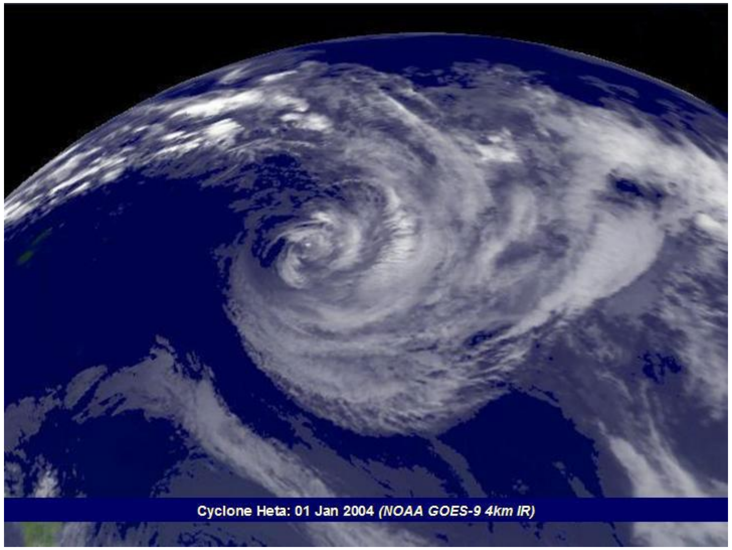
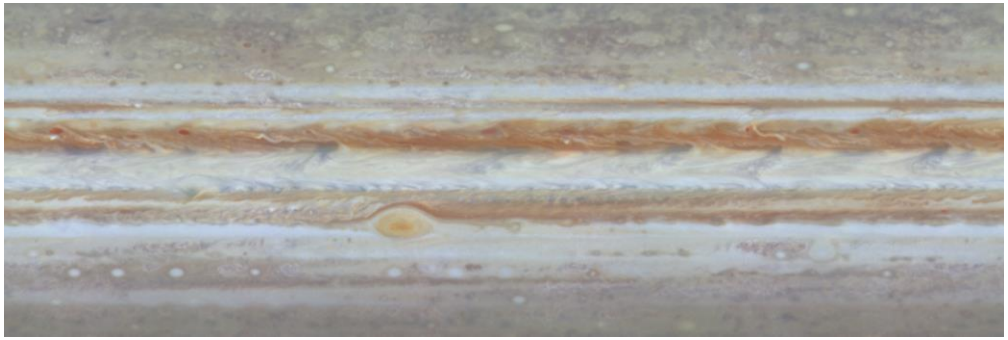
PLANETARY ATMOSPHERES

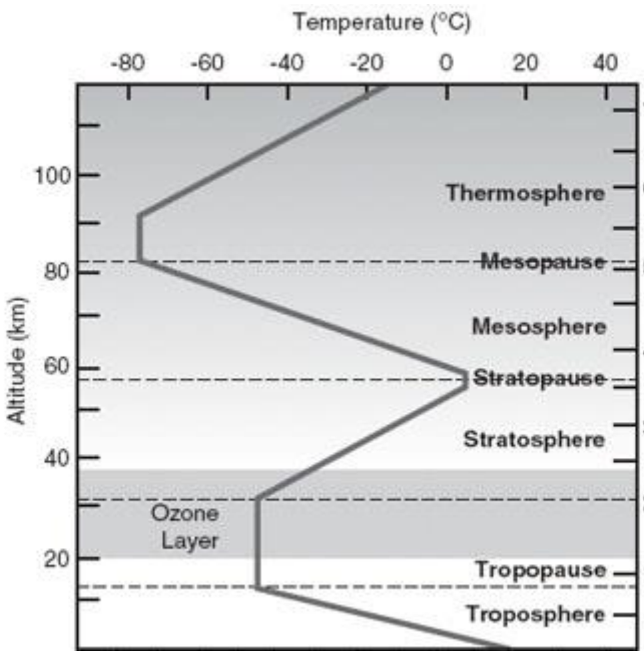




SUMMARY

1. VERTICAL MOVEMENTS

1.1. Large-scale vertical structure



For Earth

Photodissociation and ionisation of CO_2

CO_2 radiative cooling

Ozone absorption

Weather, clouds etc.

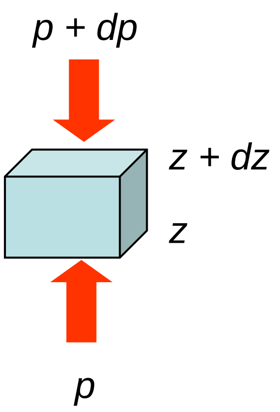
Layers ("…spheres"): temperature gradient with altitude
Regions ("…pauses"): fixed temperatures

Pressure/density of atmospheric gases decrease exponentially with altitude.

Atmosphere components can be in solid, liquid or vapour form (e.g. water on Earth)

1.2. Hydrostatic equation

$$\boxed{dp = -\rho_d \, g \, dz}$$



Equation for hydrostatic equilibrium

For dry air: R_d = 287 kJ kg^{-1} K^{-1}

$$pV = nRT = \frac{mass}{Molecular \;\; weight} RT$$

For a mixture of ideal gases, partial pressure is:

$$p_i V = nRT \Rightarrow p = \frac{m_i}{V} \frac{R}{M_i} T = \rho_i RT / M_i$$

$$p_{total} = \sum_i p_i = T \sum_i \rho_i R_i / M_i = \rho_d R_d T / M$$

Not examinable

Substituting for the hydrostatic equation and integrating:

$$\frac{dp}{p} = -\frac{Mg}{R_d T} dz \Rightarrow p(z) = p(0) e^{-\frac{z}{z_0}} \qquad z_0 = \frac{R_d T}{Mg}$$

Consistent with observed exponential dependence on altitude.

