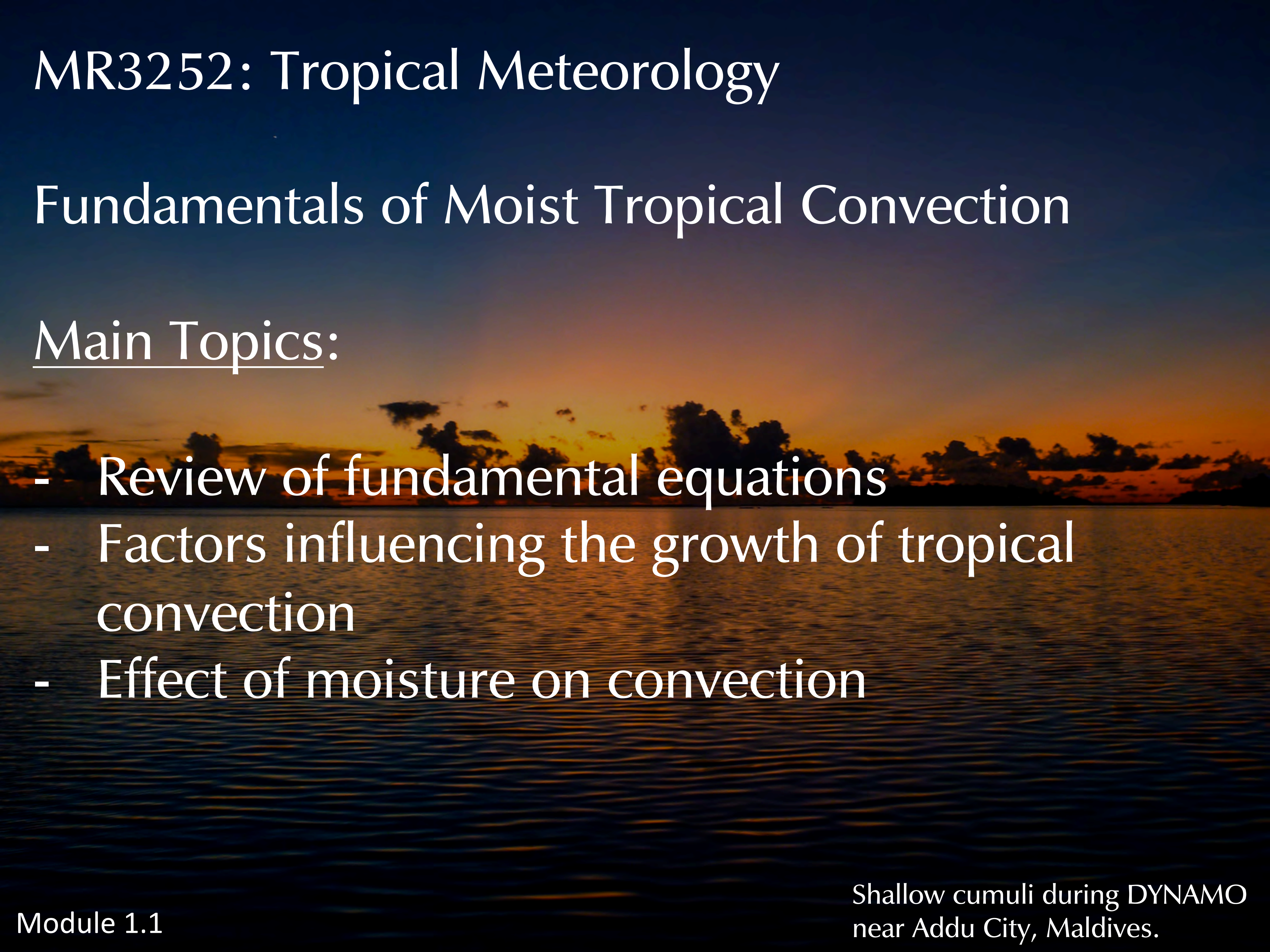
MR3252: Tropical Meteorology

Fundamentals of Moist Tropical Convection



Shallow cumuli during DYNAMO near Addu City, Maldives.

Main Topics:

- Review of fundamental equations
- Factors influencing the growth of tropical convection
- Effect of moisture on convection

Module 1.1

In mid-latitude dynamics, **quasi-geostrophic** theory is largely used to explain synoptic-scale motions.

Start with the basic primitive equations in isobaric coordinates, neglecting friction (from Holton, Chapter 3):

Horizontal Momentum

$$\frac{D\mathbf{V}}{Dt} + f\mathbf{k}\times\mathbf{V} = -\nabla_p \phi$$

Mass Continuity

$$\nabla\cdot\mathbf{u} = 0 \quad \longrightarrow \quad \frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial \omega}{\partial p} = 0$$

Thermodynamic

$$\frac{D_h T}{Dt} - S_p\omega = \frac{J}{c_p} = Q$$

Key to QG theory is the assumption that, to first order, motions are in **geostrophic balance.**

<u>Why does this not work in the Tropics?</u>

$$\frac{D\mathbf{V}}{Dt} + f\mathbf{k} \times \mathbf{V} = -\frac{1}{\rho}\nabla_p \phi$$

Key to QG theory is the assumption that, to first order, motions are in **geostrophic balance.**

Why does this not work in the Tropics? $f \approx 0$.
Only for planetary length scales in the Rossby number small enough (e.g. << 1) to be able to assume geostrophic balance.

On spatial scales of O(1000km) there cannot be a balance between pressure gradient force and Coriolis in the Tropics.

All three terms above are similar enough in magnitude that none (including the Lagrangian of the wind vector) can be neglected.

Also consider the thermodynamic equation:

$$\frac{D_h T}{Dt} - S_p \omega = Q$$

In QG theory, diabatic heating (Q) can be ignored to estimate many large-scale mid-latitude motions. This leaves the time tendency of temperature affected by advection and vertical motion.

In the Tropics, moist convection is central to dynamics!

Moist heating drives vertical motion, but at the same time, an increase in upward motion (usually realized as a decrease in clear-air downward motion) can promote diabatic heating.

$$\frac{\partial T}{\partial t} + u\cancelto{0}{\frac{\partial T}{\partial x}} + v\cancelto{0}{\frac{\partial T}{\partial y}} = Q + S_p \omega$$

The feedback and interaction between convection and larger-scale tropical dynamics remains an open topic of research today.

What factors influence moist tropical convection? *Many, and the non-linear interactions between them are not fully understood!*

Leading factors impacting convection **(by altering buoyancy of convective updrafts**)**:**

1) Moisture availability (A tautology to say that moist convection needs moisture, really. Of course, one needs moisture to support moist convection!)

2) Static stability

3) Surface fluxes

4) Wind shear

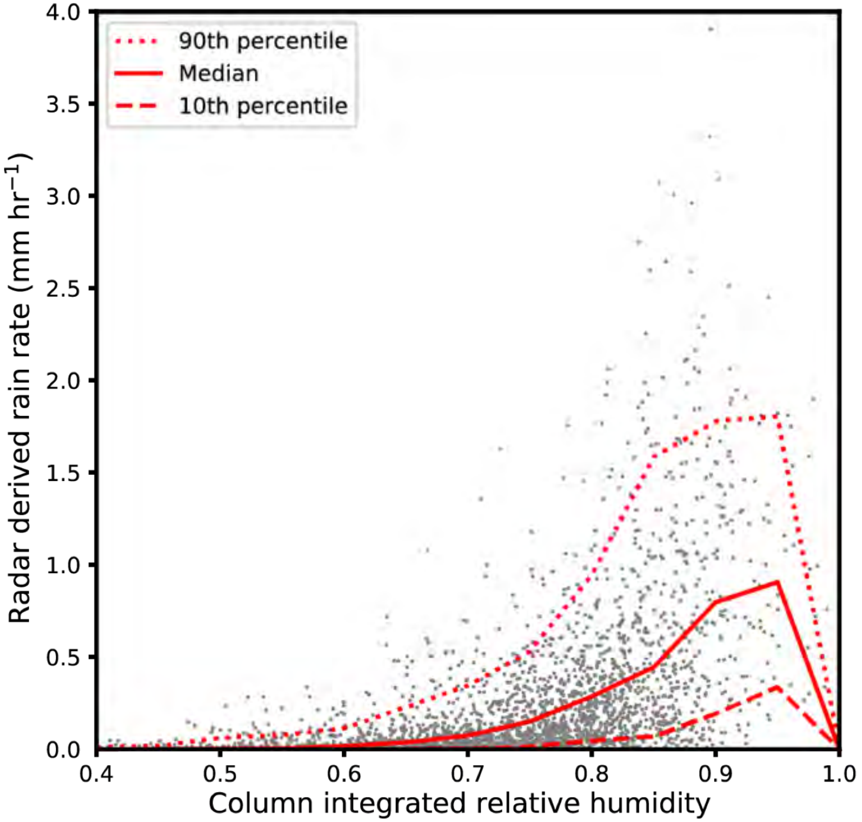
5) Low-level convergence (provides forcing)

However, these are impacted by other factors that are part of the large-scale dynamics that are themselves influenced by the convection:

