Sao Paulo School of Advanced Science on Atmospheric Aerosols

Introduction to Cloud Physics
And
Cloud - Aerosol Interactions

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Cloud Physics Group


Sun-Earth Day 2008: Space Weather Around the World
sunearthday.nasa.gov

Topics (I wish) to cover:

(1) What are clouds
(2) Clouds role in climate
(3) Possible cloud feedbacks in response to anthropogenic actions and climate change
(4) Components of cloud physics - Why clouds are regarded as (super) complex system
(5) Approaches to clouds research
(6) Scales
(7) Parcel(s) theory
(8) Basic nucleation
(9) Diffusion
(10) Collision – coalescence
(11) Twomey effect
(12) Bulk vs. Bin
(13) Invigoration
(14) Entrainment and mixing
(15) Non monotonic effects
(16) COG
(17) Effective radius – Effective terminal velocity
(18) Phase spaces
(19) Core vs. margins
(20) Semi-direct effect
(21) Nonlinear approaches – organization – networks
(22) Shortwave vs longwave effects – anvils vs. shallow Cu
(23) Warm clouds as the initial conditions for mix and cold
(24) Twilight

Textbooks (partial list):

Clouds in a glass of beer: simple experiments in atmospheric physics, Craig F. Bohren 1987

A Short Course in Cloud Physics - R R Rogers and M. K. Yau, 1989

Storm and Cloud Dynamics - William R. Cotton, George Bryan, Susan C van den Heever, 2010

Cloud dynamics - Robert A. Houze, 2014

Microphysics of Clouds and Precipitation – Pruppacher and Klett, 2010

Atmospheric Science: An Introductory Survey - Wallace and Hobbs, 2006

Physical Processes in Clouds and Cloud Modeling – Khain and Pinsky, 2018

Here we will disagree with the following statements:

(1) Warm clouds are well understood the challenges are understanding mix and cold phase clouds

(2) Warm, small, convective clouds are lass important for climate

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(1) Warm clouds are well understood the challenges are understanding mix and cold phase clouds

(2) Warm, small, convective clouds are lass important for climate

Hope to convince you that the above two statements are wrong

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Flowchart:

Introduction to clouds and their role in climate
Basic cloud intuition – parcel view

Life beyond parcel view – warm cloud microphysics
Aerosol effect - direct feedback related to nucleation
Links to cloud dynamics

Zooming out – cloud field view – larger scale feedbacks
Aerosol effects on the field scale – inside and outside of clouds

Time to do some cloud observations

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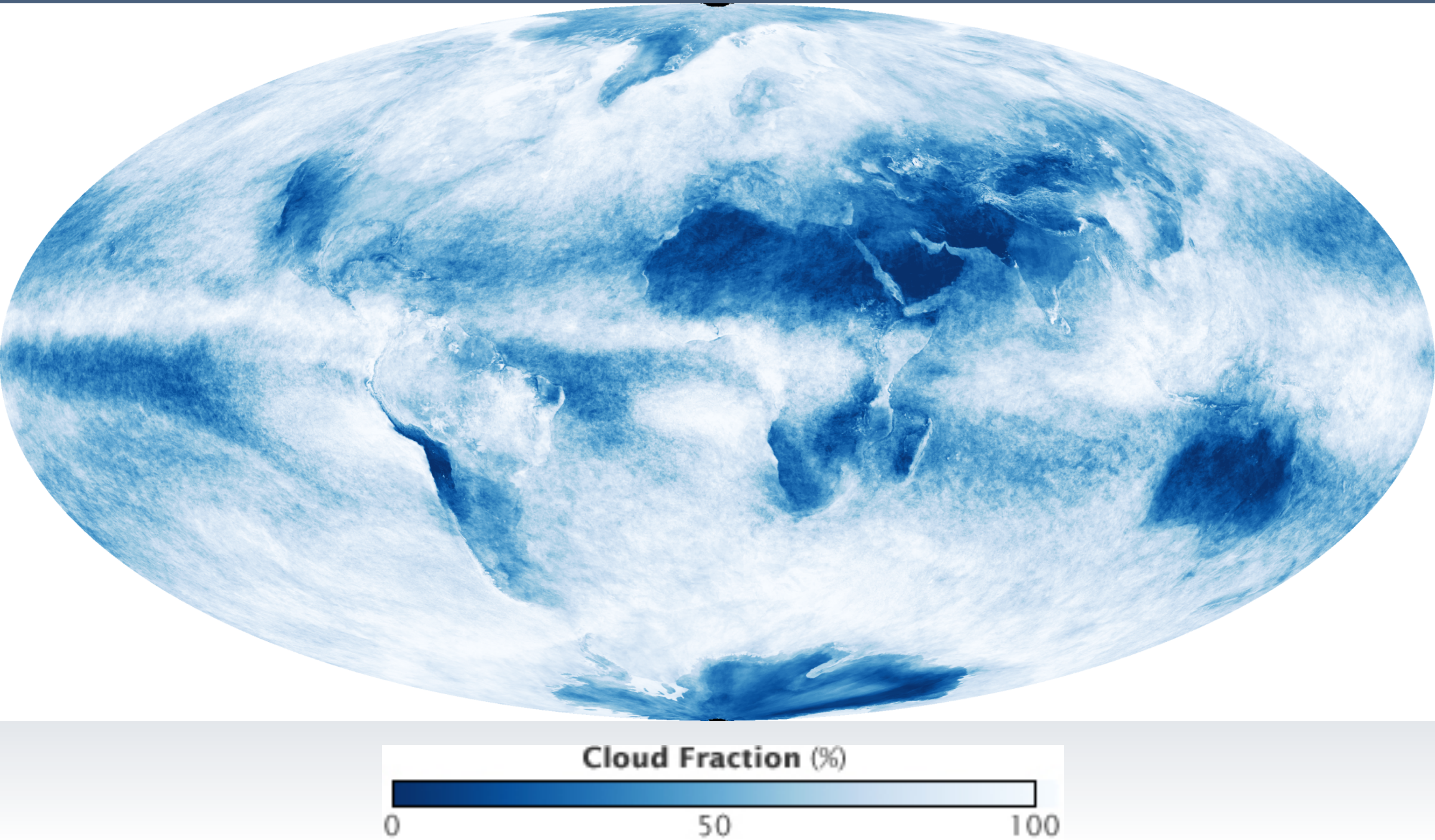
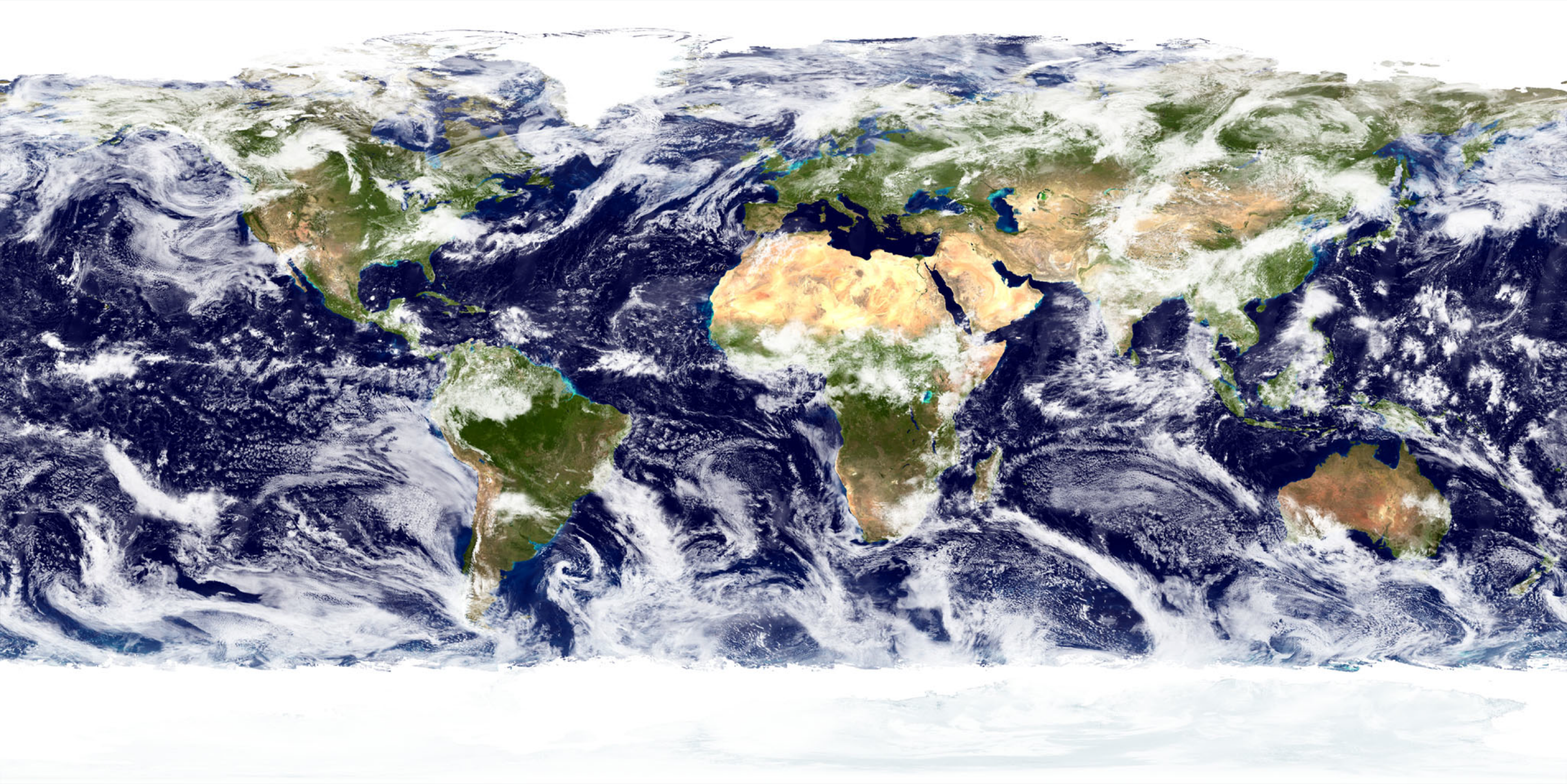
Cloud Fraction





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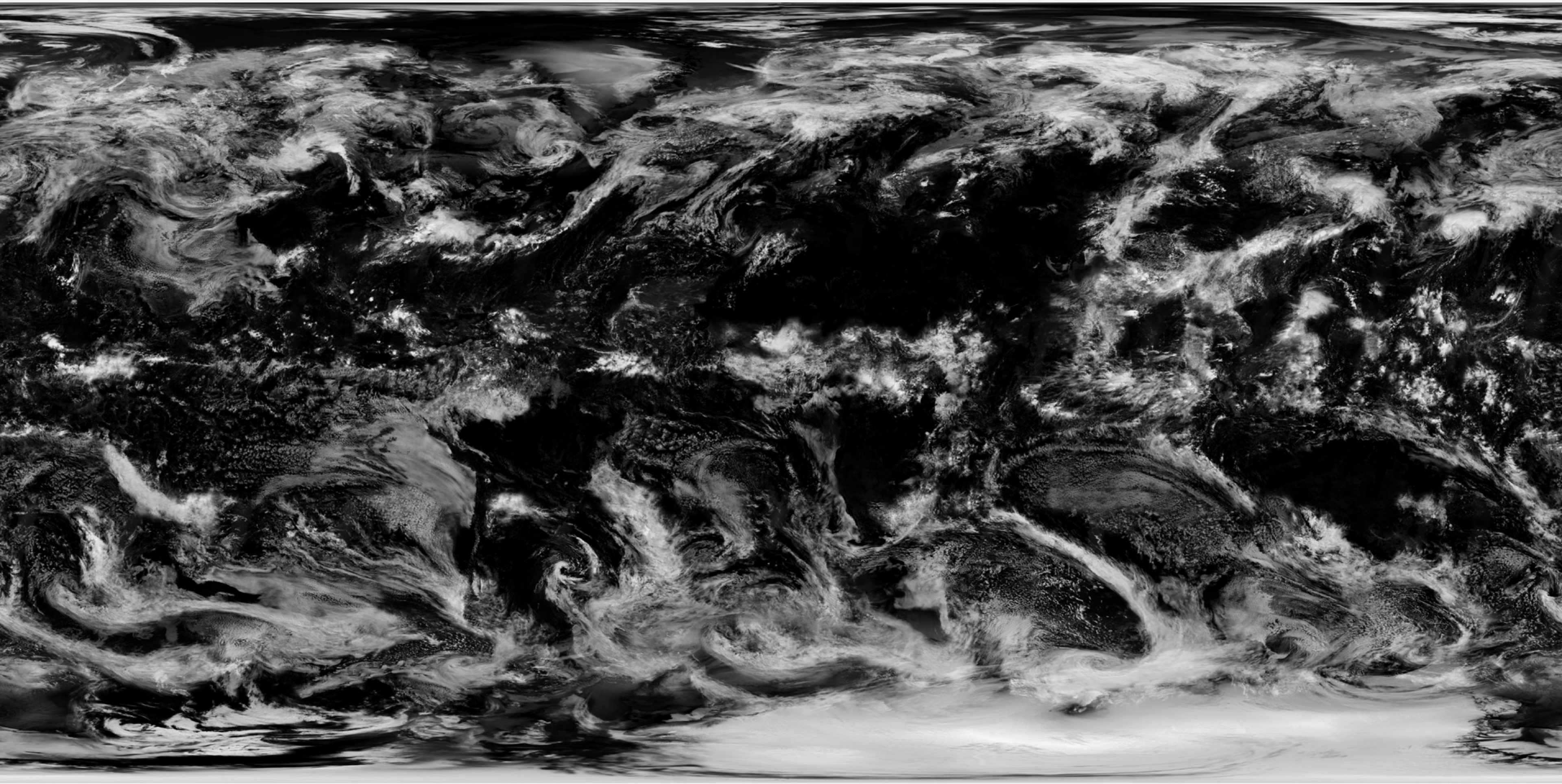

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Clouds in the climate system