Troposphere: T_{mean} *= 250 K* $\Rightarrow z_0 \simeq$ *7.3 km for dry air*

1.3. Lapse rate - Adiabat

Lapse rate = minus rate of temperature change with altitude: $-\dfrac{\partial T}{\partial z}$

Derived from First Law of thermodynamics.
For adiabatic case, amount of extra heat writes:

$$\delta Q = c_V\, dT + p\, dV = 0$$

c_V: *specific heat at V constant*

As air volume rises, p decreases and it expands:

$$pV^{\gamma} = cst \qquad d\left(pV^{\gamma}\right) = V^{\gamma} dp + \gamma p V^{\gamma-1} dV = 0 \qquad \gamma = \frac{c_P}{c_V}$$

$$pdV = -\frac{V}{\gamma} dp$$

$$\frac{V}{\gamma} dp = c_V\, dT + p\, dV + \frac{V}{\gamma} dp = c_V dT \Rightarrow \quad V\, dp = c_P dT$$

$$\frac{dT}{dp} = \frac{R_d T}{c_P\, p}$$

Not examinable

The lapse rate is therefore:

$$\boxed{\Gamma_d = -\frac{\partial T}{\partial z} = -\frac{\partial T}{\partial p} \times \frac{\partial p}{\partial z} = \frac{R_d T}{c_P P} \times \frac{g\, p}{R_d T} = \frac{g}{c_P}}$$

For dry air, c_P = 1008 J kg^{-1} K^{-1} (e.g. from Kaye and Laby) $\Rightarrow \Gamma_d = 9.8 \times 10^{-3}$ K m^{-1}

Temperature decreases by 1ºC for each 100 m up.

z(T) is called adiabat

1.4. Saturated lapse rate – Vertical stability

Air usually contains some water vapour
If the air is not saturated, Γ_d equation valid with:

$$c_P := w_m \, c_{Pw} + (1 - w_m) c_{Pd}$$

(w_m = mass fraction of water vapour)

Rising moist air will cool, relative humidity will increase. At some altitude, air is saturated.
Some of the moisture will condense. Mass fraction of water vapour decreases,
releasing evaporation heat ΔH_V, warming the air: $\delta Q \neq 0$

The new (saturated) lapse rate becomes:

$$\boxed{\Gamma_{sat} \approx \Gamma_d + \frac{\Delta H_V}{c_P} \times \frac{\partial w_m}{\partial z}}$$

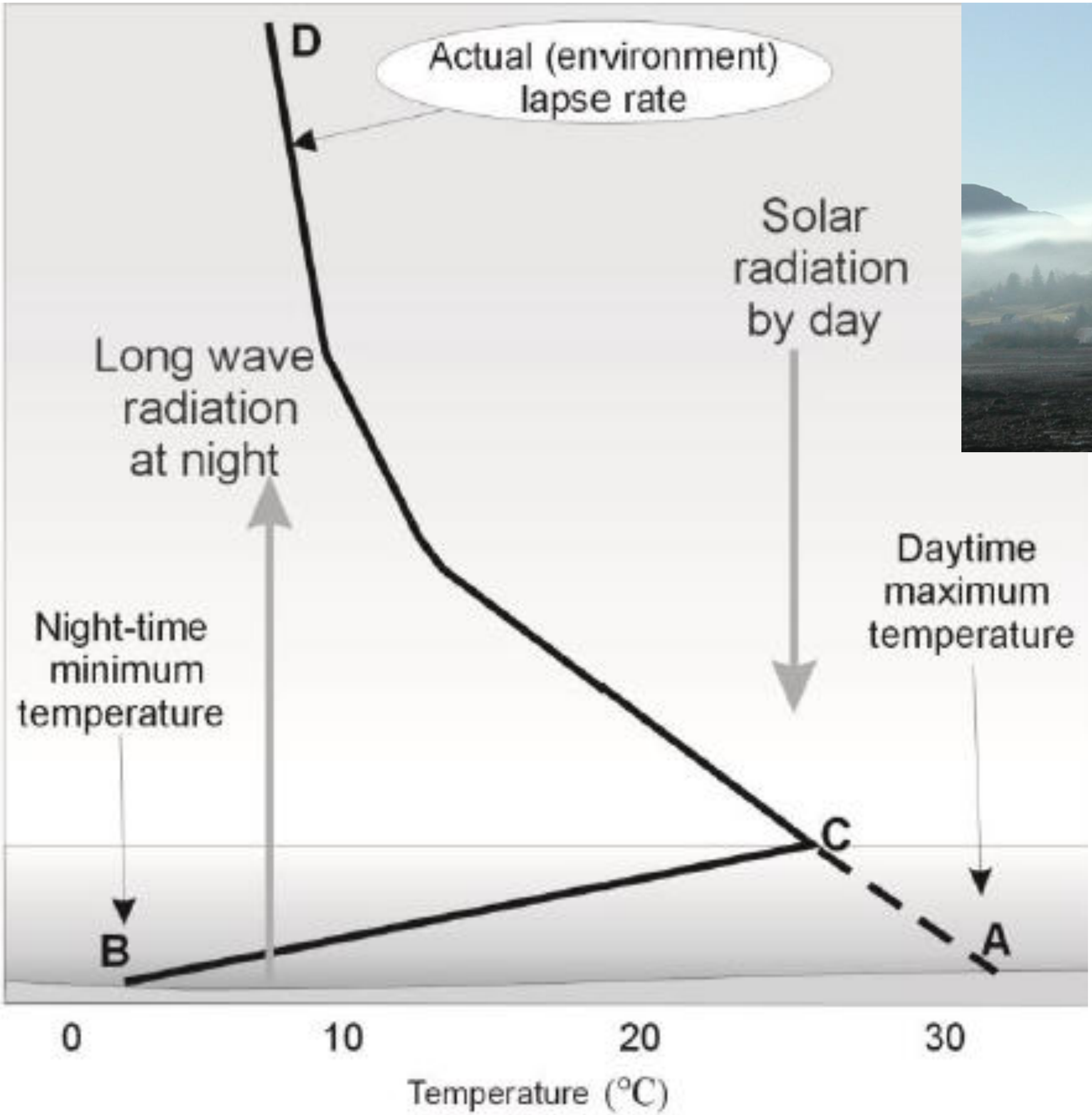
Lower than dry lapse rate: air cools more slowly as it rises, and not linearly.