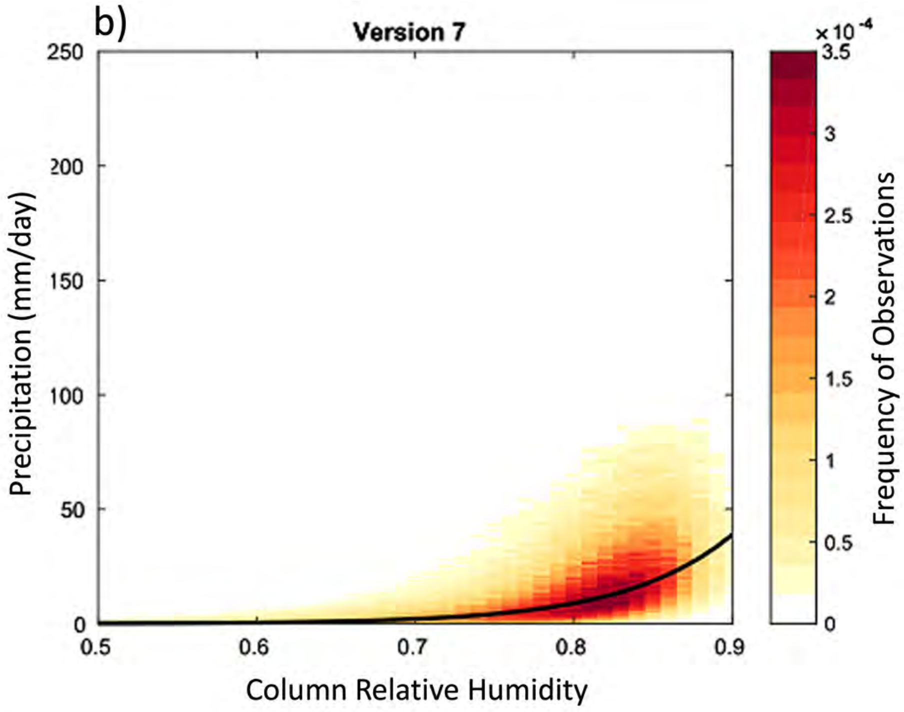
- Advection
- Near-surface wind
- Sea surface temperature
- Adiabatically driven vertical motion
- Vertical flux of moisture (i.e. moist convection itself)
- Low-level convergence

Moisture Availability

Radar Data (Powell 2019)



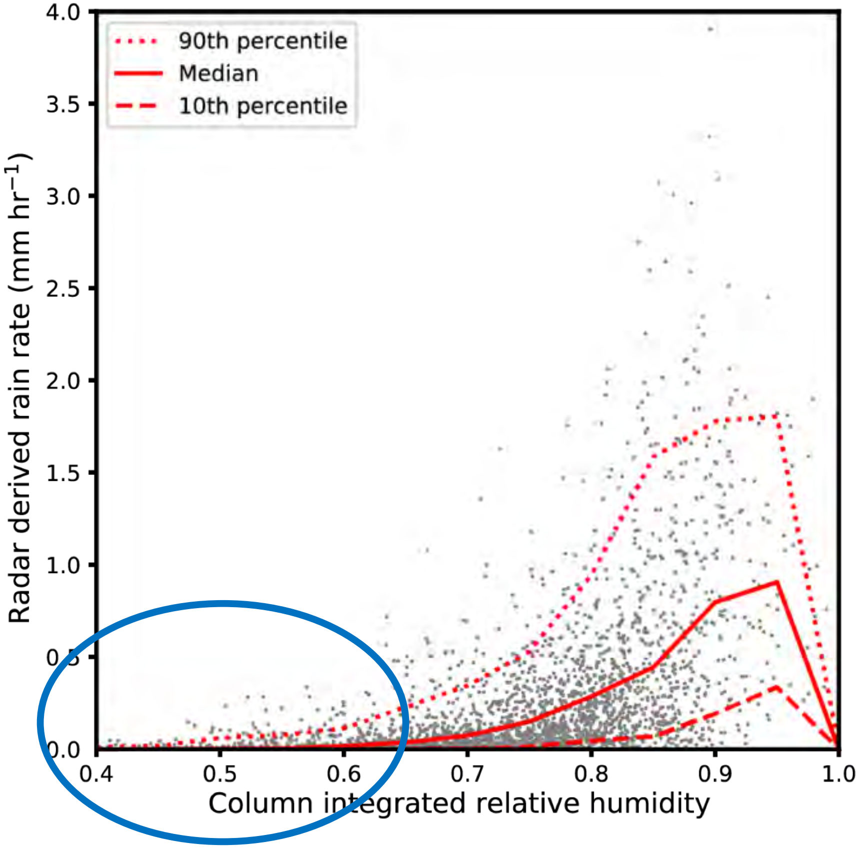
Satellite Data (Rushley et al. 2018)



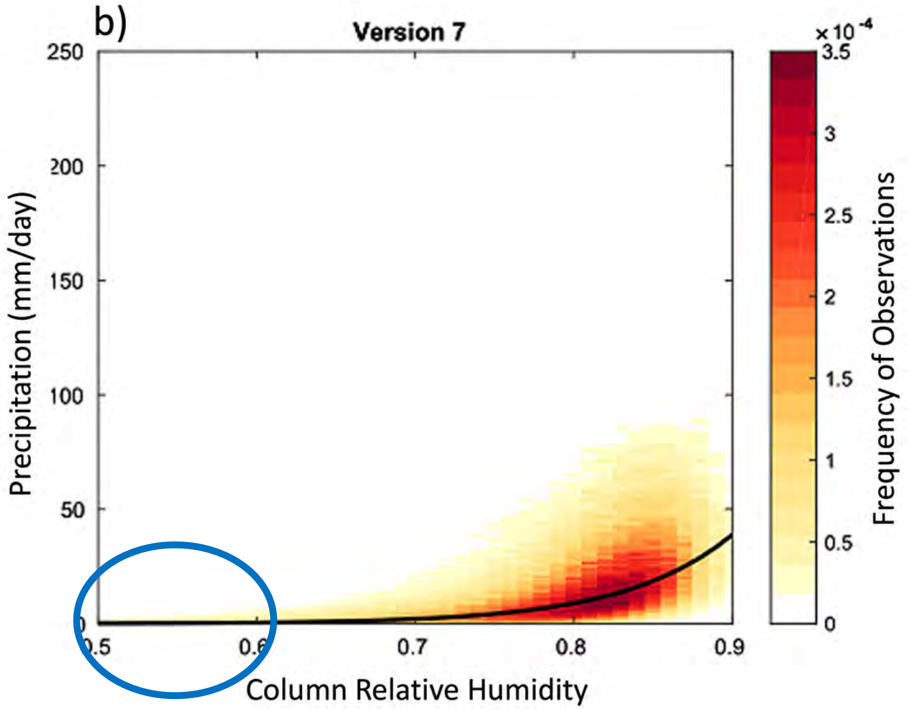
$$CRH = \frac{\int_{P_{sfc}}^{P_{top}} q \, dP}{\int_{P_{sfc}}^{P_{top}} q_{sat}(T) \, dP}$$

Moisture Availability

Radar Data (Powell 2019)



Satellite Data (Rushley et al. 2018)

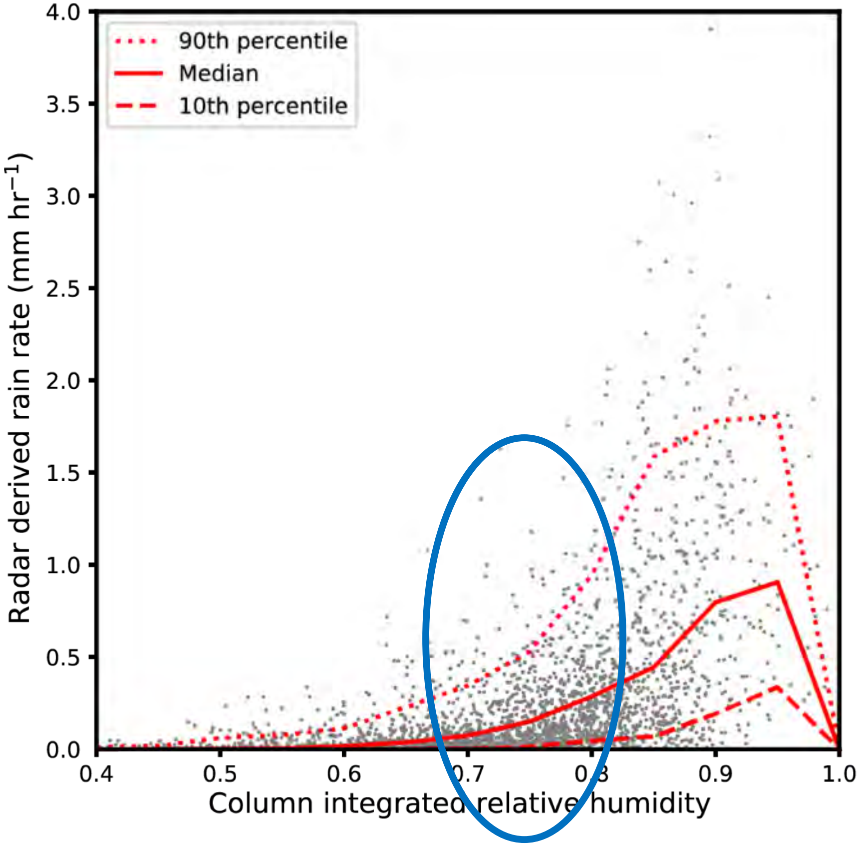


$$CRH = \frac{\int_{P_{sfc}}^{P_{top}} q\, dP}{\int_{P_{sfc}}^{P_{top}} q_{sat}(T)\, dP}$$

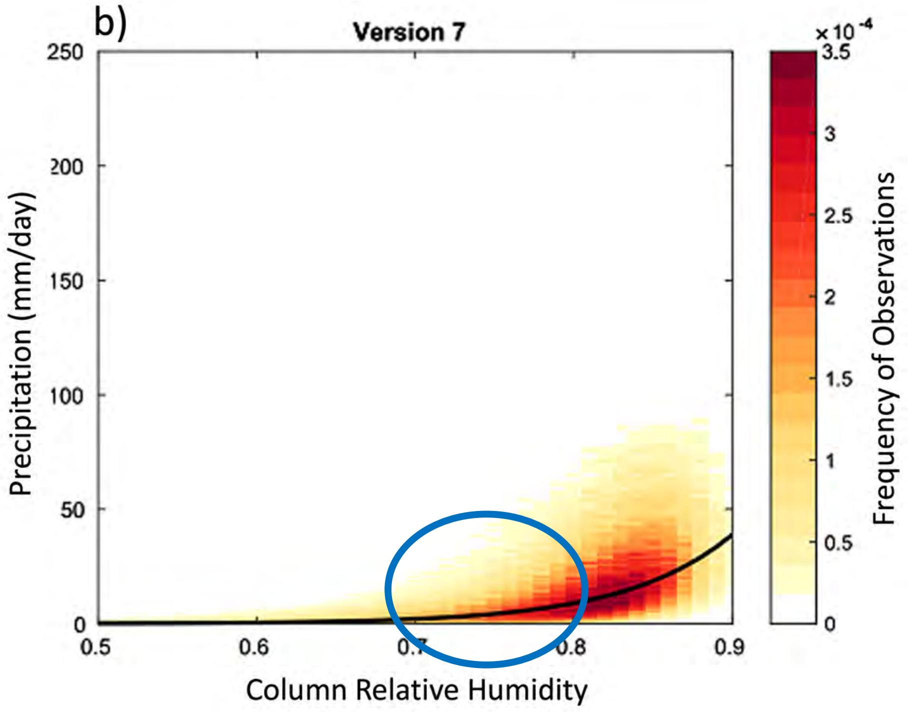
Low rain rates at low CRH.

