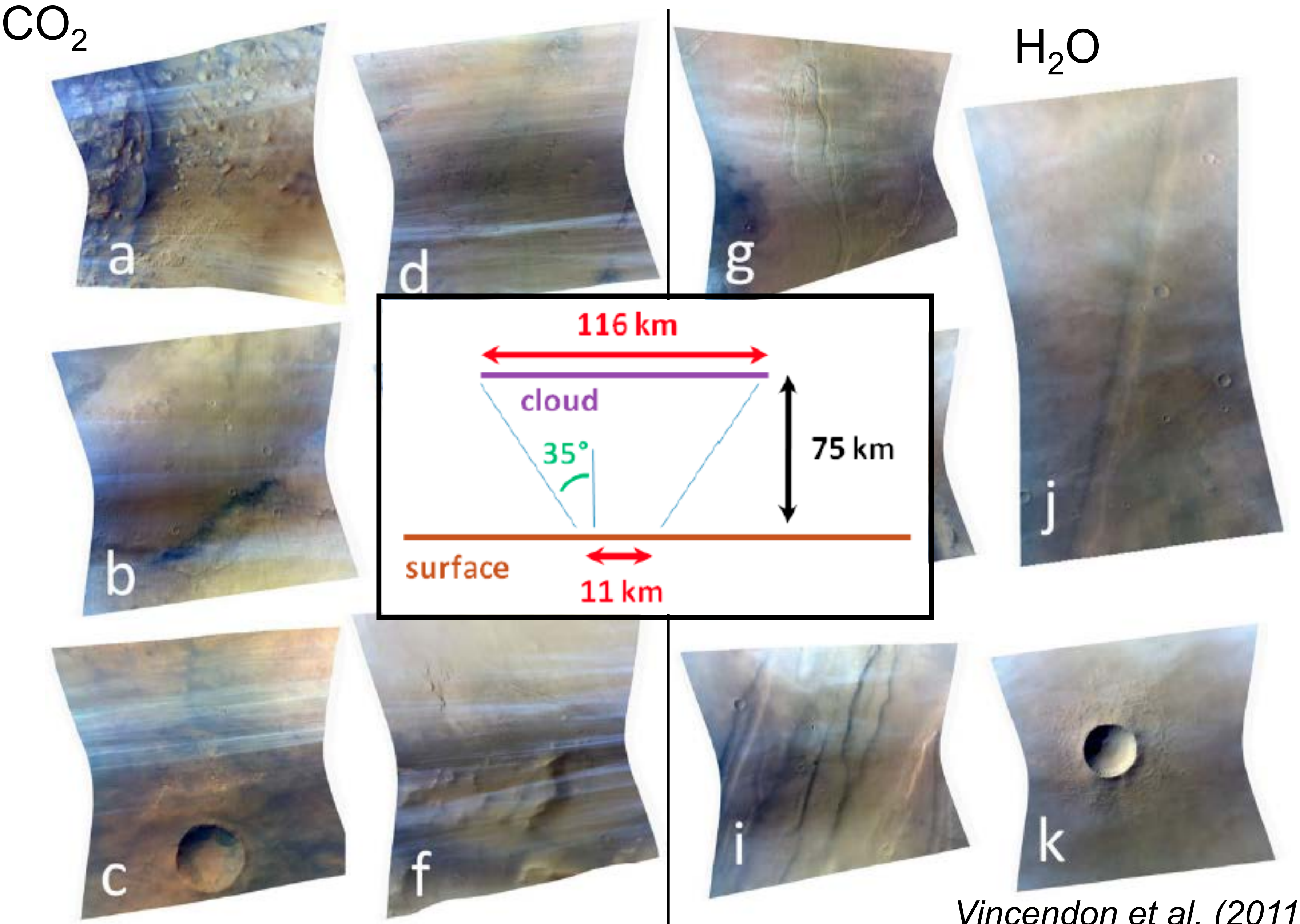
$$c_P dT = \frac{1}{\rho} dP - L_S dw_S$$

w$_s$ is the mass of water vapor that condenses out per gram of air

$$\frac{dT}{dz} = \frac{g_P}{c_P + L_S dw_S / dT}$$

= 5–6 K/km on Earth

Martian clouds



Vincendon et al. (2011)