Vertical stability is determined by difference of rising air lapse rate with environment:

$\Gamma_{env} > \Gamma_{air}$ unstable atmosphere (assists vertical motion: *convective instability*)

$\Gamma_{env} = \Gamma_{air}$ neutral atmosphere (neither assists nor resists vertical motion)

$\Gamma_{env} < \Gamma_{air}$ stable atmosphere (resists vertical motion)



Inversion layer

Air near the ground cools by conduction.

Up to 100's m high

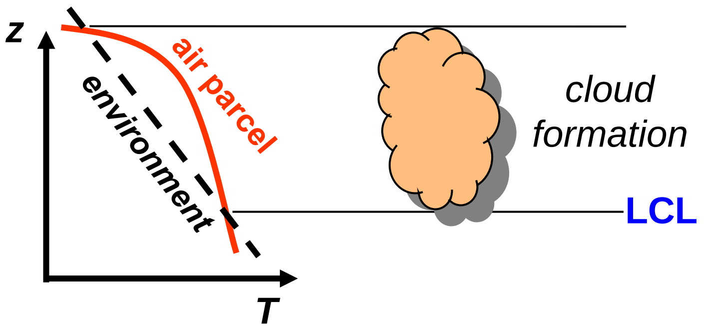
Often visible in early morning

Can trap aerosols (dust)

1.5. Cloud formation – Precipitation

In unstable atmosphere, water vapour will condense out of rising air, forming clouds, when local relative humidity reaches 100%.
This happens at the Lifting Condensation Level (LCL): cloud base.

Air parcels will follow saturated adiabat and expand, then cool …

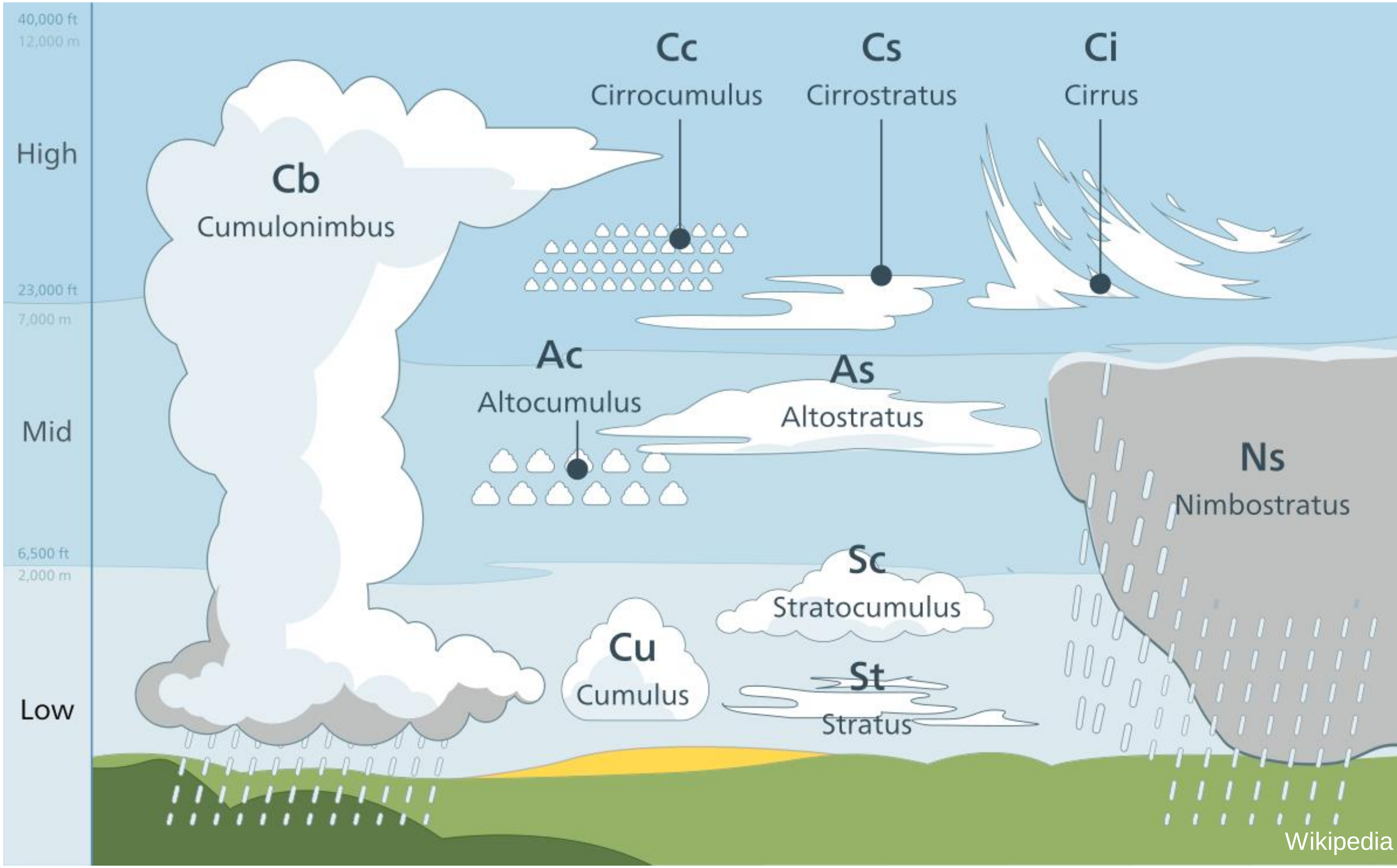


Adiabat for air parcel changes with z as it dries

Condensation: water droplets nucleate on aerosol particles (e.g. sea salt, smoke, dust)
Droplets will fall downwards to terminal velocity.
Beneath LCL, they might re-evaporate and return to cloud.