Moisture Availability

Radar Data (Powell 2019)



Satellite Data (Rushley et al. 2018)

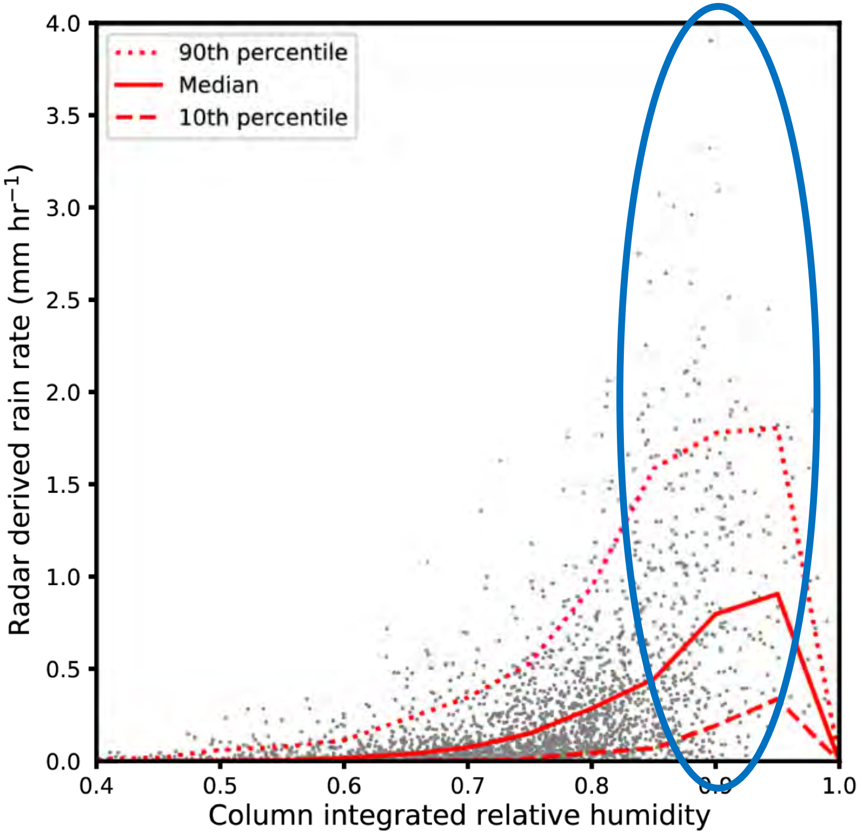


$$CRH = \frac{\int_{P_{sfc}}^{P_{top}} q \, dP}{\int_{P_{sfc}}^{P_{top}} q_{sat}(T) \, dP}$$

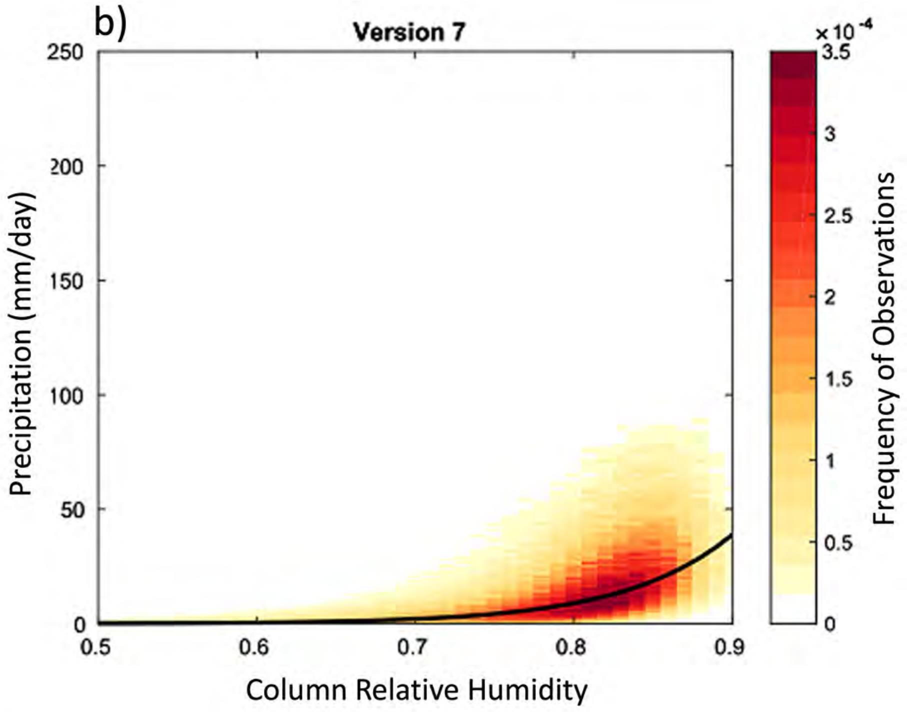
Exponential increase begins to increase quite rapidly between 70 and 80% CRH.

Moisture Availability

Radar Data (Powell 2019)

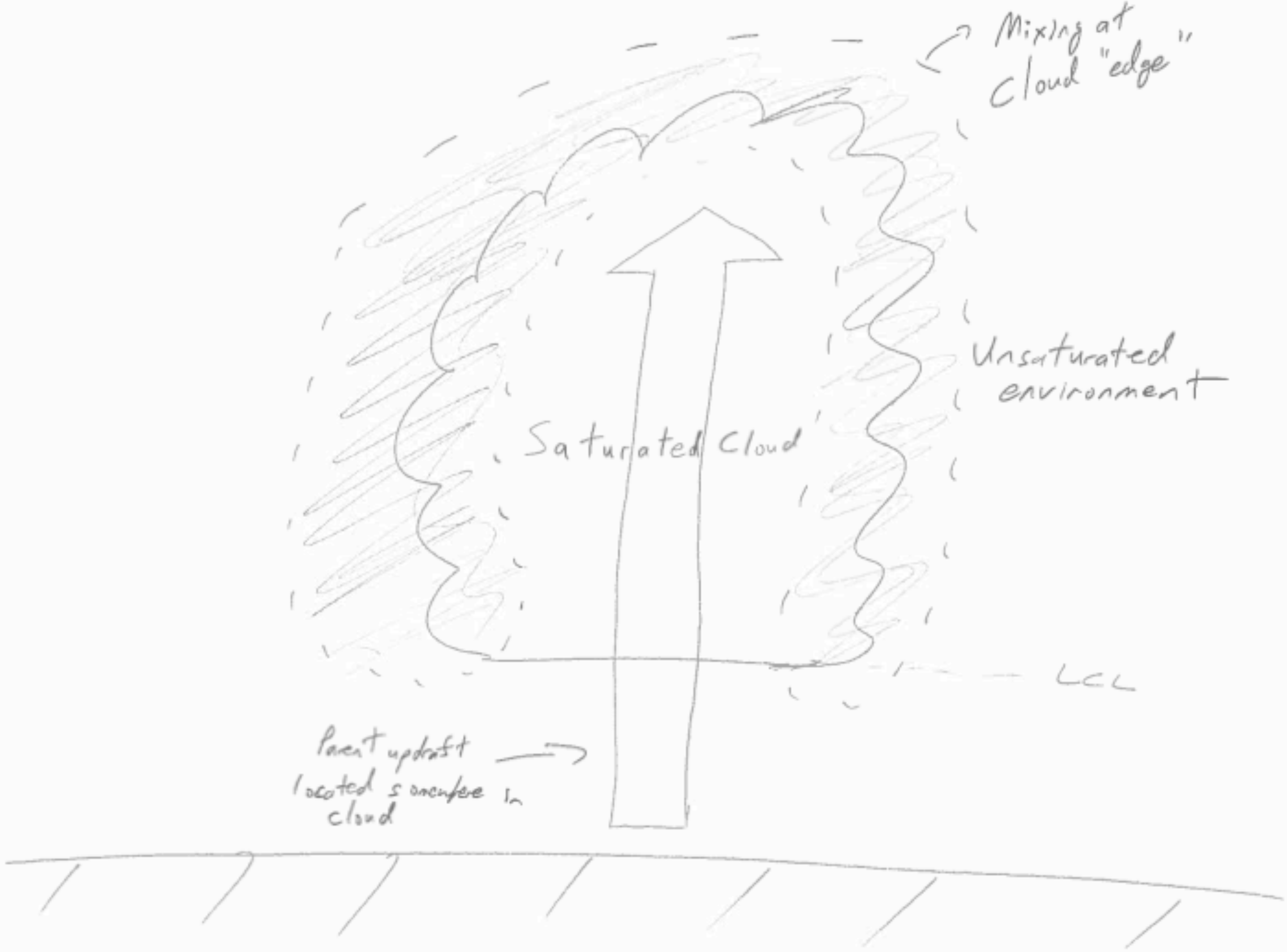


Satellite Data (Rushley et al. 2018)



$$CRH = \frac{\int_{P_{sfc}}^{P_{top}} q \, dP}{\int_{P_{sfc}}^{P_{top}} q_{sat}(T) \, dP}$$

But lots of scatter in rain rate at high CRH. Why?