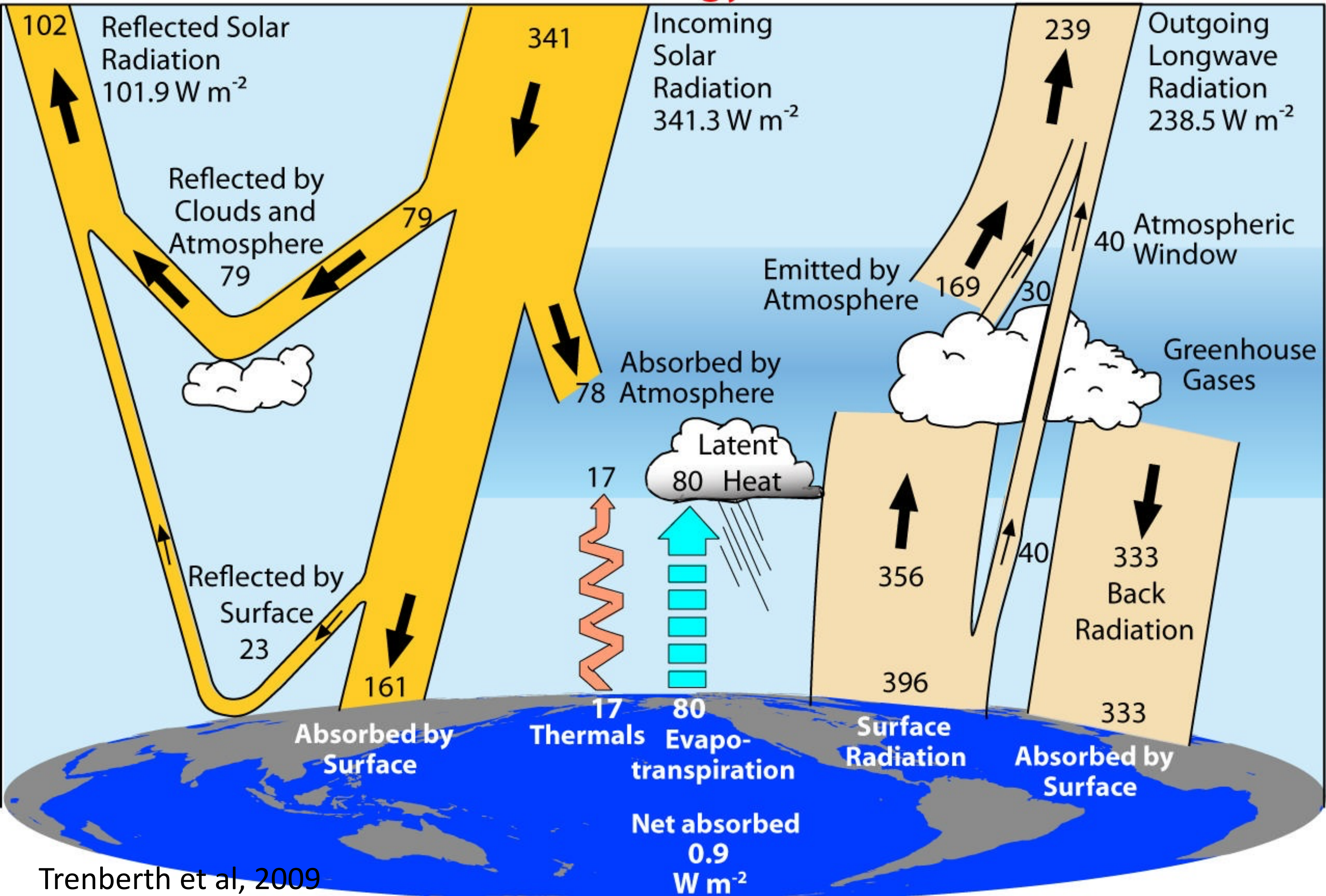
Clouds play a critical role in the Earth's climate system. They modulate the energy budget and fully control the water cycle.

Clouds are responsible for ~2/3 of the Earth albedo.

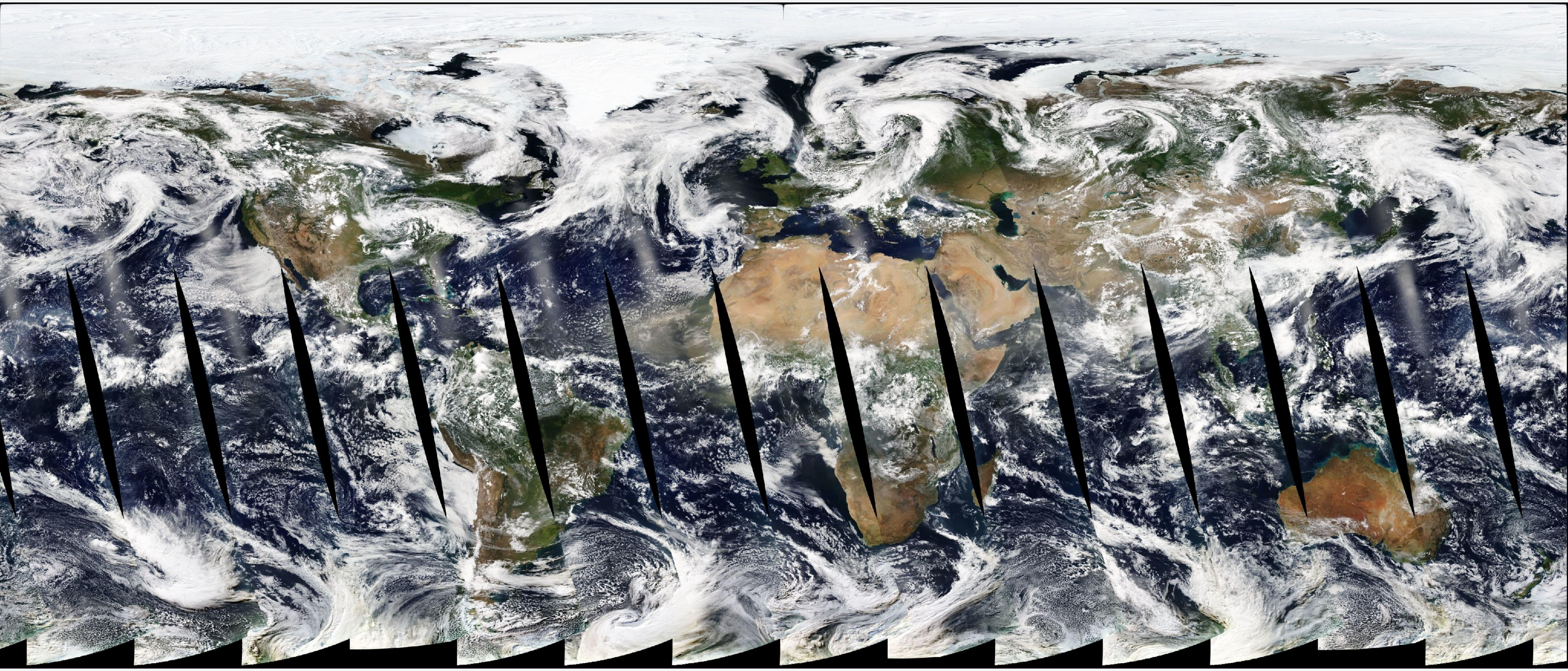
An error of ~1% in cloud properties is on the order of the climatic effect of manmade greenhouse gases.

Inaccurate description of cloud properties yields the largest errors in climate predictions.

Global Energy Flows W m^{-2}



Trenberth et al, 2009



2018 JUN 22

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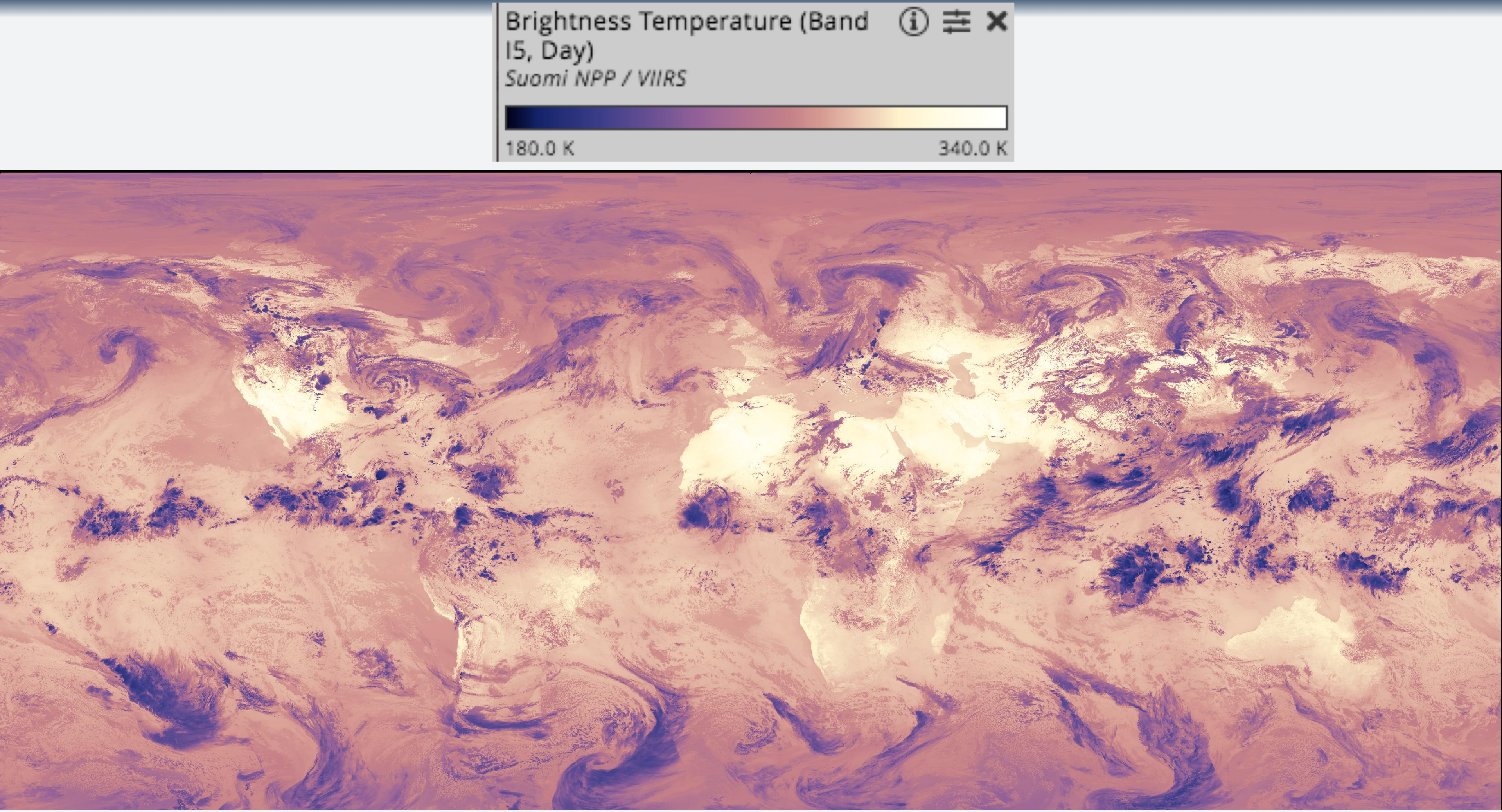