Higher cirrus clouds: T = -40ºC

Updraughts and temperature variations create different cloud types (e.g. cumulonimbus).



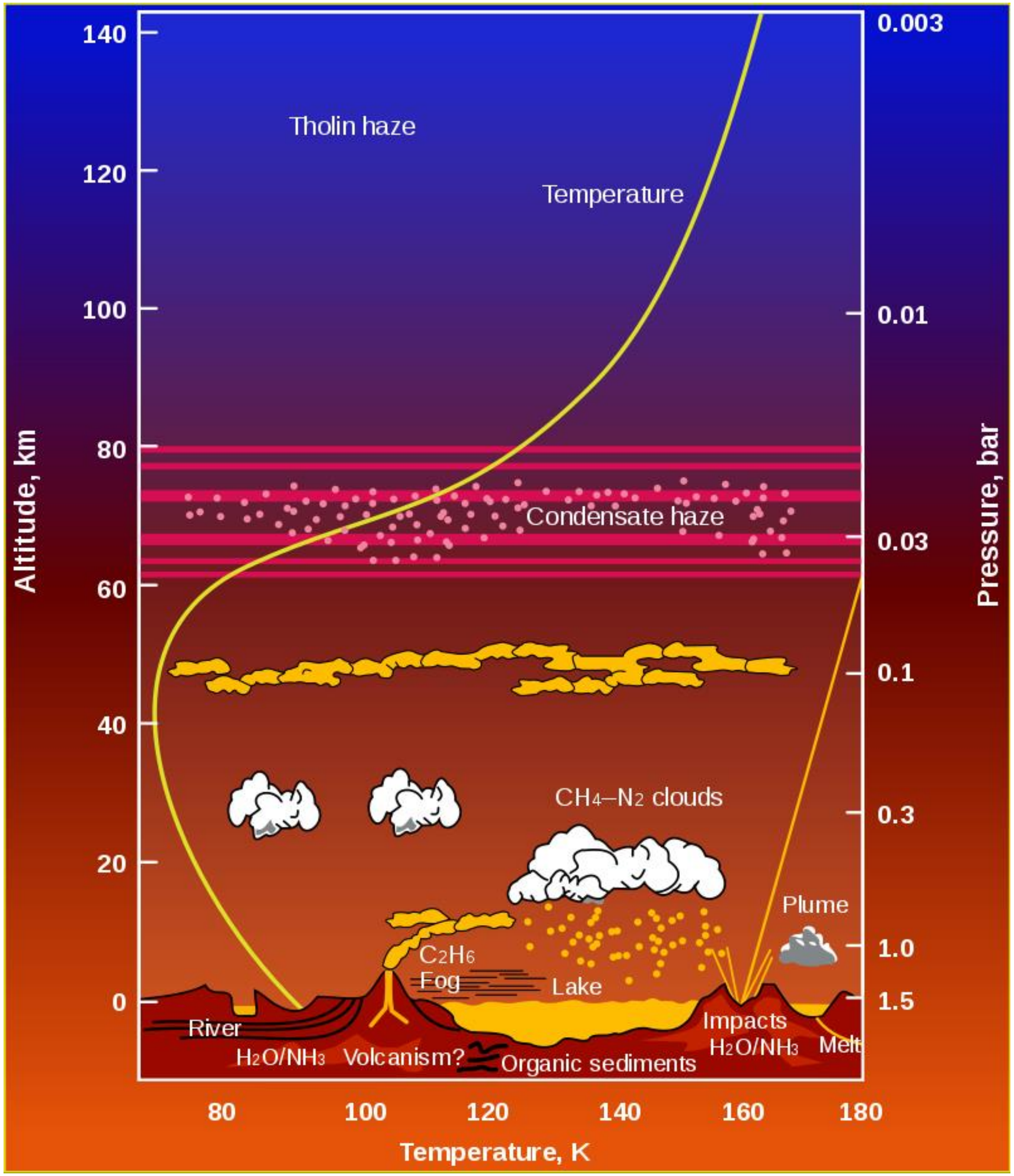
Different cloud types (on Earth) based on lapse rates and relative humidity

Atmospheres and clouds vary with the planets

Mars is famous for its regular dust storms, every 3 Mars years (5.5 Earth years), encircling the entire planet.

Mars Reconnaissance Orbiter (2010). *Credits: NASA/JPL-Caltech/MSSS*

Dust towers on Mars are concentrated clouds of dust that can rise up to 20 km above the surface.



NASA (http://www.jpl.nasa.gov/media/cassini-102504/visuals.html)

Clouds on Titan

Also rain, haze …

SUMMARY

1. Vertical Movements
   1.1. Large-scale vertical structure
   1.2. Hydrostatic equation
   1.3. Lapse rate – Adiabat
   1.4. Saturated lapse rate – Vertical stability
   1.5. Cloud formation – Precipitation

2. Horizontal Movements
   2.1. General circulation
   2.2. Driving forces
   2.3. Geostrophic flow
   2.4. Vertical wind shear
   2.5. Horizontal wind shear (weather fronts)

3. Remote Sensing of Planetary Atmospheres
   3.1. Identifying elements and volumes
   3.2. Reflected/scattered components
   3.3. Thermal component

4. Example: Jupiter
   4.1. General view
   4.2. Thermal emission
   4.3. Zones and belts
   4.4. The Great Red Spot + Red Junior
   4.5. And further away …

2. HORIZONTAL MOVEMENTS

2.1. General circulation

Horizontal movements created by pressure differences (i.e. *wind*)

Large scale circulation:

Intense solar heating near equator.