Mixing at cloud "edge"
Unsaturated environment
Saturated Cloud
LCL
Parent updraft located somewhere in cloud

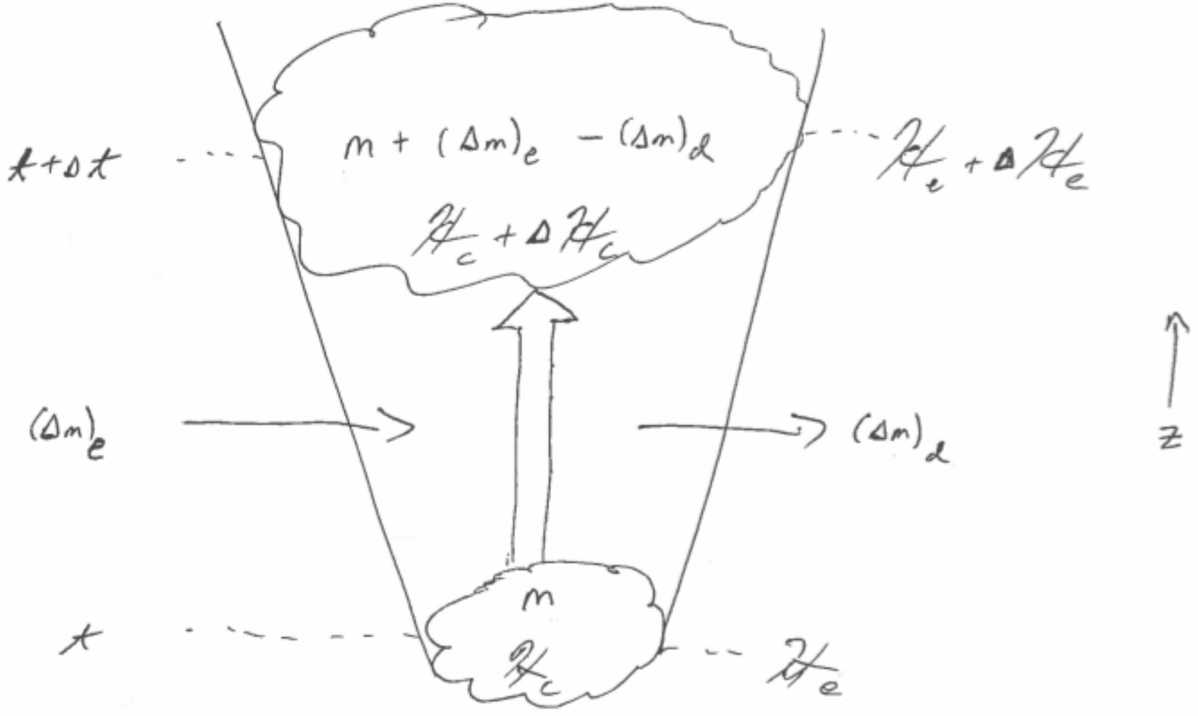
Why does moisture impact deep convection? *Via **entrainment**.*

Definition of entrainment, from AMS Glossary:

> In meteorology, the mixing of environmental air into a preexisting organized air current so that the environmental air becomes part of the current; the opposite of detrainment.
>
> Entrainment of air into clouds, especially cumulus, is said to be inhomogeneous when the timescale for mixing of environmental air is very much greater than the timescale for drop evaporation. Under these conditions, which are often found when environmental air is first entrained into cumulus, regions of cloud and entrained air are intertwined, with evaporation occurring only on the edges of the interface between the cloudy and entrained environmental air.

Mathematically, consider the case of continuous, homogenous entrainment. This presumes that entrainment is an instant process that is experienced throughout a cloud in the same way.

See Houze (2014), Fig. 7.3
Stommel (1947)



$$[m + (\Delta m)_e - (\Delta m)_d](\mathcal{H}_c + \Delta\mathcal{H}_c) = m\mathcal{H}_c + (\Delta m)_e\mathcal{H}_e - (\Delta m)_d\mathcal{H}_c + \overbrace{\left(\frac{\Delta\mathcal{H}_c}{\Delta t}\right)_s}^{\text{Source term}} m\Delta t$$

$$[m + (\Delta m)_e - (\Delta m)_d](\mathcal{H}_c + \Delta\mathcal{H}_c) = m\mathcal{H}_c + (\Delta m)_e\mathcal{H}_e - (\Delta m)_d\mathcal{H}_c + \overbrace{\left(\frac{\Delta\mathcal{H}_c}{\Delta t}\right)_s}^{\text{Source term}} m\Delta t$$

Take limit for

$\Delta t \rightarrow 0$

$$\frac{D\mathcal{H}_c}{Dt} = \left(\frac{D\mathcal{H}_c}{Dt}\right)_s + \frac{1}{m}\left(\frac{Dm}{Dt}\right)_e (\mathcal{H}_e - \mathcal{H}_c)$$

What happens if we let $\mathcal{H}$ represent moist enthalpy ($h = c_p T + L_v q$)?

$$\frac{Dh_c}{Dt} = M + \frac{1}{m}\left(\frac{Dm}{Dt}\right)_e (h_e - h_c)$$

Source/sink due to phase changes (mainly evaporation minus condensation)

Difference in temperature and/or environmental vapor and in-cloud vapor concentrations.

Why is dry air detrimental to deep, moist convection?

$$\frac{Dh_c}{Dt} = M + \frac{1}{m}\left(\frac{Dm}{Dt}\right)_e (h_e - h_c)$$

If mass is entrained into cloud, then the 2nd term means that the moist enthalpy in the cloud decreases.

Difference in environmental and in-cloud temperature/humidity.

Consider equation for buoyancy (B) (here prime means deviation from hydrostatically balanced state):

$$B = -g\frac{\rho'}{\rho_0}$$

And the 3-D momentum equation neglecting Coriolis and friction:

$$\frac{\partial \mathbf{v}}{\partial t} = -\frac{1}{\rho_0}\nabla p' + B\mathbf{k} - \mathbf{v}\cdot\nabla\mathbf{v}$$

Changing $q_{v,c}$ alters ρ'. Water vapor (H_2O) is lighter than dry air (N_2 and O_2). **An decrease in water vapor increases $\boldsymbol{\rho'}$ and inhibits upward acceleration.**

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Other Thermodynamic Factors Impacting Moist Convection

Main Topics:

- Effects on convection by static stability
- Global distribution of lightning

Module 1.2

What factors influence moist tropical convection? *Many, and the non-linear interactions between them are not fully understood!*

Leading factors impacting convection **(by altering buoyancy of convective updrafts)**:

1) Moisture availability (A tautology to say that moist convection needs moisture, really. Of course, one needs moisture to support moist convection!)

2) Static stability

3) Surface fluxes

4) Wind shear (may impact buoyancy and affects structure of convection)

5) Low-level convergence (provides forcing)

However, these are impacted by other factors that are part of the large-scale dynamics that are themselves influenced by the convection:

- Advection
- Near-surface wind
- Sea surface temperature
- Adiabatically driven vertical motion
- Vertical flux of moisture (i.e. moist convection itself)
- Low-level convergence

Static Stability appears as S_p in the thermodynamic equation:

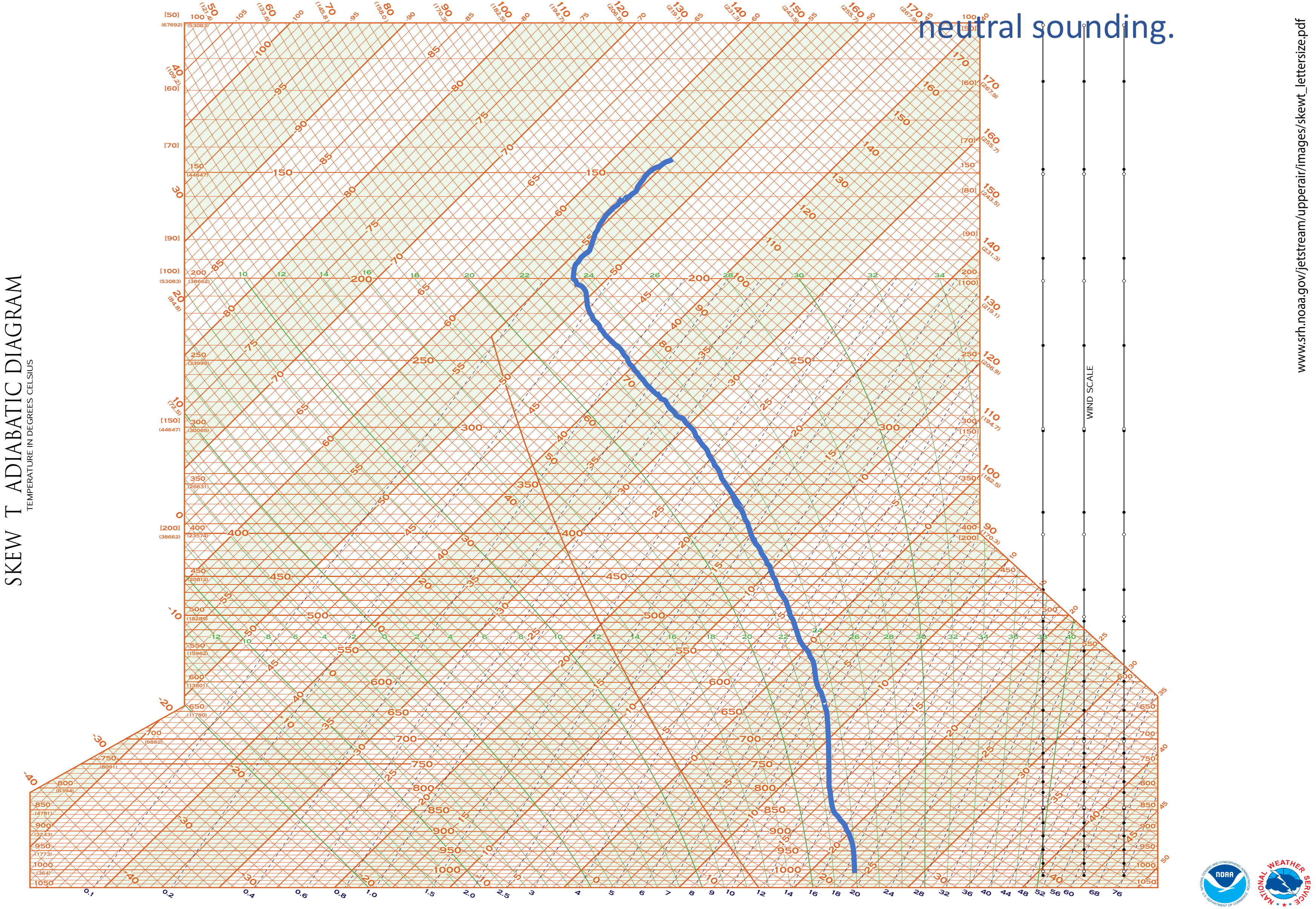
$$S_p = \frac{\Gamma_d - \Gamma}{\rho g} = -\frac{1}{c_p}\frac{\partial s}{\partial p}$$