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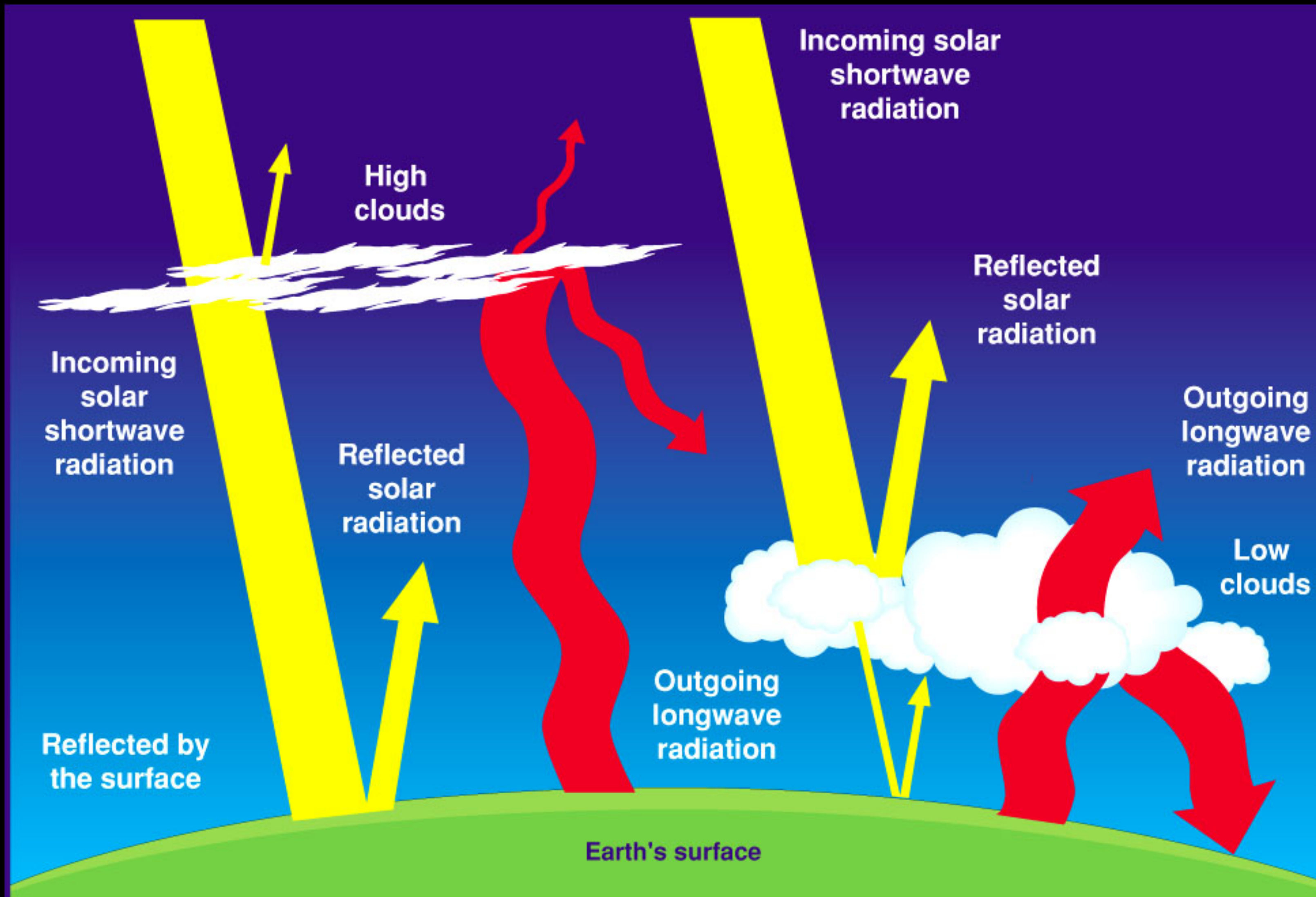
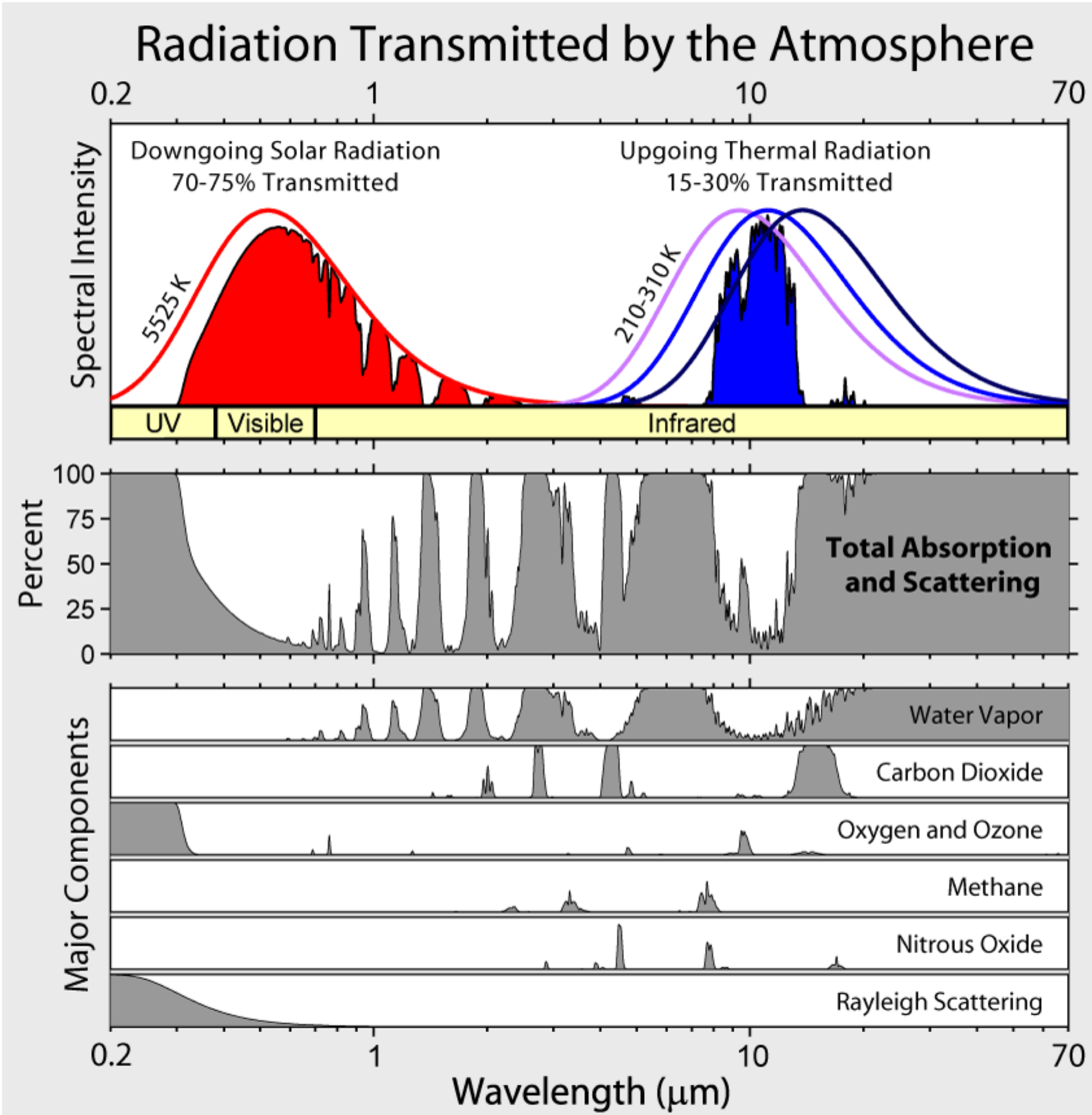
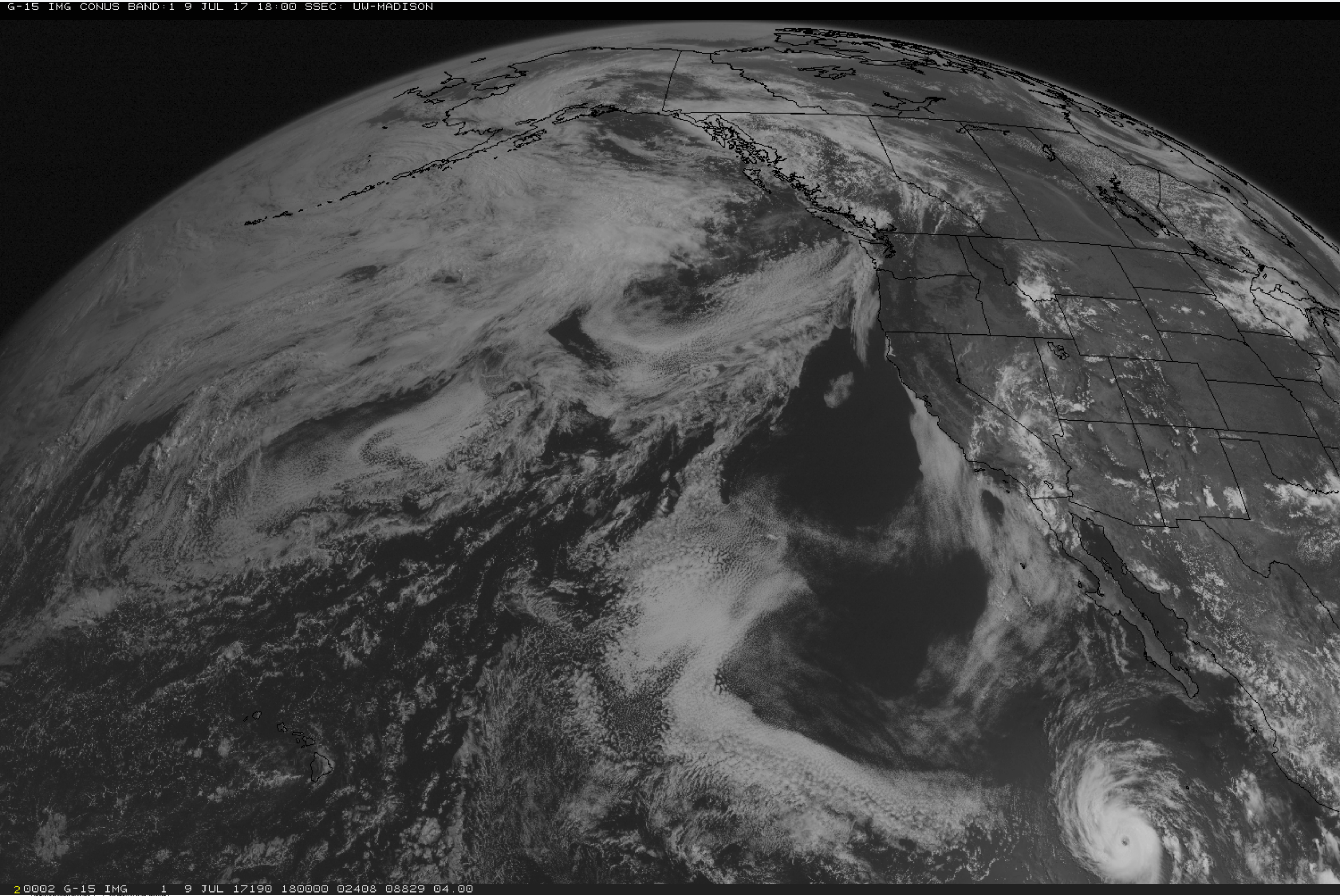
Cloud Effects On Earth's Radiation



Water vapor feedback vs. cloud feedback



Robert A. Rohde, the Global Warming Art project.