Warm air rises and cools, moves polewards and goes down again

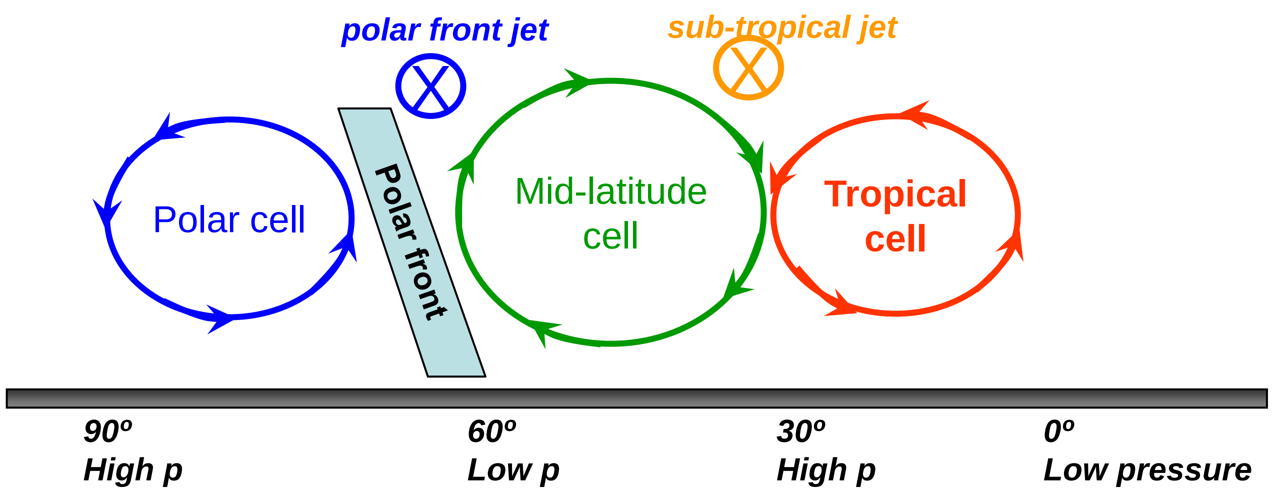
subtropical high pressure regions

Spreads out toward equator and poles

Air moving poleward will meet colder polar air

subpolar low pressure belt

Polar air mass rises and returns to the pole



General model: needs to account for rotation of Earth.

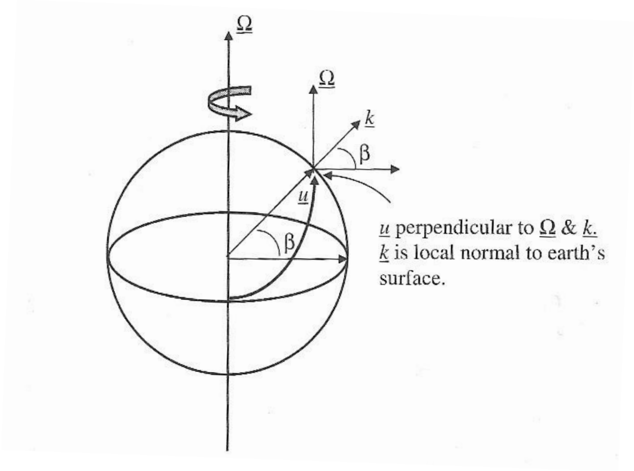
2.2. Driving forces

A volume dV of air, with mass ρdV and velocity $\mathbf{u}$, will feel forces:

$$\rho\, dV \; \frac{d\boldsymbol{u}}{dt} = \boldsymbol{F}_{\boldsymbol{Pressure}} + \boldsymbol{F}_{\boldsymbol{Coriolis}}$$

Viscous drag forces (from neighbouring layers of air at different velocities) are negligible above 500 m (because of small vertical gradient of wind velocity)

Coriolis force $\boldsymbol{F_C}$ arises from rotation of the Earth:



$$\mathbf{F}_\mathrm{C} = -2m(\mathbf{\Omega}\times\mathbf{u})$$

Ω= angular velocity of rotating reference frame

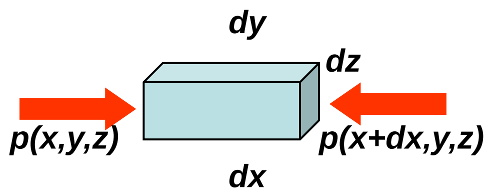
Its vertical component depends on latitude β

$$\mathbf{\Omega}_{vertical} = \sin\beta\,\Omega\mathbf{k}$$

$$\boxed{\mathbf{F}_\mathrm{C} = -2\rho\, dV \sin\beta\; \Omega(\mathbf{k}\times\mathbf{u})}$$

Pressure gradient along x direction:

$$F_P = p(x,y,z)\,dy\,dz - p(x+dx,y,z)\,dy\,dz$$
$$= p(x,y,z)\,dy\,dz - \left(p(x,y,z) \frac{\partial p}{\partial x} dx \right) dy\,dz$$
$$= -\frac{\partial p}{\partial x} dV$$