$$\frac{T}{\theta}\frac{\partial \theta}{\partial z} = \Gamma_d - \Gamma$$

$$s = c_p T + \phi \quad \longleftarrow \quad \text{Dry static energy}$$

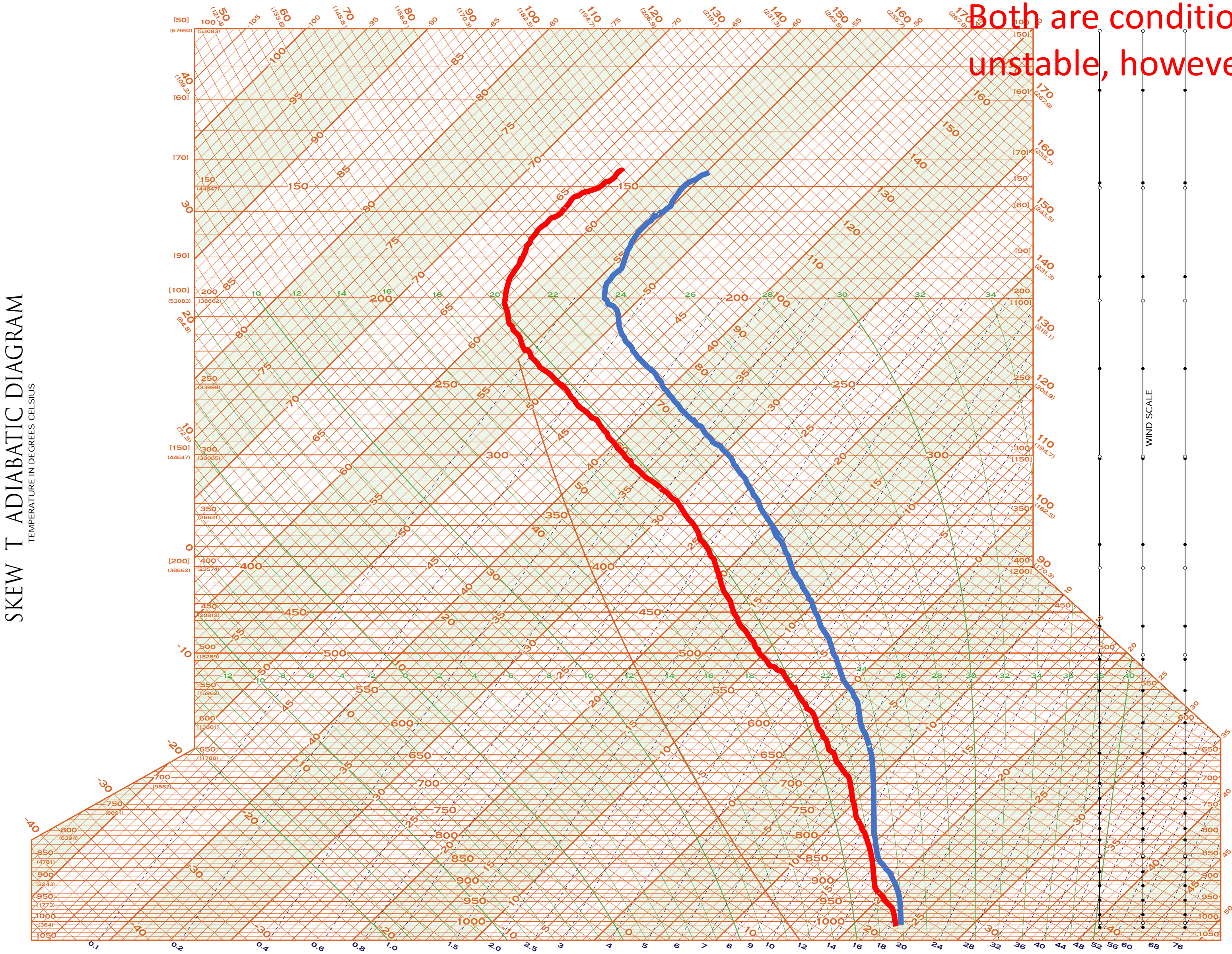
If the environmental lapse rate is large (i.e. atmosphere cools quickly with height), then air rising from the PBL is likely to be warmer (more buoyant) than the environment. A low lapse rate means the difference in environmental temperature and parcel temperature will be less (lower buoyancy; smaller vertical acceleration).

This is a conditionally unstable, but nearly moist neutral sounding.



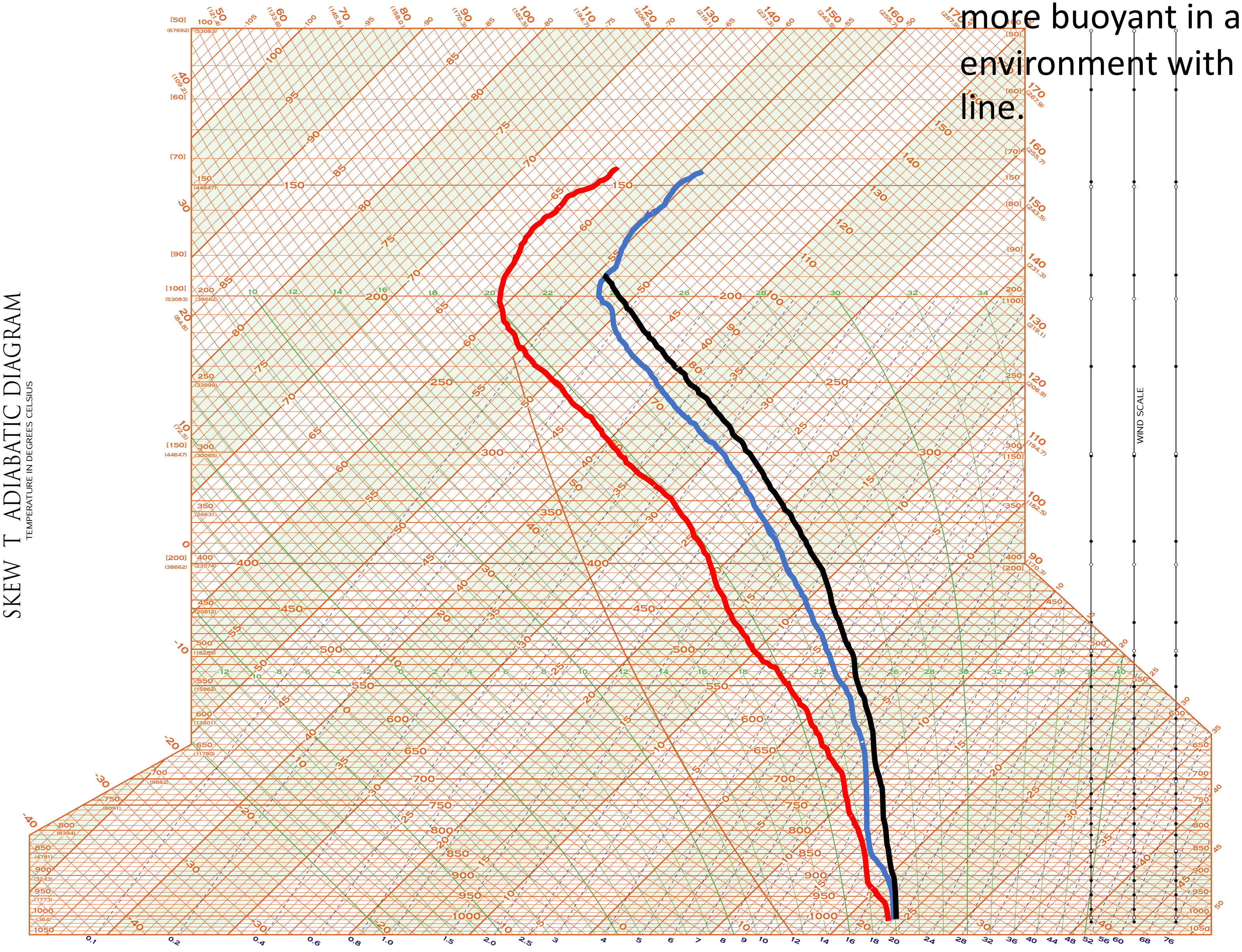
Red line is less stable than the blue line.