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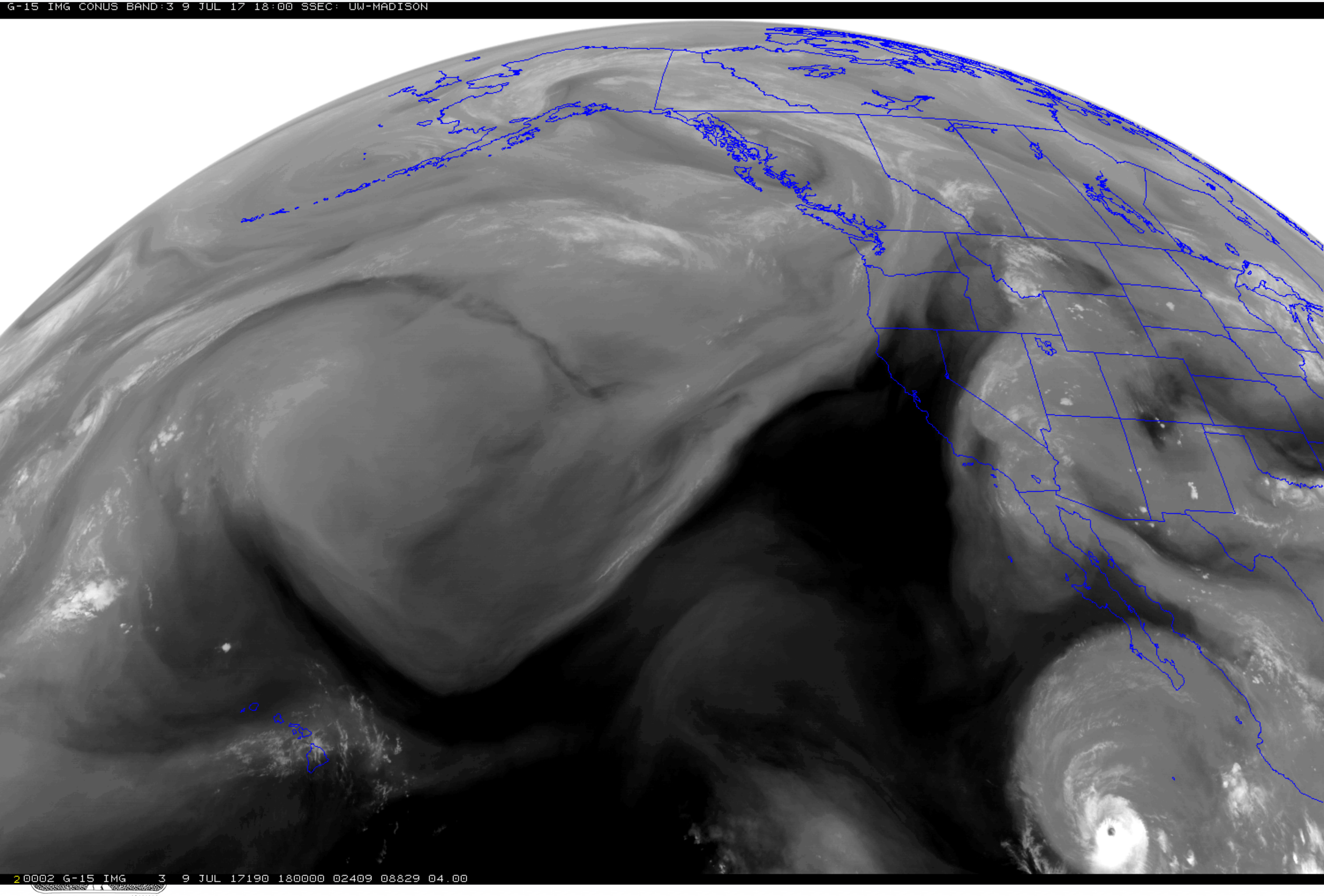
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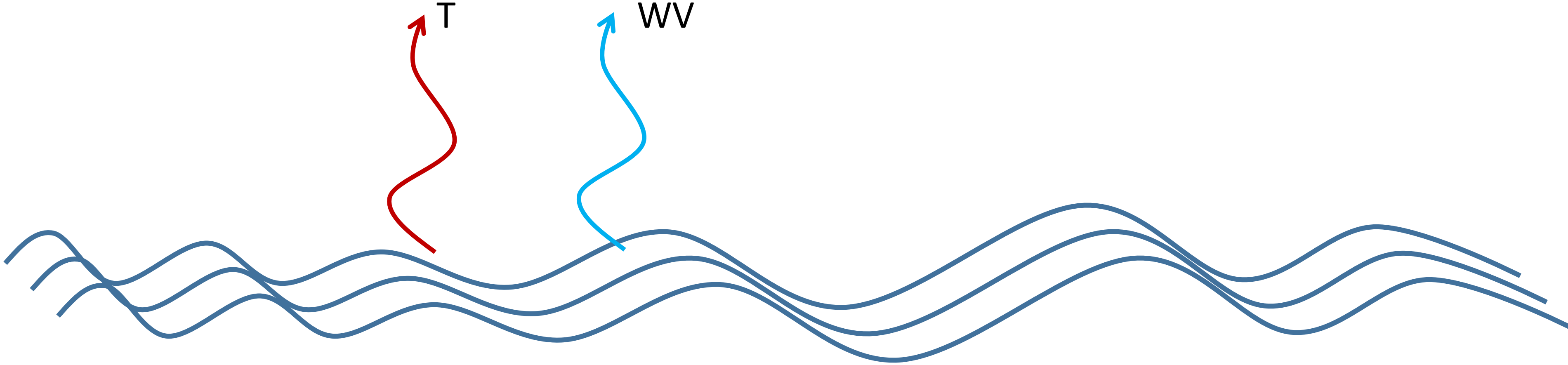
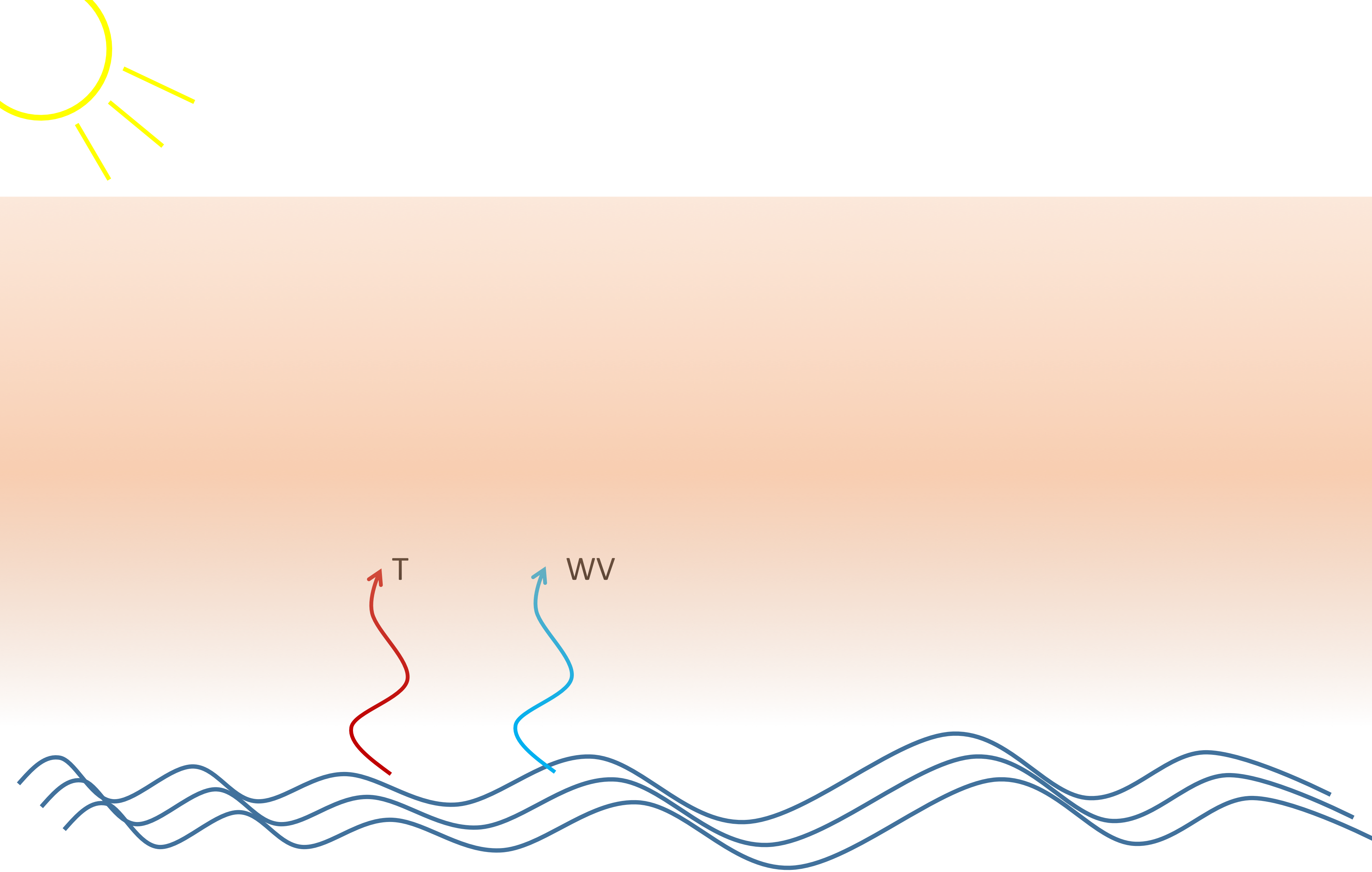
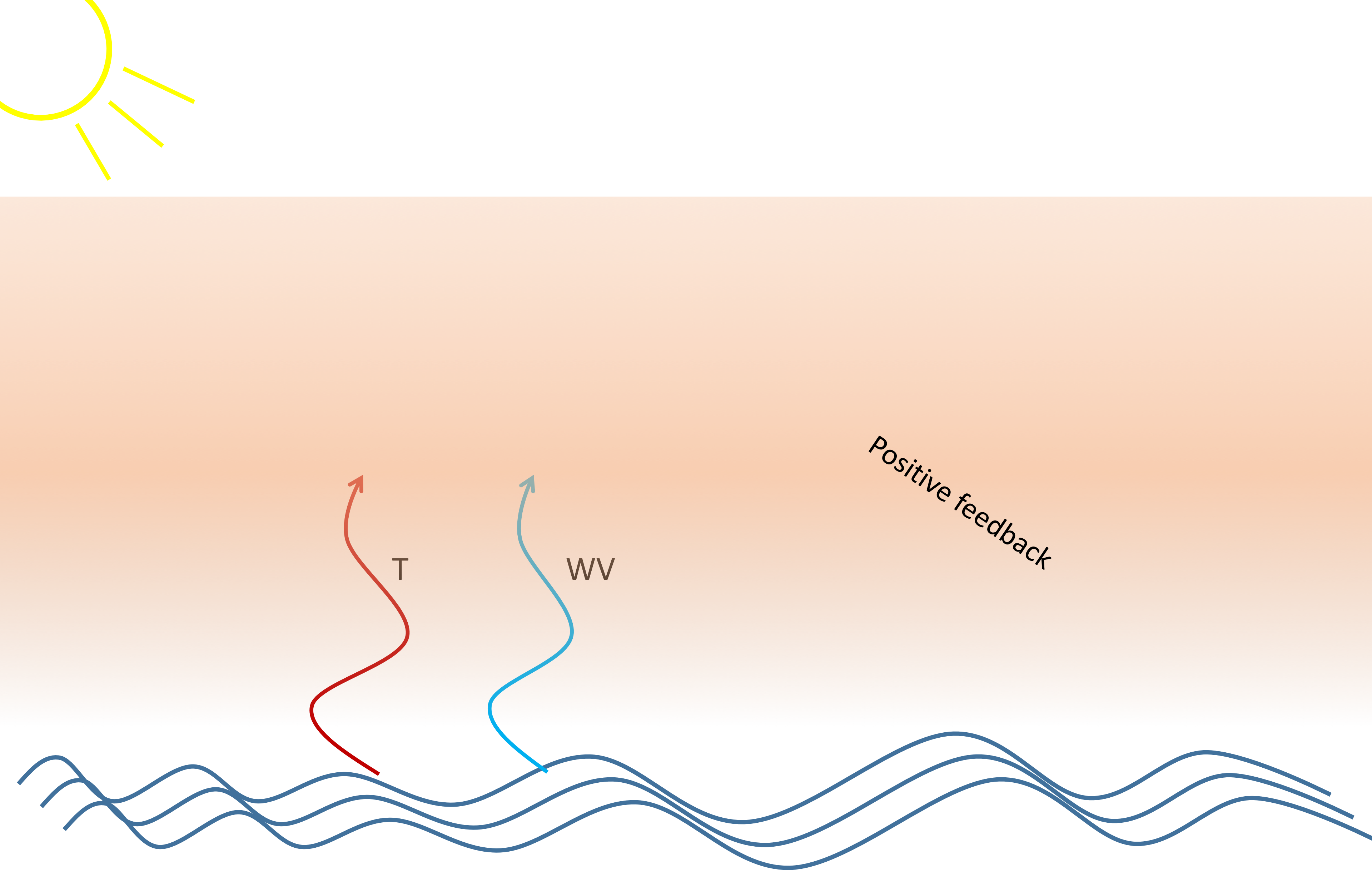
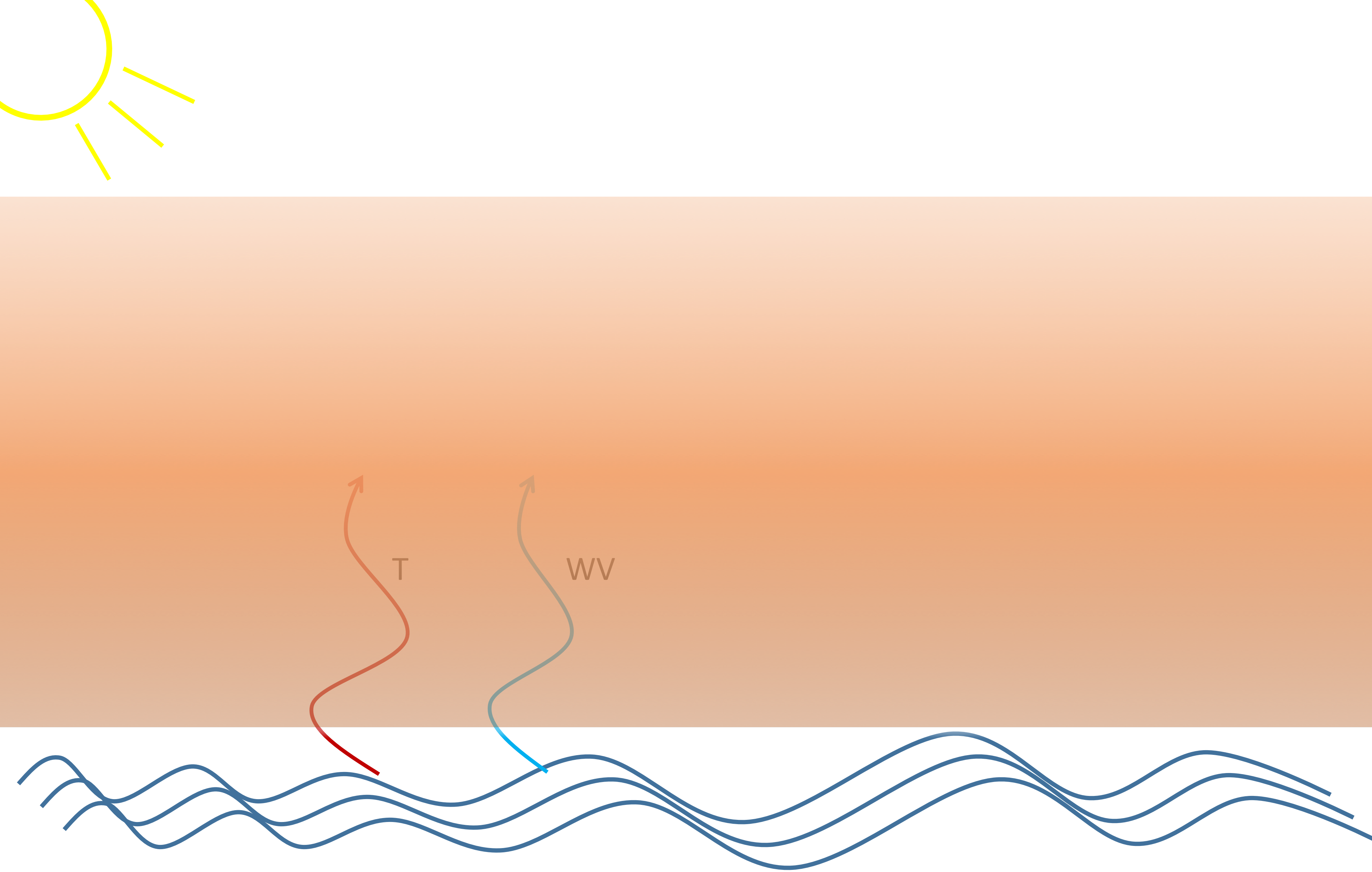
WV feedback:

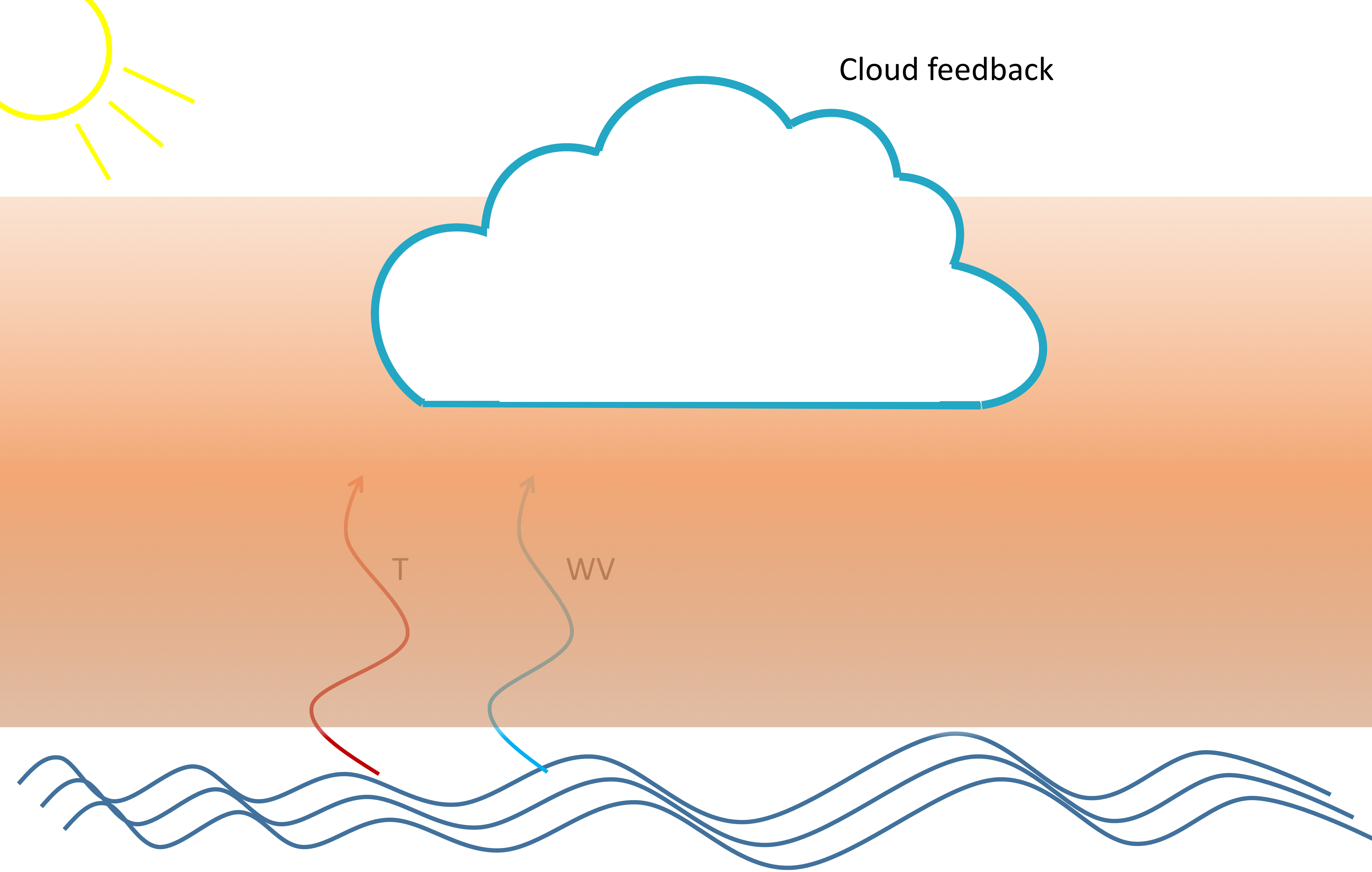
Suppose we face global warming



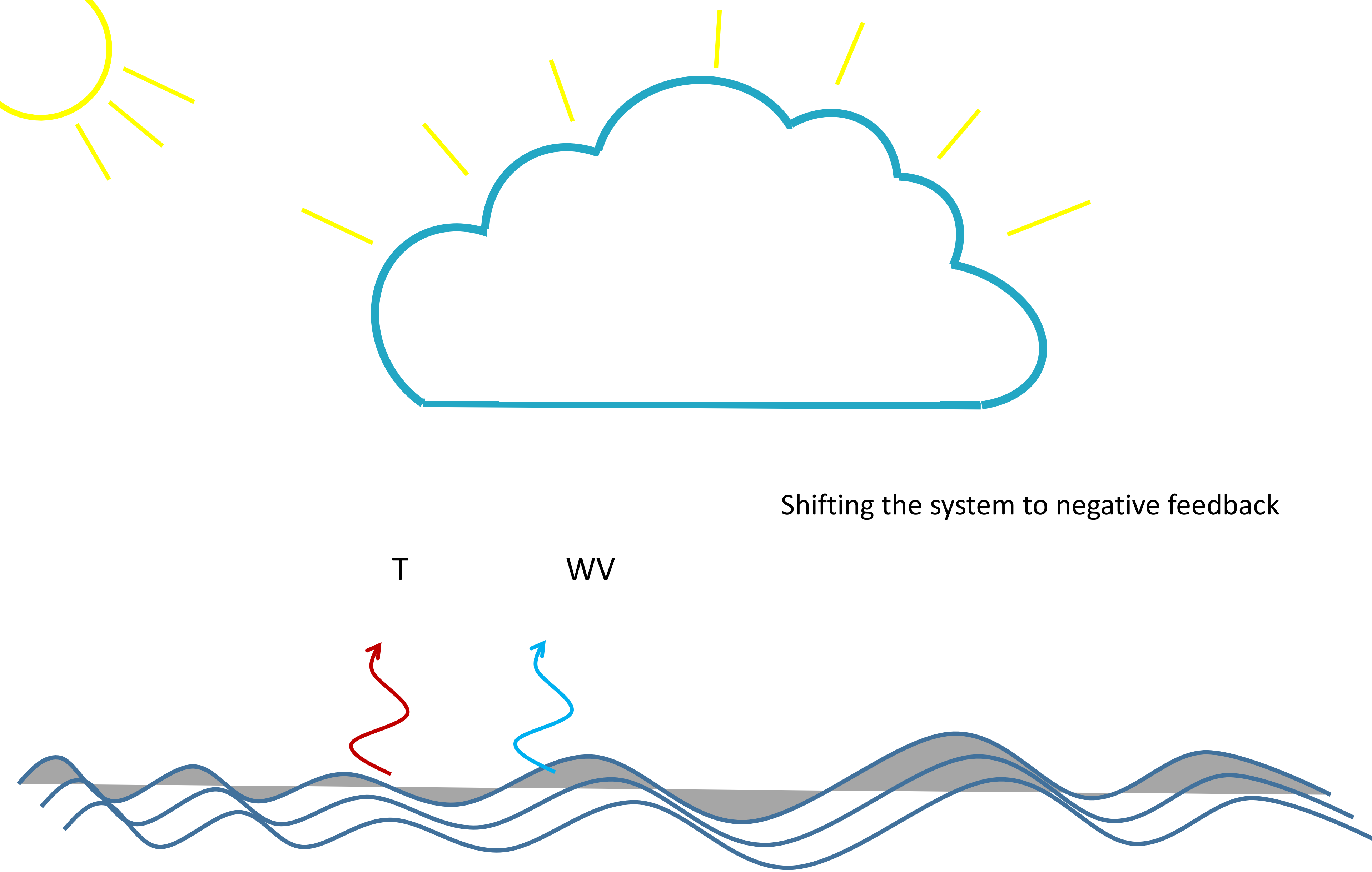

T
WV


T
WV
Positive feedback

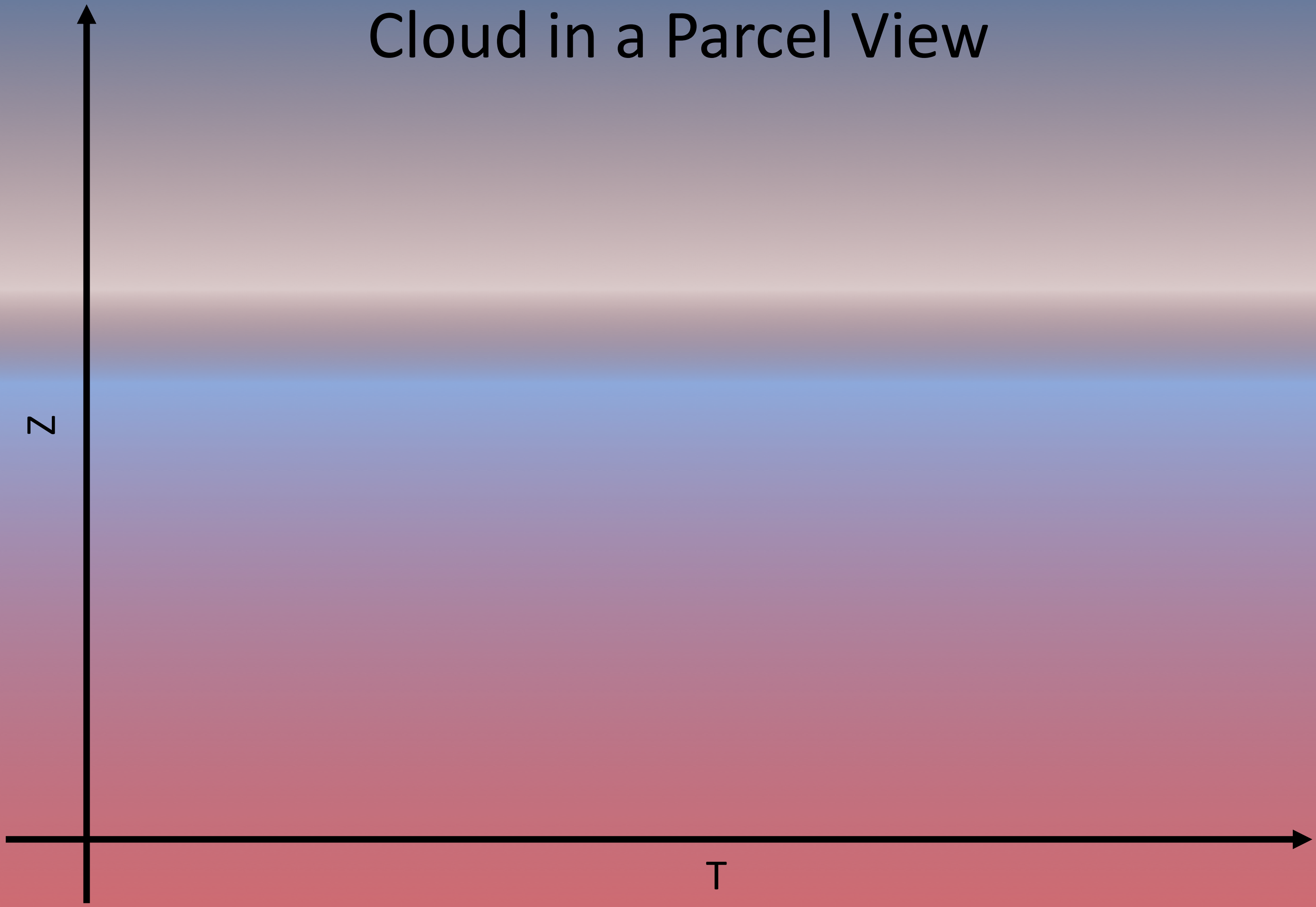

T
WV

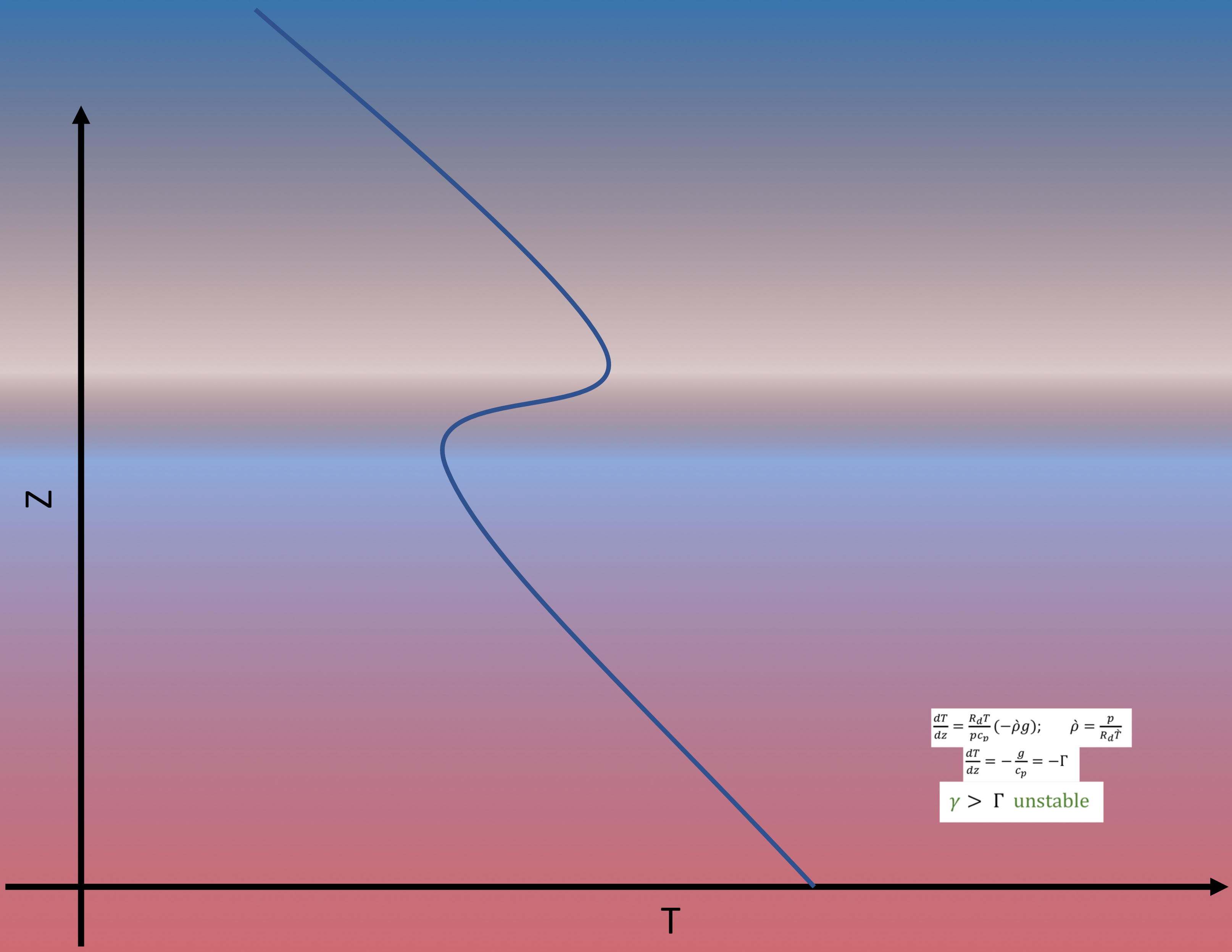


Cloud feedback
T
WV


Shifting the system to negative feedback
T
WV

Cloud in a Parcel View

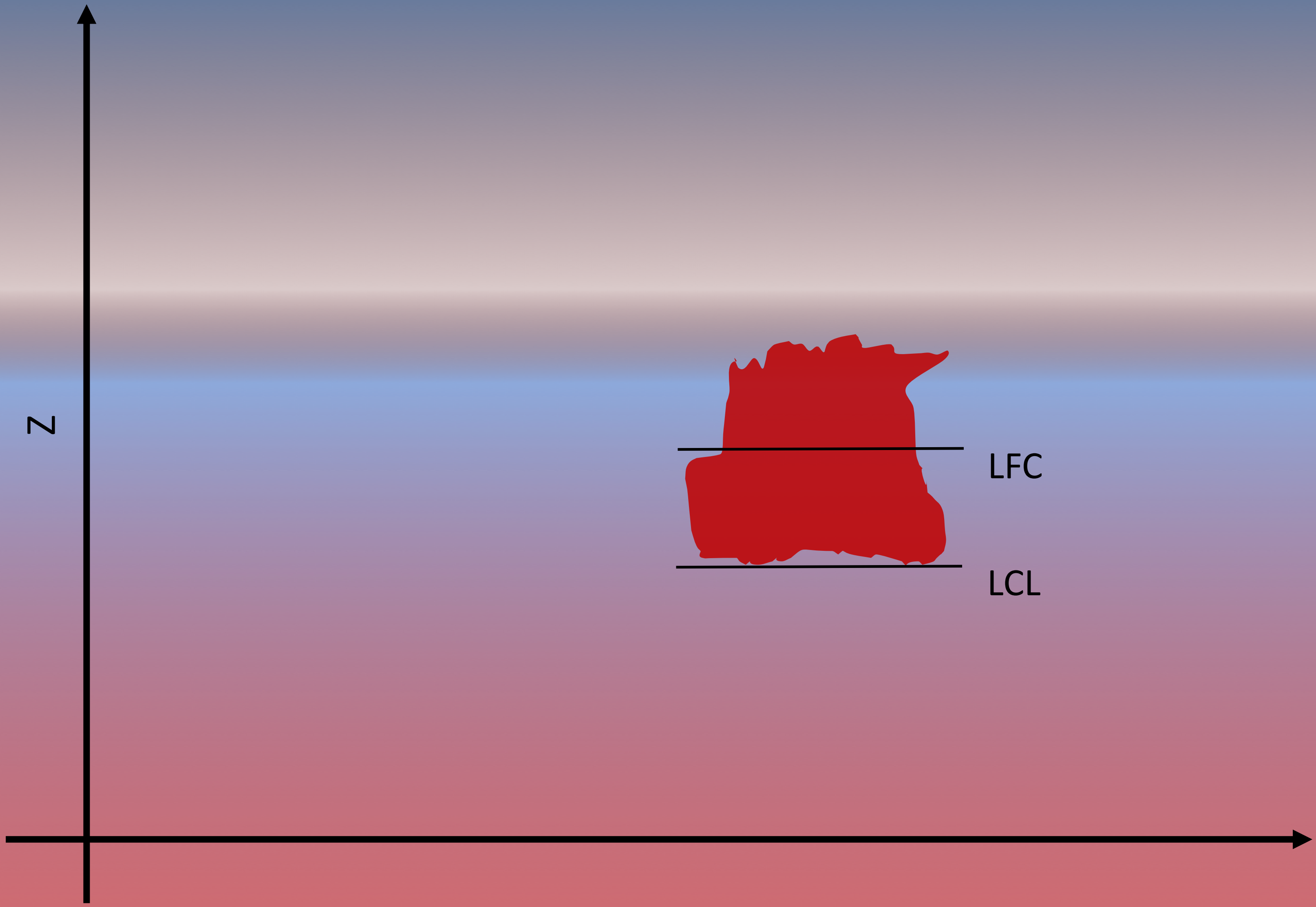




$$\frac{dT}{dz} = \frac{R_d T}{p c_p}(-\grave{\rho} g); \qquad \grave{\rho} = \frac{p}{R_d \grave{T}}$$

$$\frac{dT}{dz} = -\frac{g}{c_p} = -\Gamma$$

$\gamma > \Gamma$ unstable


Z
LFC
LCL

Setting the stage for a parcel model

Scanning fast the basic equations:

First law of Thermodynamics

(1) $dq = du + dw$
(2) $dq = c_v dT + p d\alpha$

Equation of state for ideal gas (Avogadro's number, Dalton's law, Charles' laws, Boyle's law)
per unit mass
(3) $p\alpha = R_d T$
(4) $\alpha dp + p d\alpha = R_d dT$

Second law of Thermodynamics - Entropy
(5) $d\phi = \frac{dq}{T}$

(1) $dq = du + dw$

(2) $dq = c_v dT + p d\alpha$

(3) $p\alpha = R_d T$

(4) $\alpha dp + p d\alpha = R_d dT$

(5) $d\phi = \frac{dq}{T}$

Using (4) eq (2) could be expressed:

(6) $dq = c_v dT + R_d dT - \alpha dp$

In isobaric process $dp = 0$

(7) $dq = (c_v + R_d)dT$, but $\left(\frac{dq}{dt}\right)_p = c_p dT$

(8) $c_v + R_d = c_p$

The first law can then have the forms:

(9) $dq = c_p dT - \alpha dp = c_v dT + p d\alpha$

In adiabatic process $dq = 0$

From eq. 6

(10) $c_p dT = \alpha dp = R_d T \frac{dp}{p}$

(11) $\frac{dT}{T} = \frac{R_d}{c_p}\frac{dp}{p}$

integrating

(12) $\frac{T}{T_0} = \left(\frac{P}{P_0}\right)^k ; k = \frac{R_d}{c_p}$

let $T_0 = \theta$, and P_0 be the ~surface pressure 100kPa

(13) $\theta = T\left(\frac{100kPa}{P}\right)^k$

θ is defined as the potential temp

(14) $d\phi = \frac{c_p dT - \alpha dp}{T} = c_p \frac{dT}{T} - R_d \frac{dp}{p} = c_p\left(\frac{dT}{T} - k\frac{dp}{p}\right) = c_p \frac{d\theta}{\theta}$

Integration links entropy with θ showing that adiabatic processes are also isentropic ones

(15) $\phi = c_p ln\theta + Const$

Potential temperature

$$\theta = T\left(\frac{P_{ref}}{P}\right)^{k}$$

θ is the temperature of a parcel lifted from P_{ref} & T adiabatically to pressure P

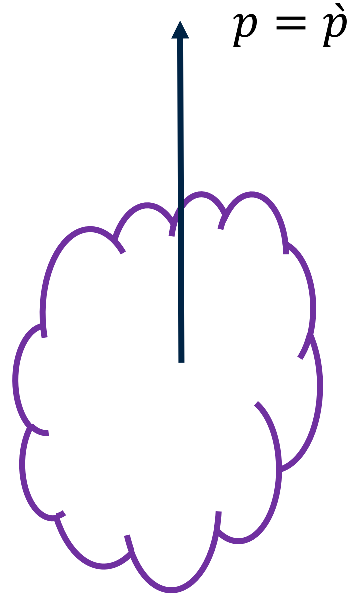
Adiabatic processes are also isentropic ones

$$\phi = c_p ln\theta + Const$$

Buoyancy force of a dry-air parcel

Consider a parcel with volume V that displace the same volume of the ambient air. The pressure gradient force is the same on the parcel and its surroundings. The parcels acceleration per unit mass will differ from the displaced air

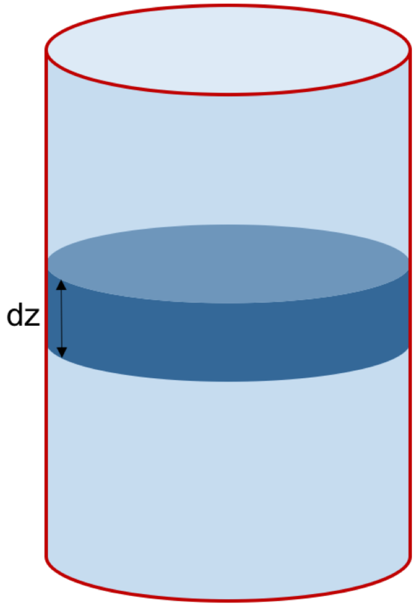
(23) $B = g(\frac{\grave{\rho}-\rho}{\rho})$



Since V and p are the same the density scales inversely to the temperature:

(24) $B = g(\frac{T-\grave{T}}{\grave{T}})$

About hydrostatic equilibrium



Pressure at height Z is the
force per unit area extracted by all of the air mass above it.

$$P(z) = -g\int_z^{\infty} \rho dz$$

(16) $\frac{\partial p}{\partial z} = -\rho g$

A force balance per unit mass between the pressure gradient force and gravity.

$$\frac{\partial p}{\partial z} = -\rho g = -\frac{P}{RT} g$$

if the air contains WV (moist air) the gas constant R has to be changed to account for it. Instead we can use a correction to the temperature that takes in account the air WV content. We define this temperature as the Virtual Temperature (T_v). T_v (which will be defined more precisely later) is a useful thermodynamic variable in application that include moist air. It can be easily calculated once the WV content is known and it allows using the dry air constants.

$$(17)\ \frac{\partial p}{\partial z} = -\frac{P}{R_d T_v} g$$

As a first approximation we can estimate p using the average T_v and integrating eq. (17) to:

$$(18)\ P = P_0 e^{-\frac{z}{H}}$$

where $H = \frac{R_d \overline{T_v}}{g}$ is the local scale height. $R_d = 287 J kg^{-1}$ and assuming $T_v \approx 270K$ yields H~8km.

Dry adiabatic lapse rate

Recall $dq = c_p dT - \alpha dp$

In adiabatic process $dq = 0$, and replacing α using the eq. of state, the above yields:

$$c_p dT = R_d T \frac{dp}{p}$$

Taking the z derivative:

$$\frac{d}{dz}(c_p dT) = \frac{d}{dz}(R_d T \frac{dp}{p})$$

(19) $\frac{dT}{dz} = \frac{R_d T}{p c_p} \frac{dp}{dz}$

Using the hydrostatic eq. assuming an instantaneous adjustment of the pressure to the ambient (surrounding) one.

(20) $\frac{dT}{dz} = \frac{R_d T}{p c_p}(-\grave{\rho} g)$; $\quad \grave{\rho} = \frac{p}{R_d \grave{T}}$

Where $\grave{\rho}$ and $\grave{T}$ are the ambient density and temperature. Combining eqs. 19 and 20:

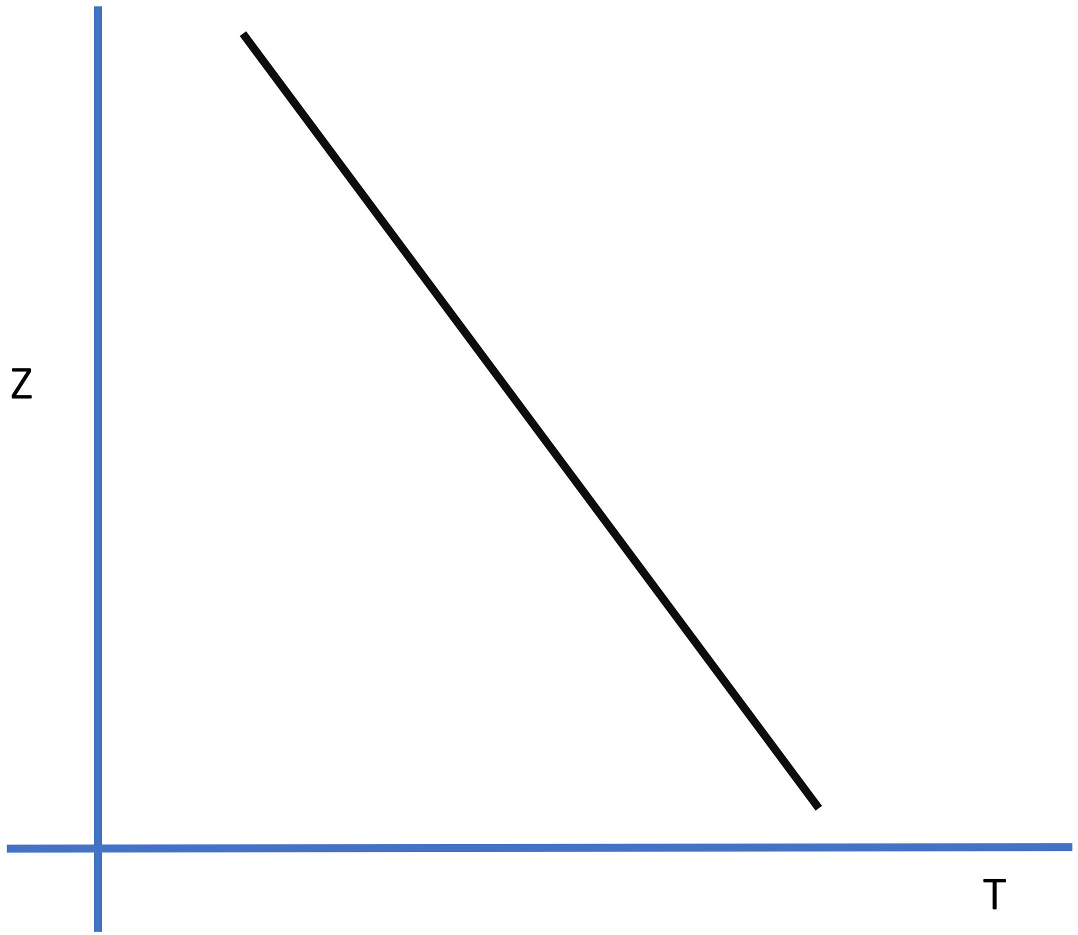
(21) $\frac{dT}{dz} = -\frac{g}{c_p}\frac{T}{\grave{T}}$

Assuming that the temperature of a parcel is not significantly different than the surrounding (O ~1K), $\frac{T}{\grave{T}} \cong 1$ and therefore we get a constant value for the dry adiabatic lapse rate:

(22) $\frac{dT}{dz} = -\frac{g}{c_p} = -\Gamma$

Stability and parcel perturbations

Parcel lapse rate Γ
Ambient lapse rate γ



Stability and dry parcel perturbations

Parcel lapse rate Γ
Ambient lapse rate γ

$$\gamma < \Gamma \text{ stable}$$

$$\gamma = \Gamma \text{ neutral}$$

$$\gamma > \Gamma \text{ unstable}$$

Recall that the driving force per unit mass is the buoyancy. With some approximations, we can estimate it as (HW):

(25) $$B = g\left(\frac{T - \grave{T}}{\grave{T}}\right) \approx -g\left(\frac{\Gamma - \gamma}{T}\right) z$$

Therefore if B represents the total forces acting on the parcel, we can express it as equation of motion:

(26) $$\frac{d^2 z}{dt^2} + \frac{g}{T}(\Gamma - \gamma) z = 0$$

which has harmonic structure and if we assume that the lapse rates are close to be constants (in the dry case they are), eq. (26) can be solved analytically:

$$z(t) = z_0 \cos(\omega t + \varphi)$$

for which ω is the Brunt–Väisälä frequency expressed as

(27) $$\omega = \left(\frac{g}{T}(\Gamma - \gamma)\right)^{1/2}$$

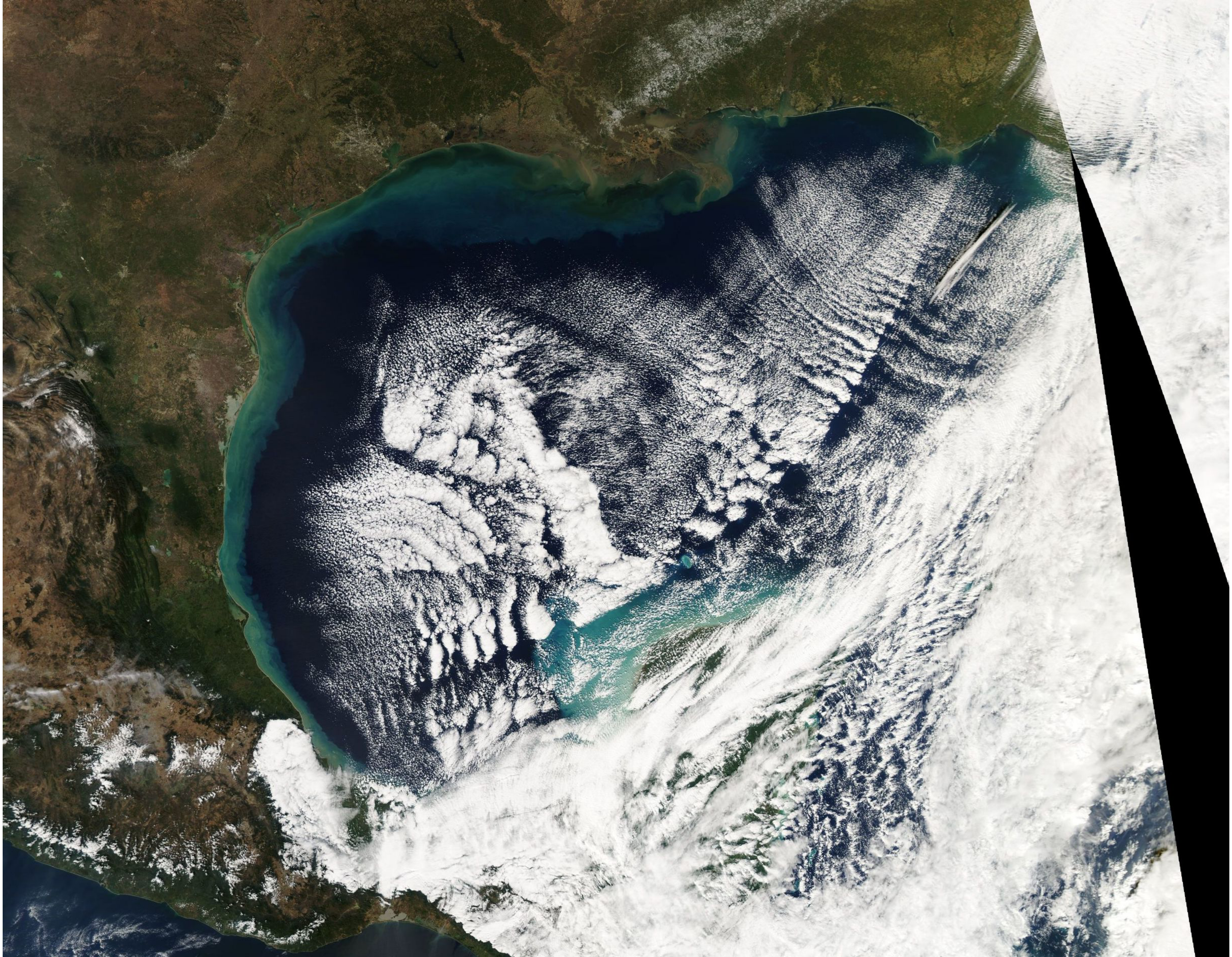
(27) $\omega = \left(\frac{g}{T}(\Gamma - \gamma)\right)^{1/2}$