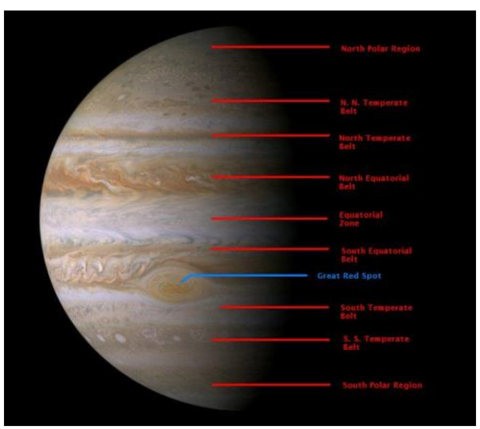


In 3 dimensions: $\underline{F_P} = -\nabla \underline{p}\, dV$

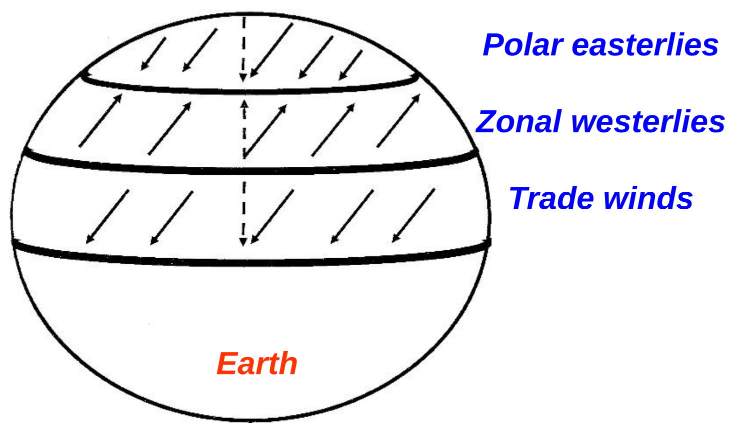
Adding the Coriolis force: $\boxed{\dfrac{d\underline{u}}{dt} \rho\, dV = -\nabla \underline{p}\, dV - 2 \sin\beta\, \Omega (\underline{k} \times \underline{u}) \rho\, dV}$

(equation of motion for the elemental volume)

Planet rotation deflects the winds: cells become belts



Jupiter



2.3. Geostrophic flow

Horizontal motion described by:

$$\boxed{\frac{d\underline{u}}{dt}\rho\, dV = -\nabla \underline{p}\, dV - 2\sin\beta\,\Omega\left(\underline{k}\times\underline{u}\right)\rho\, dV}$$

Neglects friction (between layers) and gravity. Flow is said to be *geostrophic*

Under ordinary conditions, wind velocity varies slowly over time.

$$\frac{d\underline{u}}{dt}\approx 0 \Rightarrow 2\sin\beta\,\Omega\left(\underline{k}\times\underline{u_g}\right)\rho = -\nabla \underline{p}$$

$\underline{u}_g$ is the geostrophic wind velocity.

Balance between Coriolis and pressure gradient forces

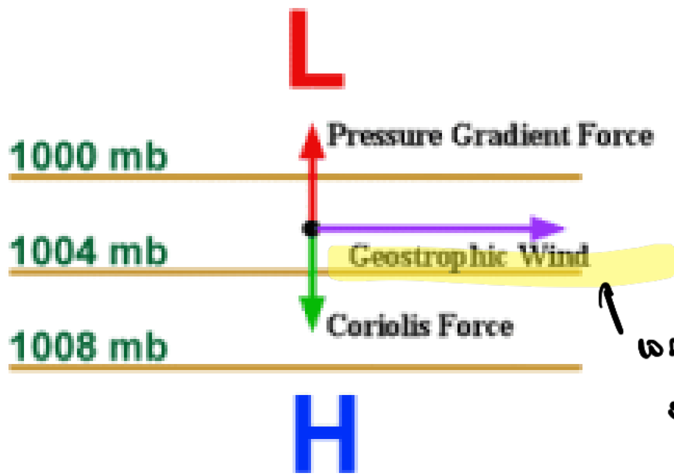
As $\underline{k}$ is vertical (zenith), wind direction must be parallel to isobars (in 3-D)

Geostrophic wind speed varies with latitude:

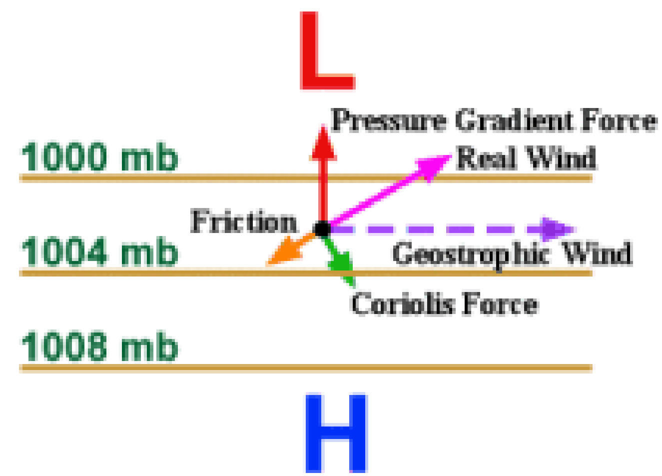
$$\boxed{u_g = \frac{|\nabla p|}{2\rho\sin\beta\,\Omega} = \frac{|\nabla p|}{f\,\rho}}$$

f is the Coriolis parameter

Wind speed increases with pressure gradient (not surprising) and with latitude



coordinate
of the lines



Geostrophic wind **Ageostrophic wind**

From: http://ww2010.atmos.uiuc.edu/(Gh)/guides/mtr/fw/fric.rxml

Geostrophic wind direction must be parallel to isobars

This usually occurs at higher latitude (1-2 km), where surface features affect the wind less, because they create friction:

- A calm ocean is rather smooth, not affecting the wind much;
- Conversely, hills and forests can slow down surface winds and/or change their direction

Below the friction (boundary) layer, wind is not geostrophic.

nice approx but
does not accurately

represent "cplx" atmosphere

2.4. Vertical wind shear

Wind shear = difference in wind speed between neighbouring layers → *shear vector*

Controlled by horizontal temperature gradient (a.k.a. "thermal wind")