Both are conditionally unstable, however.



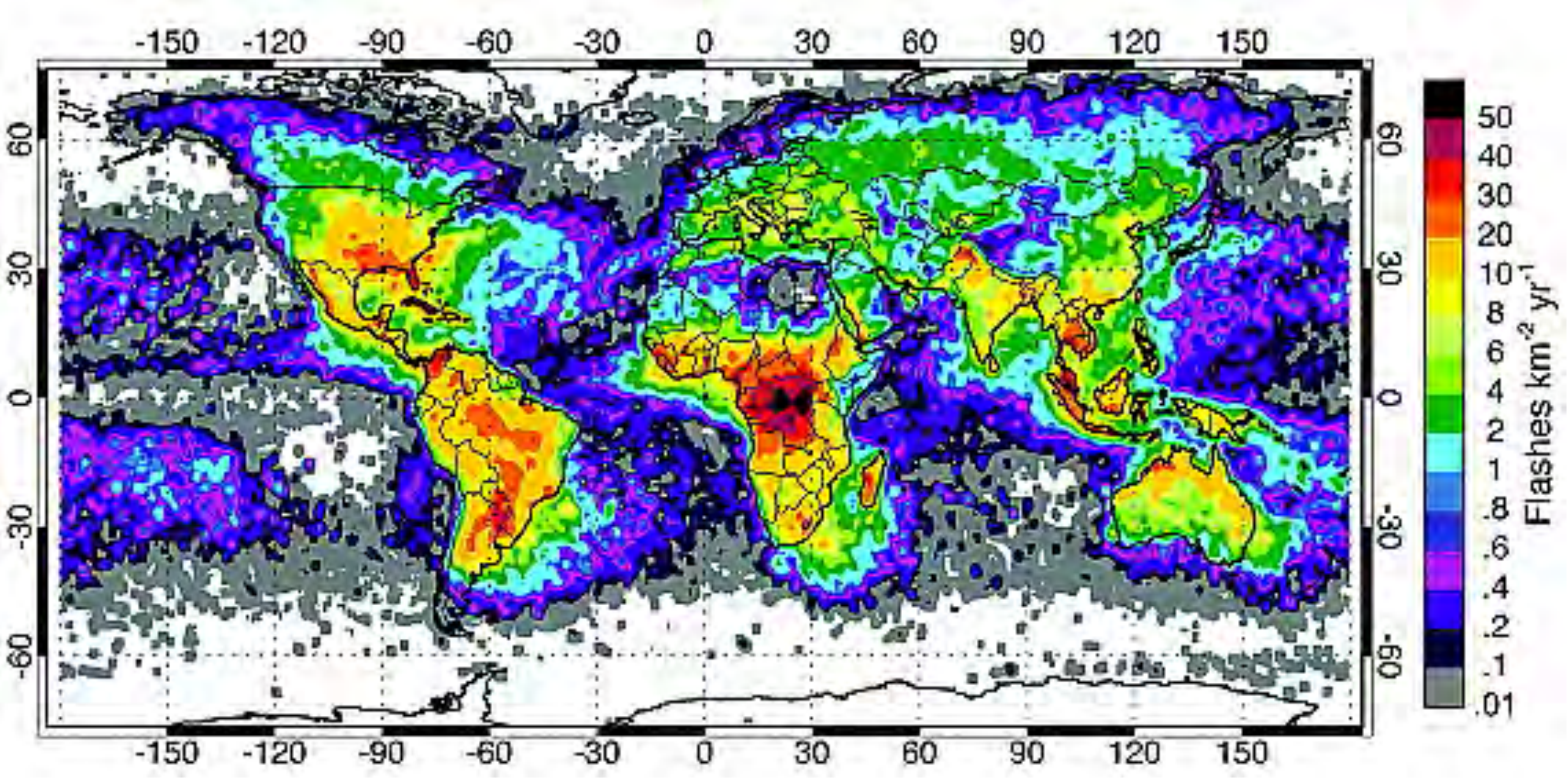
www.srh.noaa.gov/jetstream/upperair/images/skewt_lettersize.pdf

A parcel following the black line (a moist adiabat) will be more buoyant in an environment with the red line.



www.srh.noaa.gov/jetstream/upperair/images/skewt_lettersize.pdf

Global distribution of lightning: Far more common over land (Figure from Christian et al. 2003)



MR3252: Tropical Meteorology

Other Thermodynamic Factors Impacting Moist Convection

Main Topics:

- Effects on convection by surface fluxes
- Buoyancy flux
- Latent heat flux
- Sensible heat flux
- Global distribution of precipitation

Module 1.3

Surface Fluxes

Two types of surface fluxes: Latent and sensible heat fluxes

Latent heat flux: Heat flux between atmosphere and surface associated with phase change of water. For example, evaporation of ocean water into the atmosphere would be a positive flux of latent heat to the atmosphere.

Sensible heat flux: Conductive heat flux between surface and atmosphere. If the surface is warmer than the surface-layer of the atmosphere, then the sensible heat flux to the atmosphere will be positive.

$$L = \rho L_v C_q U(q_s - q)$$

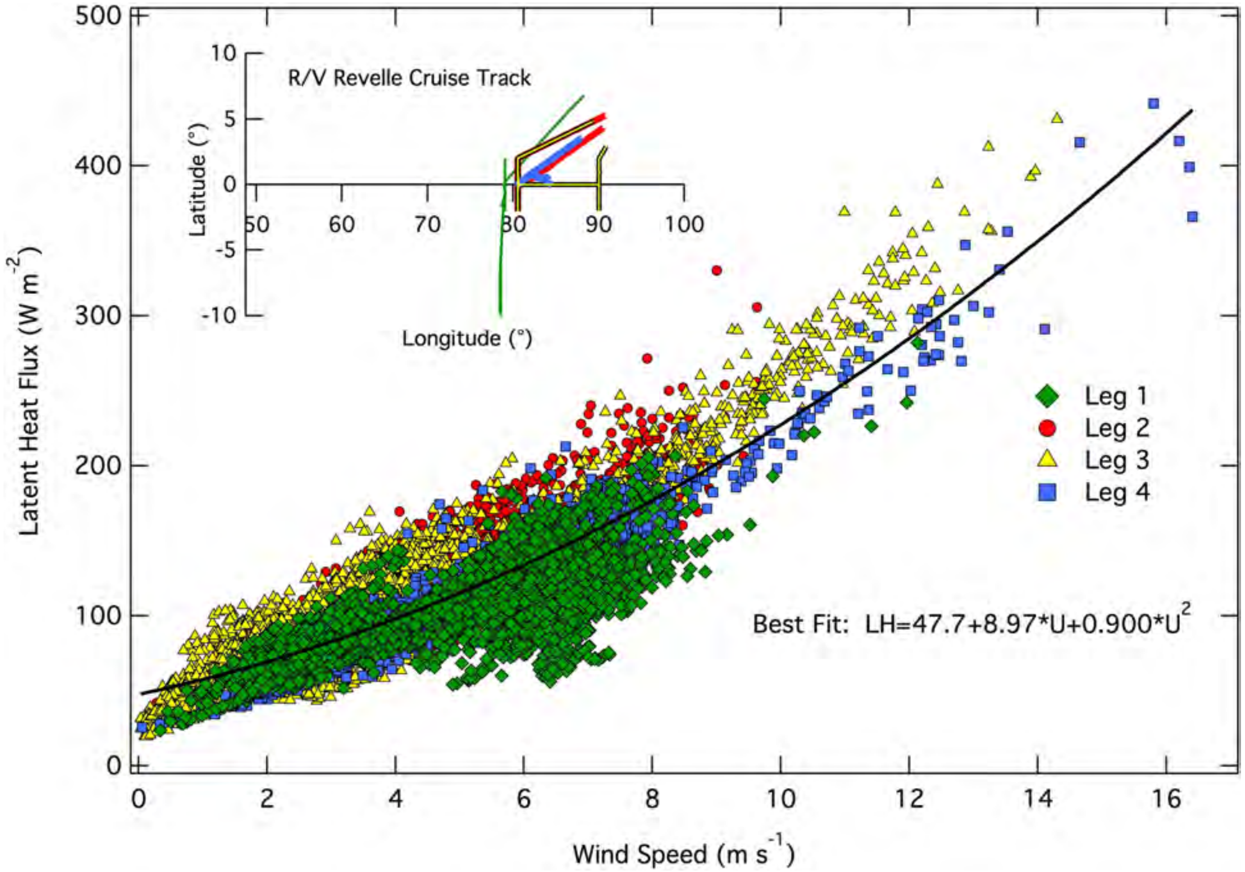
$$S = \rho c_p C_h U(\theta_s - \theta)$$

The *C* variables are exchange coefficients that are dependent on conditions at the air-sea interface.

Both fluxes are functions of wind, and the subscript *s* indicates the conditions of the ocean. q_s is the saturation specific humidity associated with air with the same temperature as the sea surface.