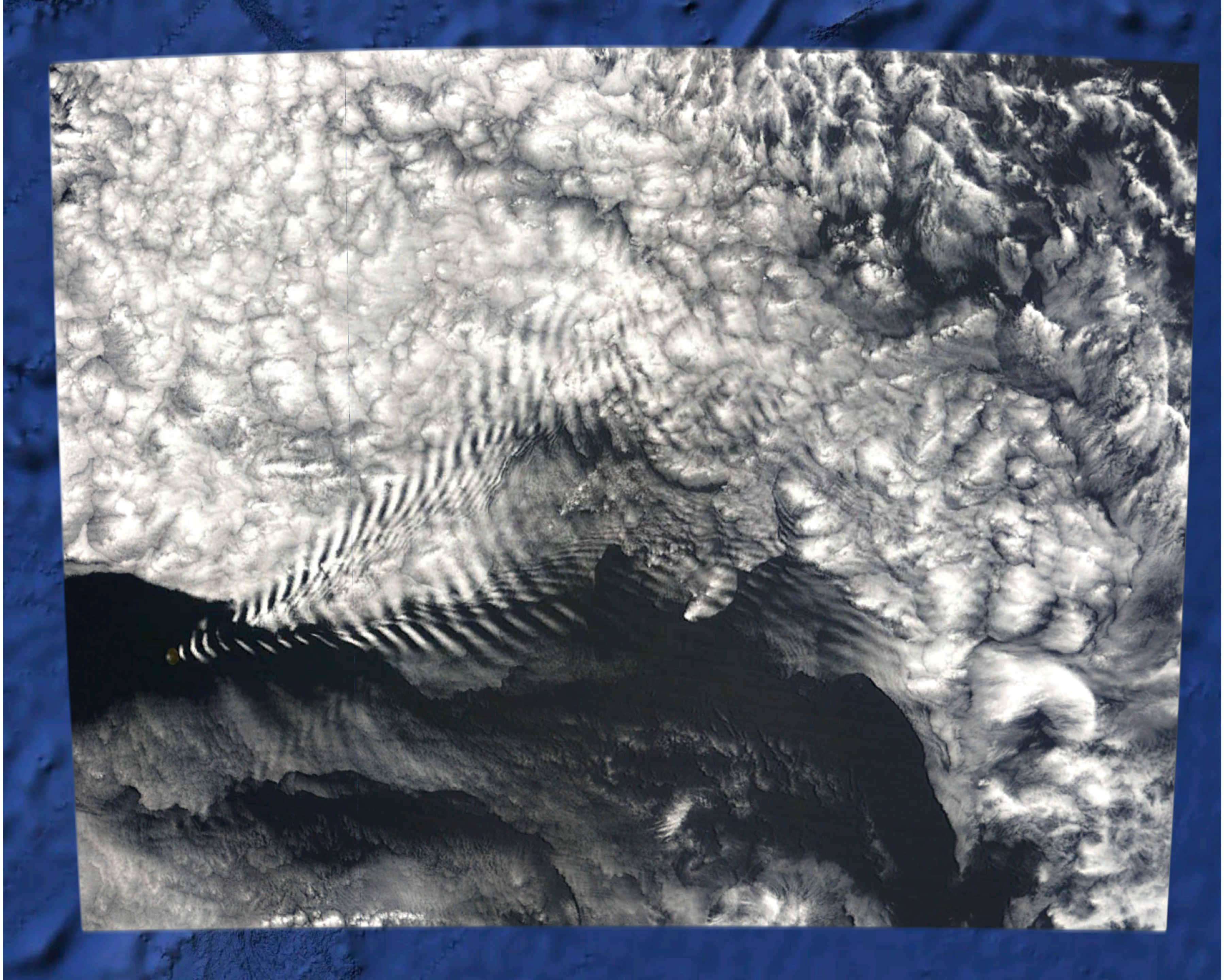
With characteristic BV periodicity

$$f = \frac{2\pi}{\omega} = \frac{2\pi}{\left(\frac{10}{300}\left(\frac{1}{1000}\right)\right)^{1/2}} \approx 20min$$

For small angles the pendulum periodicity $f = 2\pi\sqrt{\frac{l}{g}}$ the equivalent BV length is

$$\mathrm{l} = \frac{T}{(\Gamma - \gamma)} \approx 300km$$





Moist air
Ideal gases allow for several levels of linearities (superposition):

$P_{(moist\ air)} = P_{(dry)} + e_{(vapor)}$ each obey the eq. of state (with different constants)

For the moist air

$$P = \rho_m R_m T$$

for the dry

$p_d = \rho_d R_d T$ and

For the WV

(28) $e = \rho_v R_v T$

and the densities of the moist parcel

$$\rho_m = \rho_d + \rho_v$$

Recall that the i^{th} gas constant per unit mass (kg) is $R_i = 1000\frac{R^*}{M_i}$ where capital M denote the molecular weight, therefore

(29) $\varepsilon = \frac{R_d}{R_v} = \frac{M_v}{M_d} = 0.622$ and (28) could be expressed by means of R_d as

We wish to use R_d for cases of moist air for which the exact gas constant depends on the bland. To do so we define a correction for the temperature, a virtual temperature for which R_d can be used.

(30) $e = \rho_v \frac{R_d}{\varepsilon} T$

Expressing the partial eq. of state for the moist parcel above:

$$p = \rho R T = p_d + e = \rho_d' R_d T + \rho_v' R_v T$$

where ρ_d' , ρ_v' are densities of the same volume but considering only one subset (smaller). The total density are their sum:

$$\rho = \rho_d' + \rho_v' = \frac{p-e}{R_d T} + \frac{e}{R_v T} = \frac{p-e}{R_d T} + \frac{e}{\frac{R_d}{\varepsilon} T} = \frac{P}{R_d T}\left(1 - \frac{e}{p}(1-\varepsilon)\right)$$

$$\rho = \rho_d' + \rho_v' = \frac{p-e}{R_dT} + \frac{e}{R_vT} = \frac{p-e}{R_dT} + \frac{e}{\frac{R_d}{\varepsilon}T} = \frac{P}{R_dT}\left(1 - \frac{e}{p}(1-\varepsilon)\right)$$

Which is a modified form of the eq. of state:

$$\rho = \frac{P}{R_dT}\left(1 - \frac{e}{p}(1-\varepsilon)\right).$$

The correction term could be glued to the temperature which is then defined as Virtual Temperature allowing the use of dry constants for moist air:

(31) $p = \rho R_d T_v$

(32) $T_v = \frac{T}{\left(1-\frac{e}{p}(1-\varepsilon)\right)}$

(32) $T_v = \frac{T}{\left(1-\frac{e}{p}(1-\varepsilon)\right)}$

Since $\varepsilon < 1, T_v > T$ always. Some approximations can link T_v to the vapor mixing ratio

Assuming that $e \ll p$

$\frac{1}{\left(1-\frac{e}{p}(1-\varepsilon)\right)} \cong 1 + \frac{e}{p}(1-\varepsilon) = 1 + \frac{\omega}{\omega+\varepsilon}(1-\varepsilon)$, assuming that $\omega \ll \varepsilon$ and using $\varepsilon = 0.622$

(33) $T_v = T(1 + \frac{\omega}{\varepsilon}(1-\varepsilon)) \cong T(1 + 0.6\omega)$

Basic measures of moist air:
Mixing ratio (small (m) is for mass):

(34) $\omega \equiv \frac{m_v}{m_d}$

If there is no evaporation or condensation, the mixing ratio is a conserved quantity.

Vapor specific humidity is:

(35) $q \equiv \frac{m_v}{m_v+m_d} = \frac{\omega}{1+\omega} \cong \omega$

Consider a moist parcel for which: (HW1)

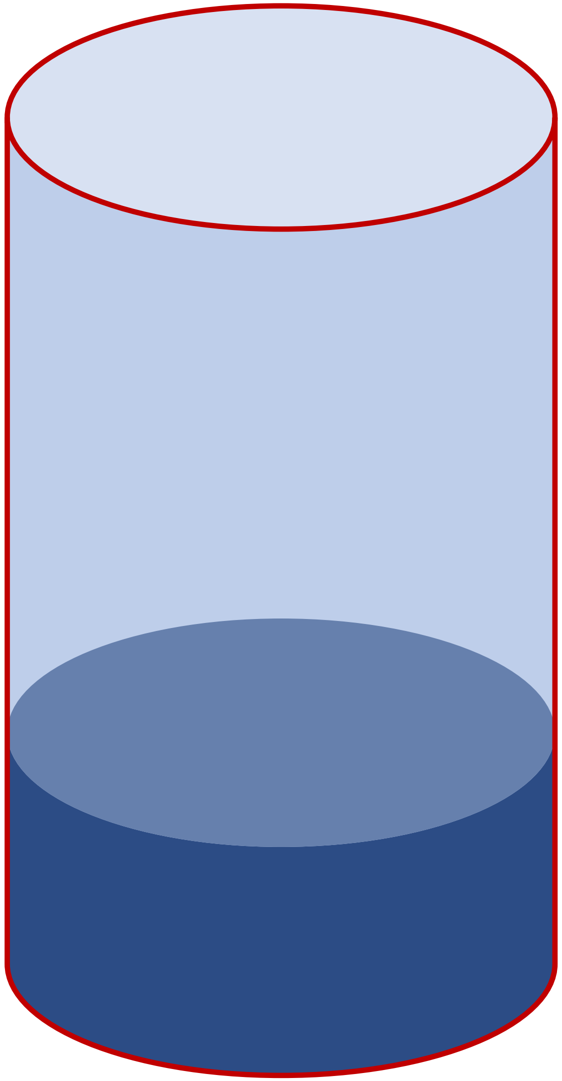
$$p = p_d + e$$
$$pV = nR^*T$$
$$eV = n_v R^*T$$
$$p_d V = n_d R^*T$$

(36) $\frac{e}{p} = \frac{n_v}{n_v+n_d} = \frac{m_v/M_v}{m_v/M_v+m_d/M_d} = \frac{\omega}{\omega+\varepsilon}$

A brief review on Clausius-Clapeyron EQ.

In a moist environment, we need to know the saturation vapor pressure and the latent heat release to understand feedbacks and estimate other key variables.

Consider an interface between water and air in a tank with fixed P and T. After enough time the system reaches a steady state such that there are no net fluxes of heat and of molecules.



$T_{air} = T_{water}$

$e = e_s$

$e_s = e_s(T)$

L the latent heat is the required energy to phase change unit of mass with fixed P and T

$T_{air} = T_{water}$; $e = e_s$ and $e_s = e_s(T)$

L [J] the specific latent heat is the required energy for a liquid (state1) to gas (state2) phase change of a unit mass with fixed P and T.
We can express L as:

(39) $L = \int_{q1}^{q2} dq = \int_{u1}^{u2} du + \int_{\alpha 1}^{\alpha 2} p d\alpha = u2 - u1 + e_s(\alpha 2 - \alpha 1)$,

Since T is constant
(40) $L = T \int_{q1}^{q2} \frac{dq}{T} = T(\phi_2 - \phi_1)$

From Eqs. (39) and (40) in isothermal isobaric processes the following constant is defined as the Gibbs function:

$$u1 + e_s \alpha 1 - T\phi_1 = u2 + e_s \alpha 2 - T\phi_2$$

and in general the Gibbs differential is:

(41) $dG = du + e_s d\alpha + \alpha de_s - T d\phi - \phi dT$

but $dq = du + e_s d\alpha = T d\phi$, therefore Eq. (41) is reduced to

(41) $dG = du + e_s d\alpha + \alpha de_s - Td\phi - \phi dT$

but $dq = du + e_s d\alpha = Td\phi$, therefore Eq. (41) is reduced to

(42) $dG = \alpha de_s - \phi dT$

but G remains constant in the transition between the two phases

(43) $\frac{de_s}{dT} = \frac{\phi 2 - \phi 1}{\alpha 2 - \alpha 1} = \frac{L}{T(\alpha 2 - \alpha 1)}$

Since $\alpha 2 \ll \alpha 1$

(43) $\frac{de_s}{dT} = \frac{L}{T\alpha 2} = \frac{Le_s}{R_v T^2}$

If L was constant, Eq. (43) would be solved analytically: $e_s = Const * \exp(-\frac{L}{R_v T})$. But $L = L(T)$ and therefore should be solved per T range. L as a good approximation (Bolton) can be expressed as:

(44) $e_s = 100 * 6.112\exp(\frac{17.67T(in\ C!!!)}{T(in\ C!!!)+243.5})$ $\quad e_s\ in\ pascal$

$$(43)\ \frac{de_s}{dT} \cong \frac{L}{T\alpha_2} = \frac{Le_s}{R_v T^2}$$

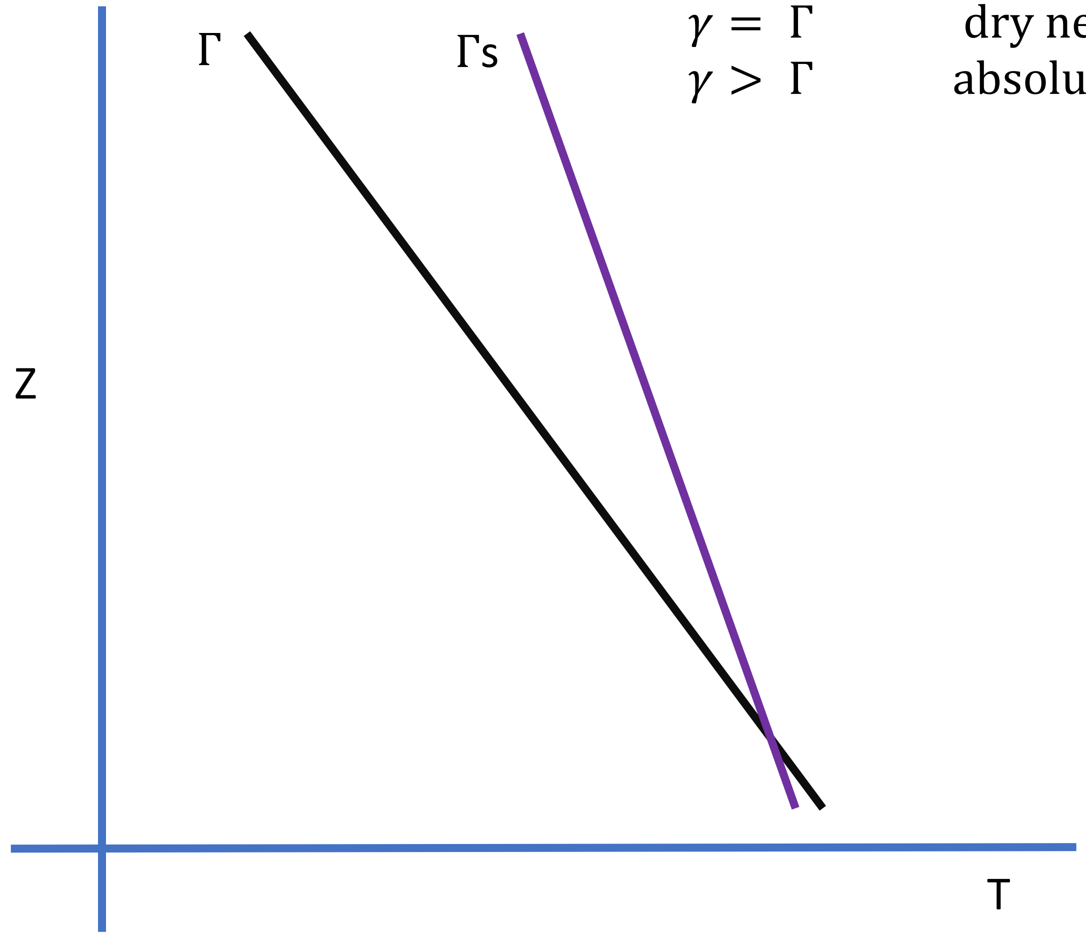
α_2 is the specific volume of wv that is in equilibrium with liquid water at temperature T.

$$e_s \cong e_s(T_0)e^{\frac{L}{R_v}\left(\frac{1}{T_0} - \frac{1}{T}\right)}$$

Stability and parcel perturbations

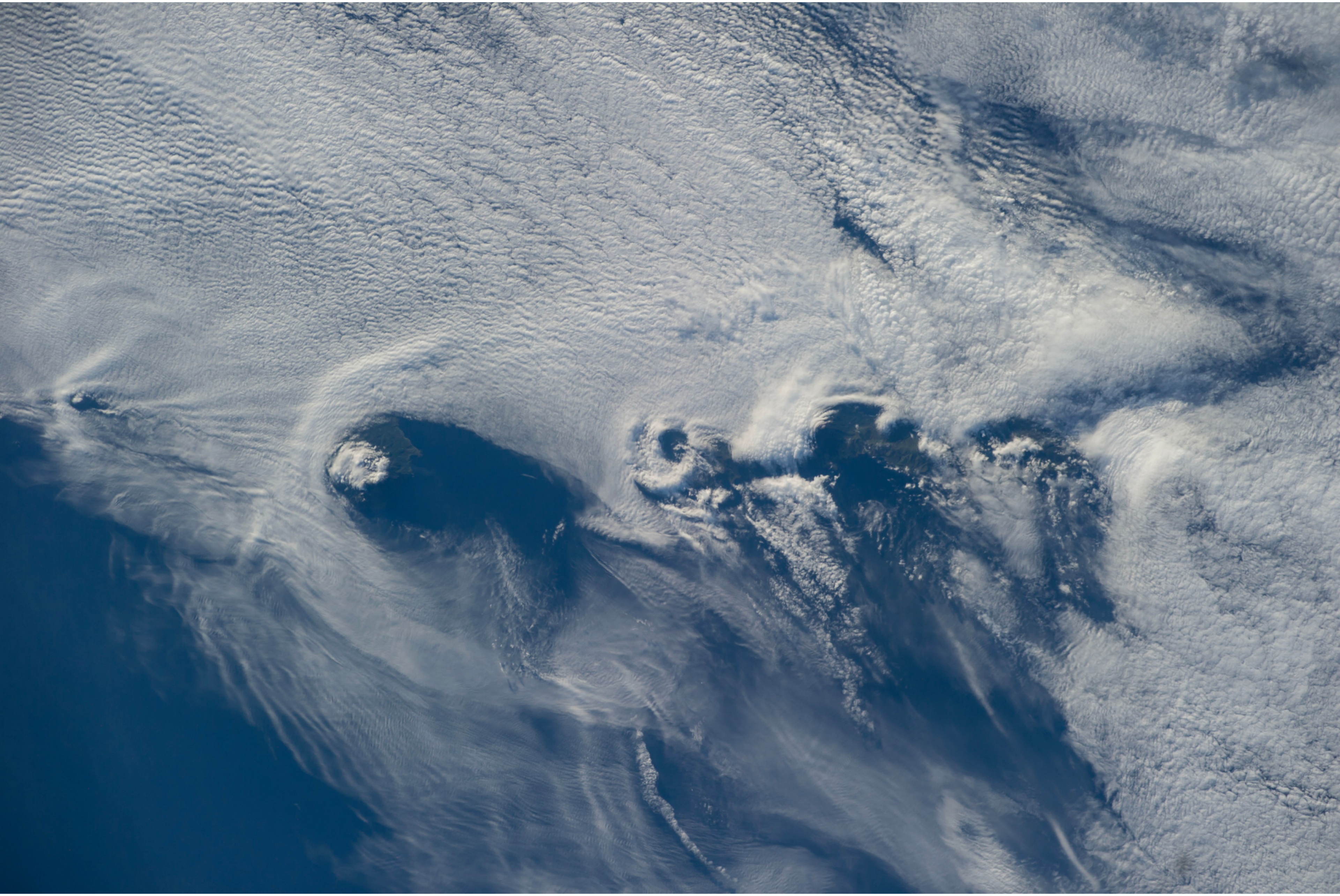
Parcel lapse rate Γ
Ambient lapse rate γ

$\gamma < \Gamma_s$	absolutly stable
$\gamma = \Gamma$	saturated neutral
$\Gamma_s < \gamma < \Gamma$	conditionaly unstable
$\gamma = \Gamma$	dry neutral
$\gamma > \Gamma$	absolutly unstable



3 possible parcel perturbations

V orographic



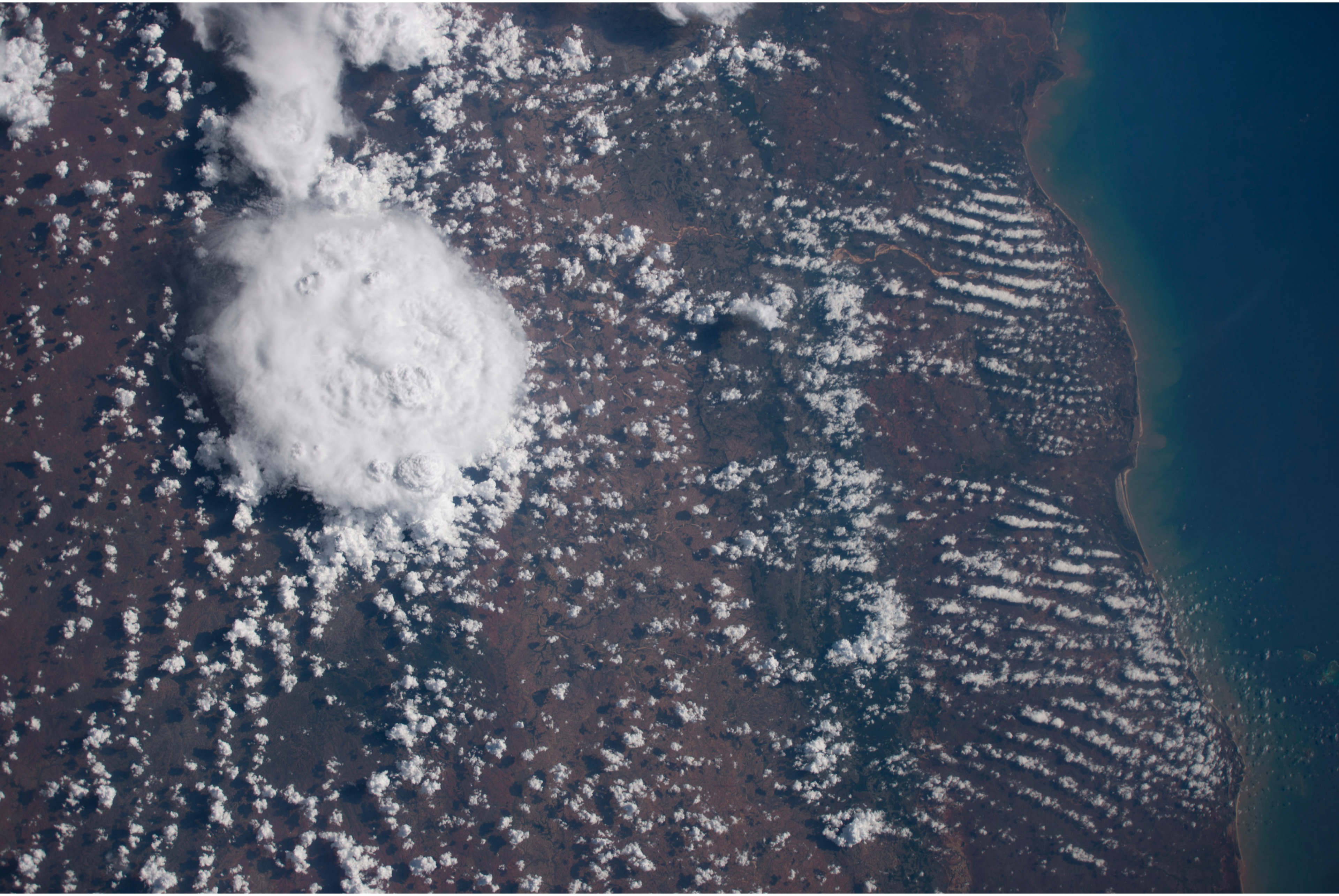
ISS Astronauts Photograph Earth - iss040e138991

ISS Astronauts Photograph Earth - Iss040e139

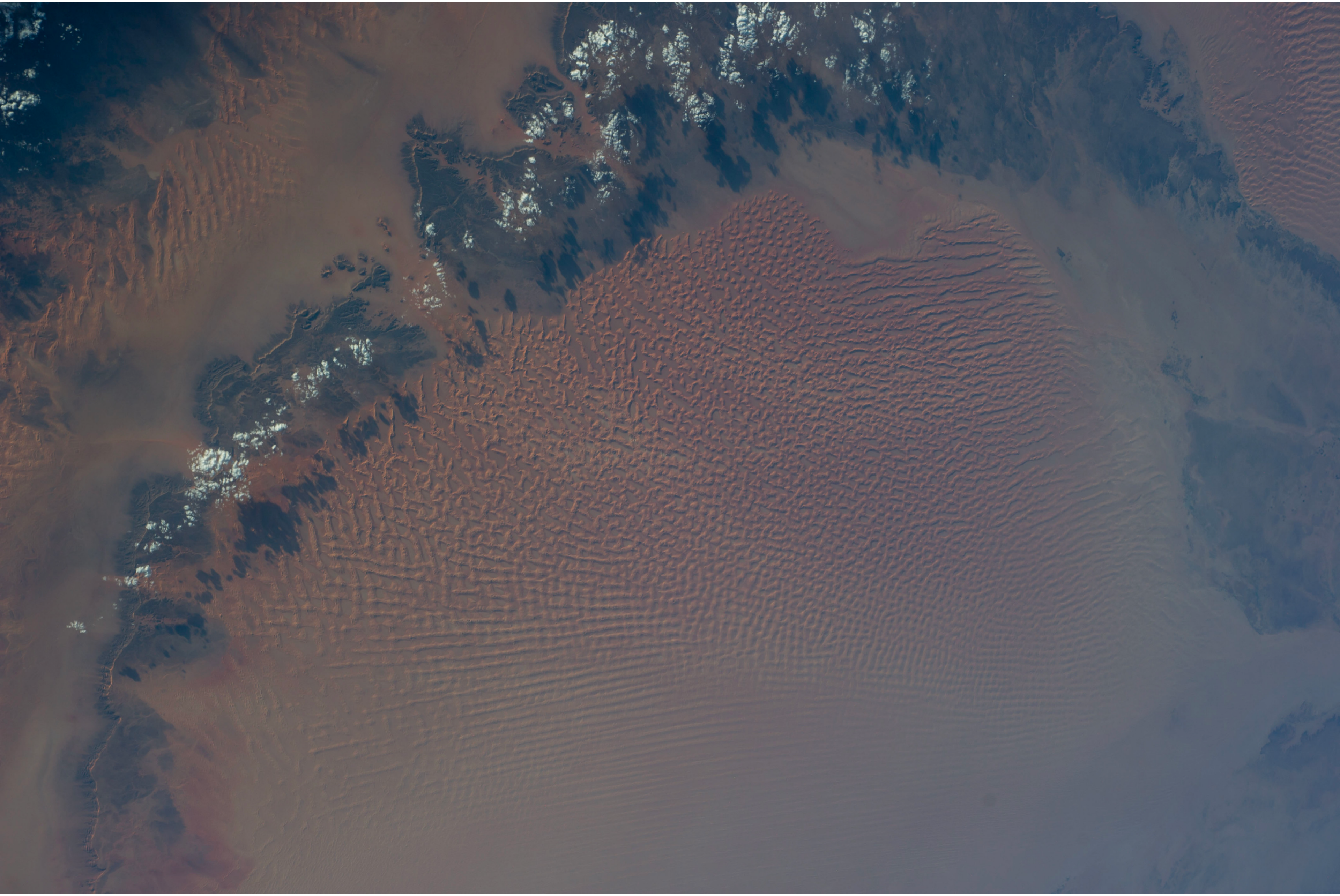
3 possible parcel perturbations

V orographic

T thermals



ISS Astronauts Photograph Earth - iss042e025076



ISS Astronauts Photograph Earth - iss040e123296

3 possible parcel perturbations

V orographic

T thermals

Q vapor pockets