We just saw:

$$\frac{d\underline{u}}{dt} \approx 0 \Rightarrow 2\sin\beta\Omega\left(\underline{k}\times\underline{u_g}\right)\rho = -\nabla \underline{p}$$

and (much earlier):

$$pV = nRT = \frac{mass}{Molecular\ weight}RT$$

For a unit mass:

$$\frac{1}{V} = \rho \quad \Rightarrow \quad \underline{k}\times\underline{u_g} = -\frac{1}{f\,\rho}\nabla p = -\frac{R_{air}}{f\,\rho}\left(T\nabla\rho + \rho\nabla T\right)$$

Dividing by *T* and using the fact that: $\frac{1}{y}\frac{\partial y}{\partial x} = \frac{\partial}{\partial x}(\ln y)$

Not examinable

$$\frac{\underline{k}\times\underline{u_g}}{T} = -\frac{R_{air}}{f}\left(\frac{\nabla\rho}{\rho} + \frac{\nabla T}{T}\right) = -\frac{R_{air}}{f}\left(\nabla\ln\rho + \nabla\ln T\right) \Leftrightarrow \boxed{\frac{\underline{k}\times\underline{u_g}}{T} = -\frac{R_{air}}{f}\nabla\ln p}$$

To find the variation of wind velocity with altitude, we differentiate with respect to *z*

Differentiating in *z*:

$$\frac{\partial}{\partial z}\left(\frac{\underline{k}\times\underline{u_g}}{T}\right)=-\frac{R_{air}}{f}\frac{\partial}{\partial z}(\nabla\ln p)=-\frac{R_{air}}{f}\nabla\frac{\partial}{\partial z}\ln p$$

Using the hydrostatic equation as a starting point, we get:

$$\frac{1}{\rho}\frac{\partial p}{\partial z}=\frac{R_{air}}{\rho}\frac{\partial}{\partial z}(\rho T)=\frac{R_{air}}{\rho}\left(\rho\frac{\partial T}{\partial z}+T\frac{\partial\rho}{\partial z}\right)=-g$$

Dividing by *T* again:

$$-\frac{g}{T}=R_{air}\left(\frac{1}{T}\frac{\partial T}{\partial z}+\frac{1}{\rho}\frac{\partial\rho}{\partial z}\right)=R_{air}\left(\frac{\partial}{\partial z}\ln T+\frac{\partial}{\partial z}\ln\rho\right)=R_{air}\frac{\partial}{\partial z}\ln p$$

And substituting in original equation

Not examinable

$$\boxed{\frac{\partial}{\partial z}\left(\frac{\underline{k}\times\underline{u_g}}{T}\right)=\frac{g}{f}\nabla\frac{1}{T}=-\frac{g}{fT^2}\nabla T}$$

$\underline{k}$ is vertical : variation in geostrophic wind velocity depends on horizontal *T* gradient

This is the *baroclinic model* of an atmosphere.

Vertical distribution of wind shear varies with latitude and season

On Earth: two westerly main jets:

subtropical jet: at 40ºN in summer, 200 mbar altitude (~13 km) (blowing at 20 m/s)

polar front jet: near polar front, 500 mbar altitude (~11 km) (blowing at up to 50 m/s)

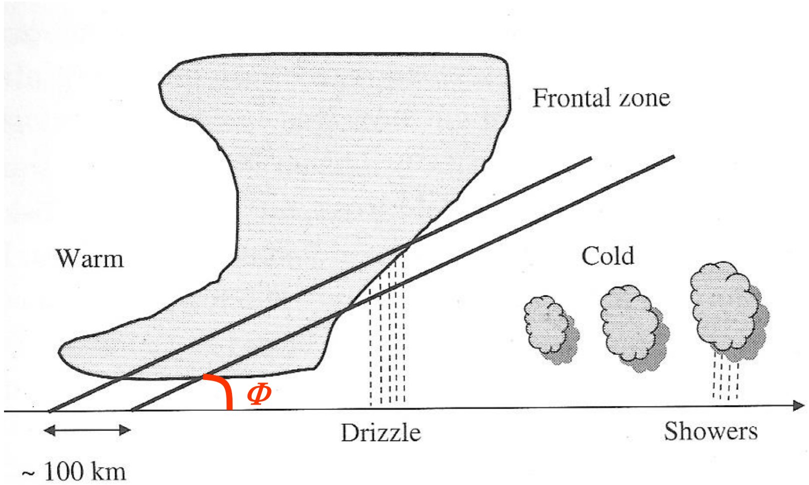
2.5. Horizontal wind shear

Boundary between two air masses: *front.*
Transition region: *frontal zone.*
Colder (denser) mass usually below the warmer (less dense) one.

Slope of front :

proportional to wind shear

inversely proportional to ΔT



Idealised frontal zone

Working as before, it is possible to demonstrate Margules' formula:

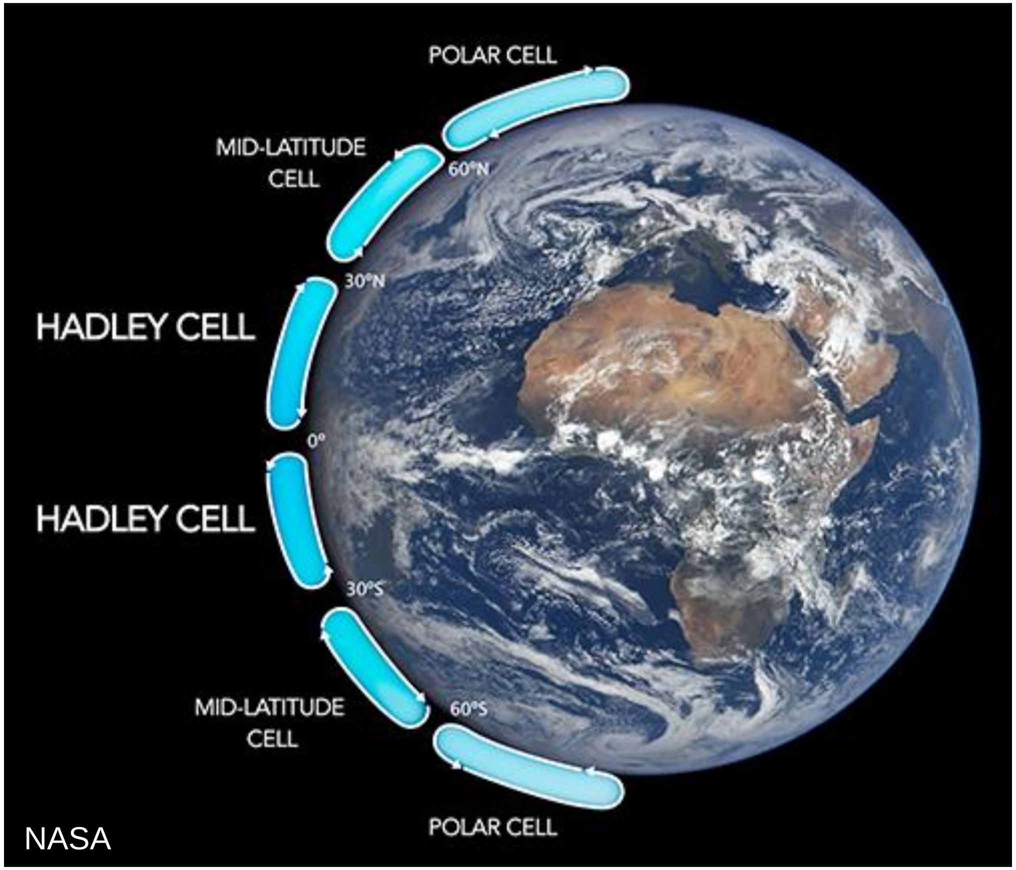
$$\boxed{\tan\Phi = \frac{dz}{dx} \approx \frac{f\,T_{mean}}{g}\left(\frac{u_{g\,cold} - u_{g\,warm}}{T_{warm} - T_{cold}}\right)}$$

When there are no clouds:

Dark surfaces absorb heat

When there are clouds:

Their tops reflect sun radiation away (cooling the Earth's surface)



Climate change is expanding the tropical cells, which pushes high-altitude clouds poleward

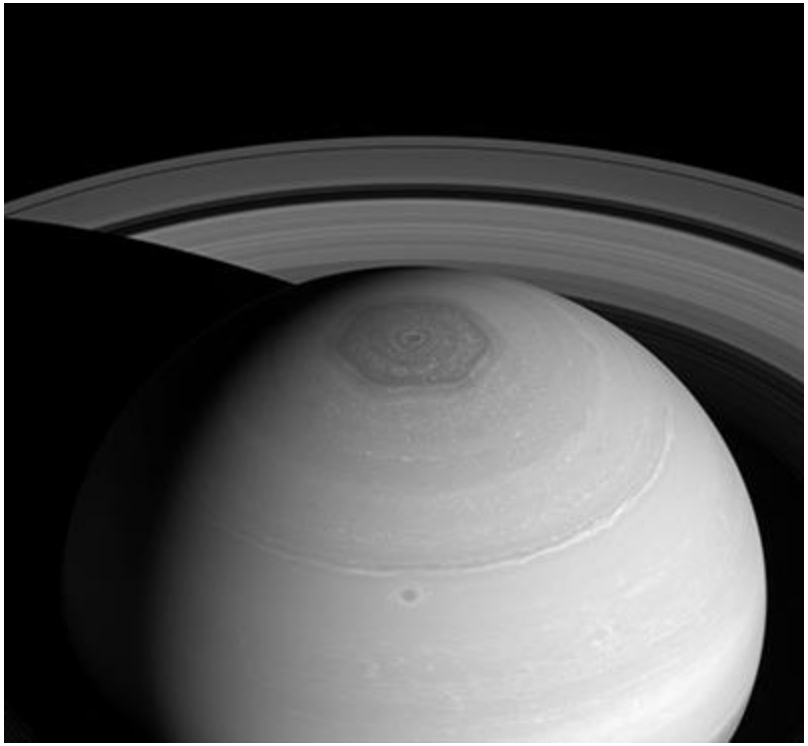
Based on analyses of 30+ years of satellite data by Tselioudis et al. (Geophys. Res. Letters, 2016)

To answer a student's question:

"The Hexagon" on Saturn is a six-sided north polar jet stream.

Voyager (1980) hinted at the structure.
Cassini (1997-2017) confirmed it.

25,000 km wide
100 km down? (from thermal imaging)



Credit: NASA/JPL/Space Science Institute

Saturn's hexagon might be what an undisturbed polar jet stream looks like in an atmosphere devoid of land and oceans.

Why the jet stream is six-sided rather than five- or eight-sided is still a mystery.

More Cassini discoveries at: https://eos.org/features/saturn-unveiled-ten-notable-findings-from-cassini-huygens

SUMMARY

1. Vertical Movements
   1.1. Large-scale vertical structure
   1.2. Hydrostatic equation
   1.3. Lapse rate – Adiabat
   1.4. Saturated lapse rate – Vertical stability
   1.5. Cloud formation – Precipitation

2. Horizontal Movements
   2.1. General circulation
   2.2. Driving forces
   2.3. Geostrophic flow
   2.4. Vertical wind shear
   2.5. Horizontal wind shear (weather fronts)

3. Remote Sensing of Planetary Atmospheres
   3.1. Identifying elements and volumes
   3.2. Reflected/scattered components
   3.3. Thermal component

4. Example: Jupiter
   4.1. General view
   4.2. Thermal emission
   4.3. Zones and belts
   4.4. The Great Red Spot + Red Junior
   4.5. And further away …