Surface Fluxes

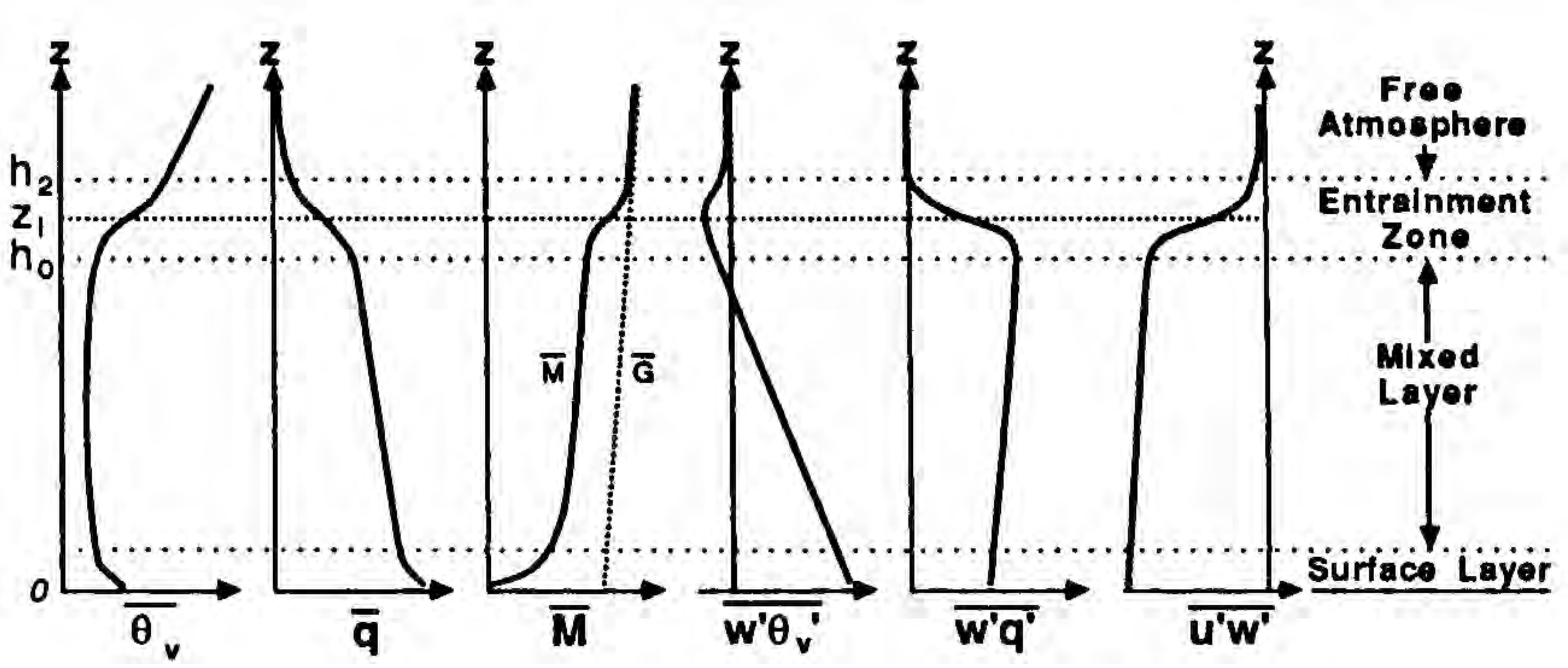
Qian et al.
(2016)



Surface latent heat fluxes vary non-linearly with near-surface wind speed. Wind will alter surface roughness, which impacts exchange coefficient. Therefore, impact of wind on surface fluxes is nonlinear!

Typical profiles of quantities in a convective boundary layer

$\theta_v \approx \theta(1 + 0.61r)$, *r* = mixing ratio



mean virtual potential temperature

mean specific humidity

mean wind speed

buoyancy flux

specific humidity flux

momentum flux

What is a buoyancy flux?

"The vertical kinematic flux of virtual potential temperature $\overline{w'\theta_v{}'}$, which when multiplied by the buoyancy parameter (g/T_v) yields a flux that is proportional to buoyancy. (AMS Glossary)

$$\text{Buoyancy Flux} = \frac{g}{T_v}\overline{w'\theta_v{}'}$$

Turbulent sensible heat flux:

$$LHF = \rho_0 L_v \overline{w'q_v{}'}$$

$$SHF = \rho_0 c_p \overline{w'\theta'}$$

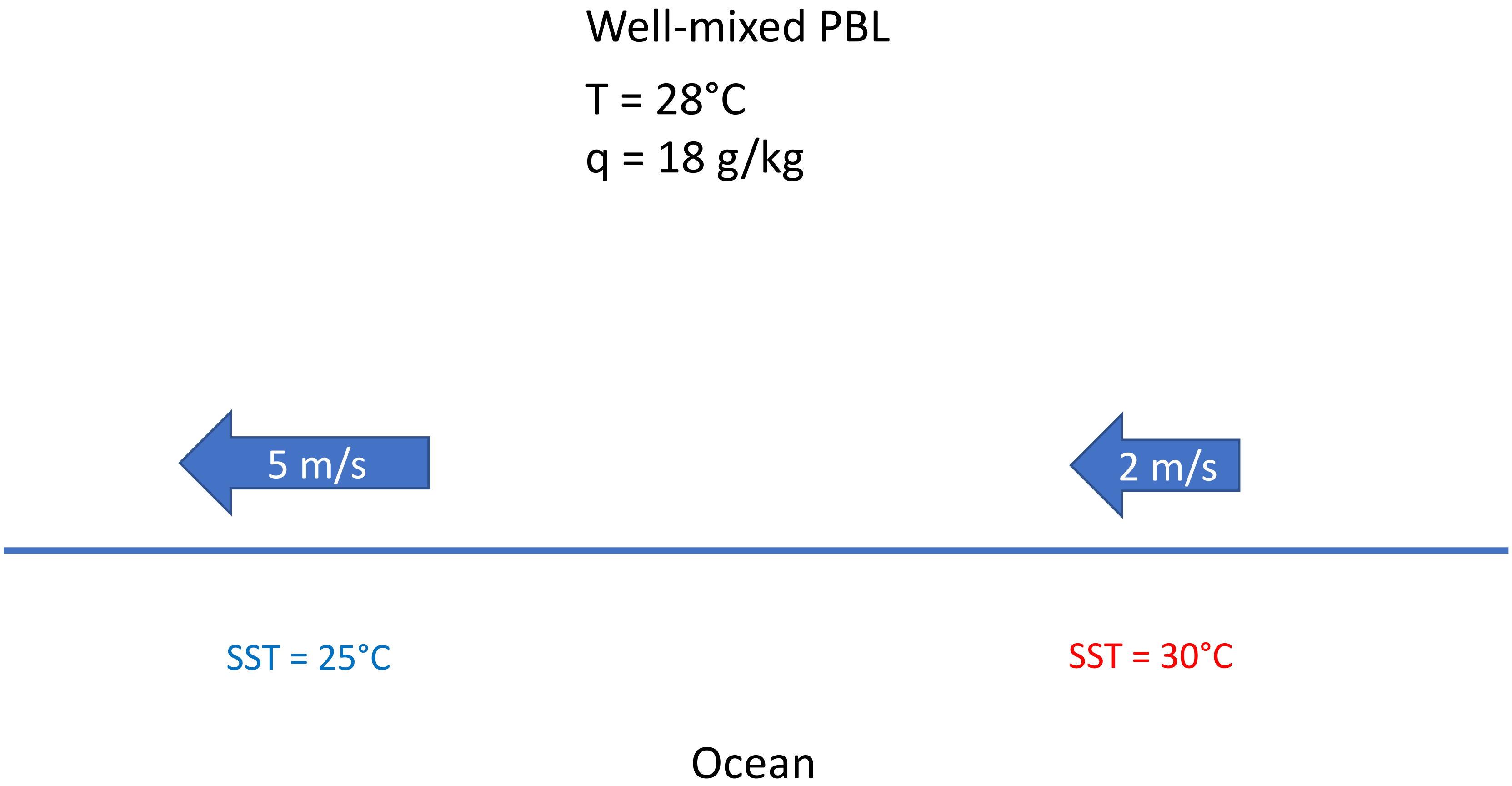
Take the equation for virtual potential temperature (ignoring hydrometeors, which we presume are not near the surface):

$$\theta_v \approx \theta(1 + 0.61r),\ r = \text{mixing ratio}$$

We eventually arrive at

$$\text{Buoyancy Flux} = \frac{g}{\rho c_p T}\left(SHF + \overbrace{0.61\frac{c_p T}{L_v}}^{\approx 0.08} LHF\right)$$

For typical temperatures, the contributions of SHF and LHF to buoyancy flux are similar in magnitude, even if LHF >> SHF.



Well-mixed PBL

T = 28°C

q = 18 g/kg

5 m/s

2 m/s

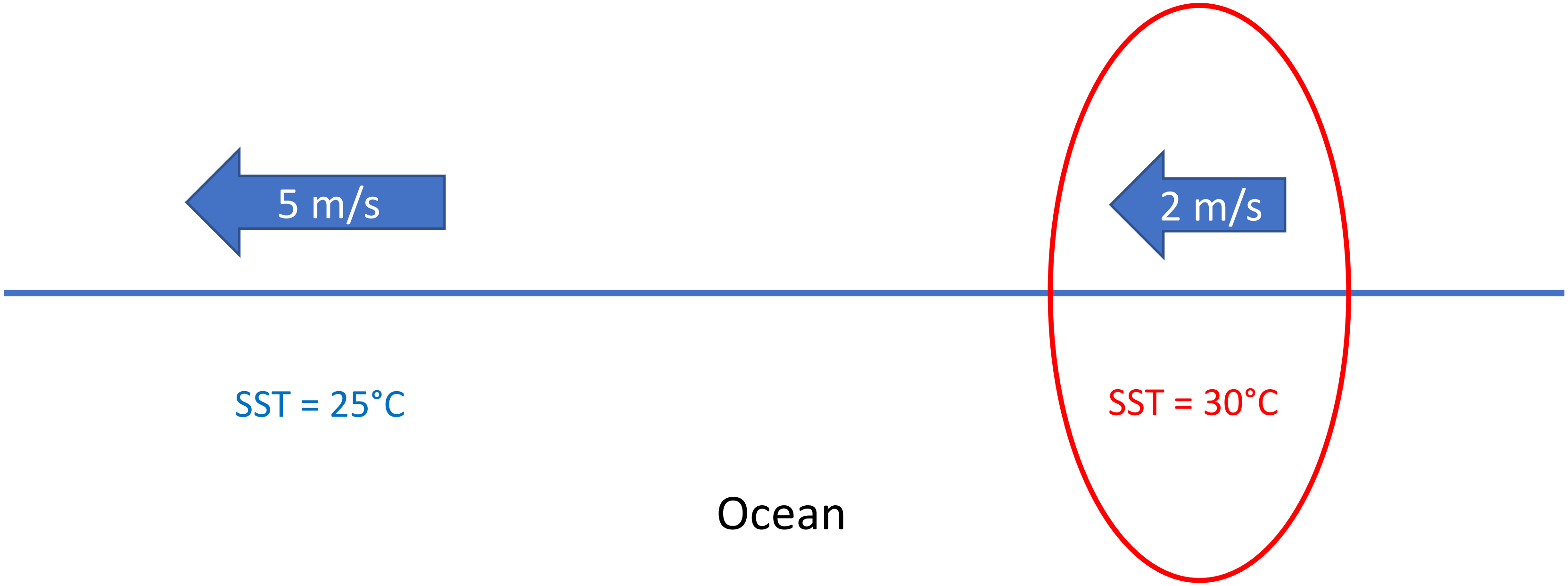
SST = 25°C

SST = 30°C

Ocean

Well-mixed PBL

T = 28°C
q = 18 g/kg



5 m/s

2 m/s

SST = 25°C

SST = 30°C

Ocean

Well-mixed PBL

T = 28°C

q = 18 g/kg

5 m/s

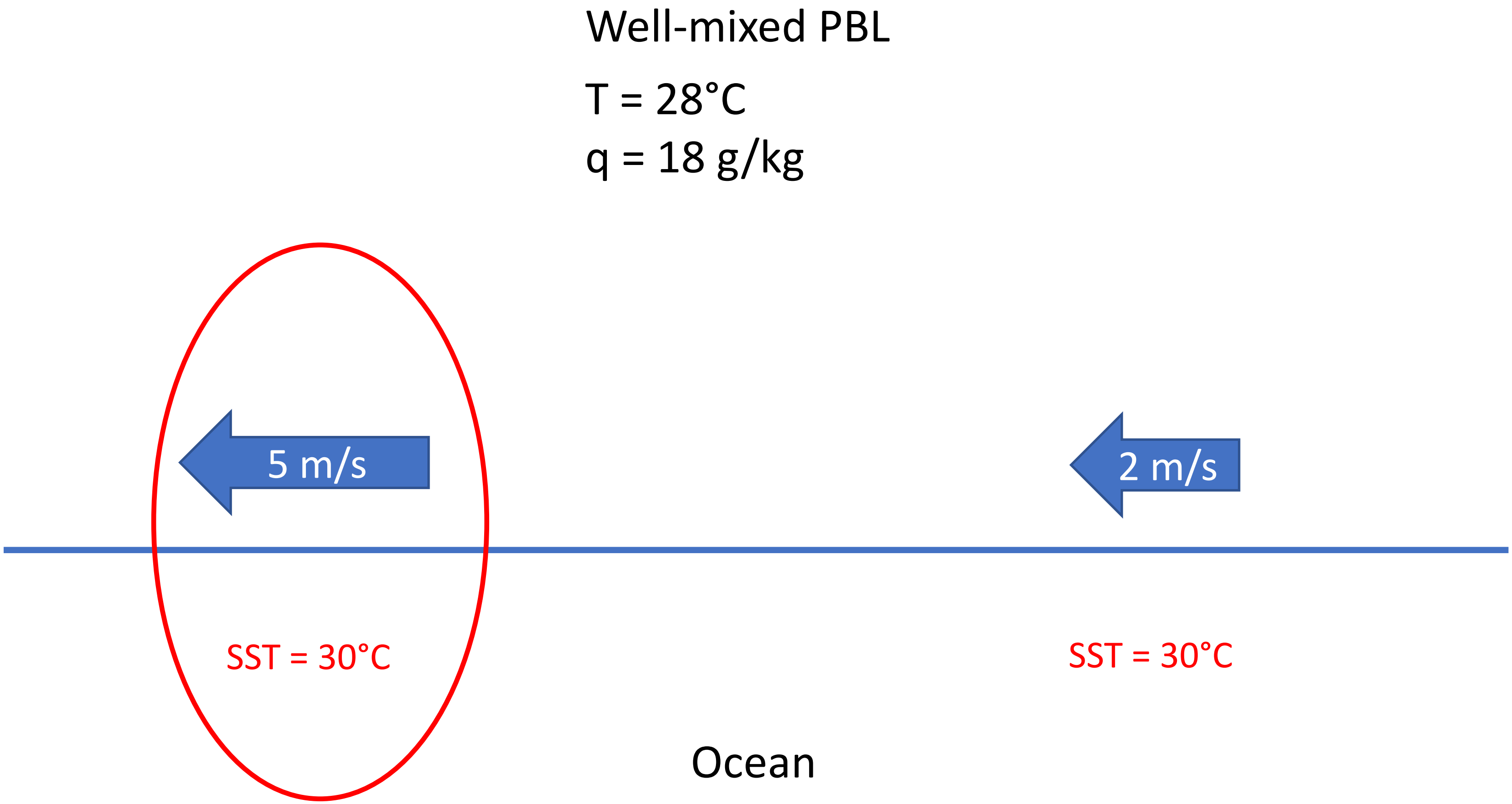
2 m/s

SST = 30°C

SST = 30°C

Ocean

Generally, high surface wind and high sea surface temperature will lead to an increase in boundary layer moist static energy (unless the PBL is already very warm and moist).



Well-mixed PBL

T = 28°C

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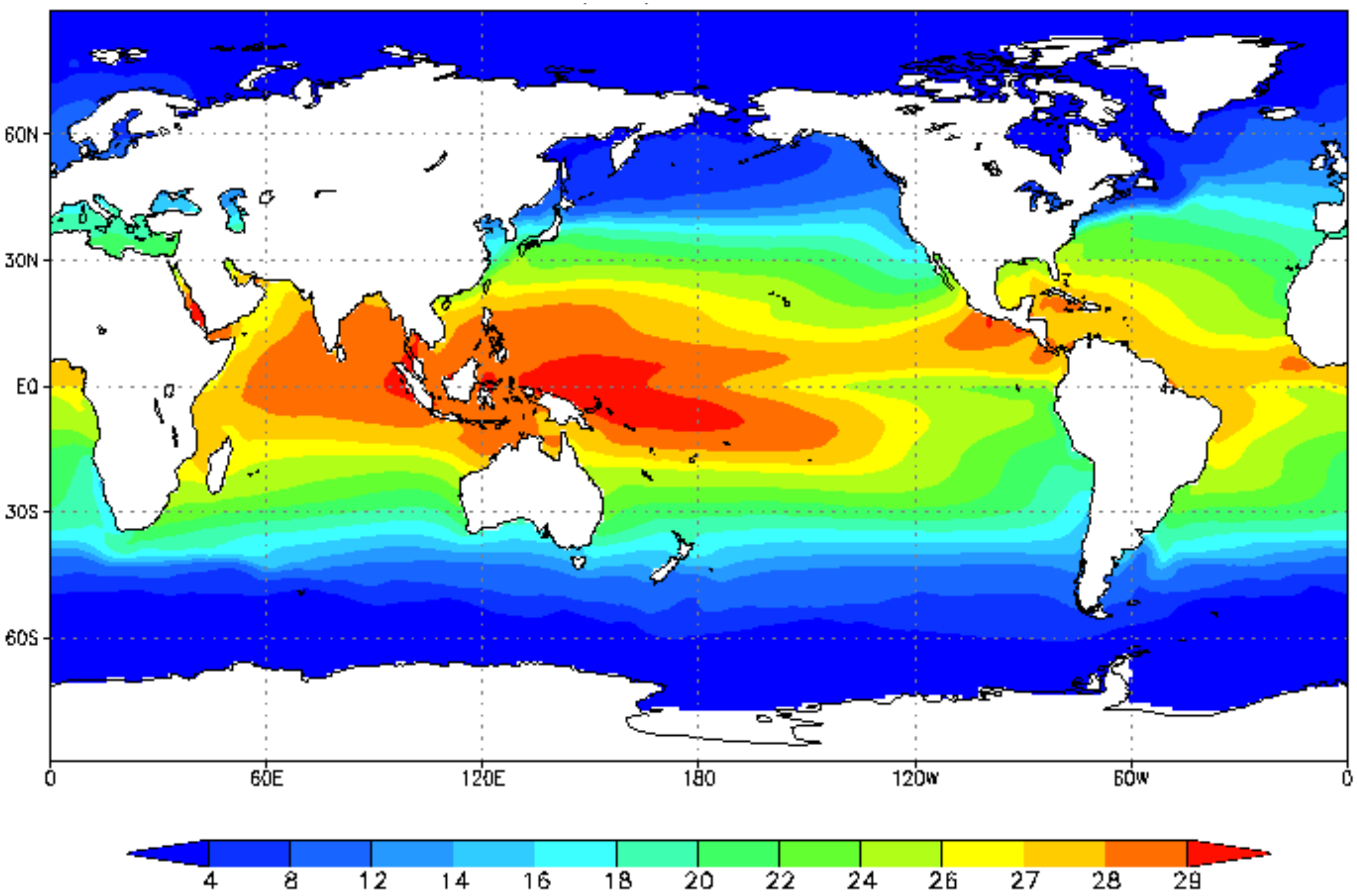
2 m/s

SST = 30°C

SST = 30°C

Ocean


60N
30N
EQ
30S
60S
0
60E
120E
180
120W
60W
0
4
8
12
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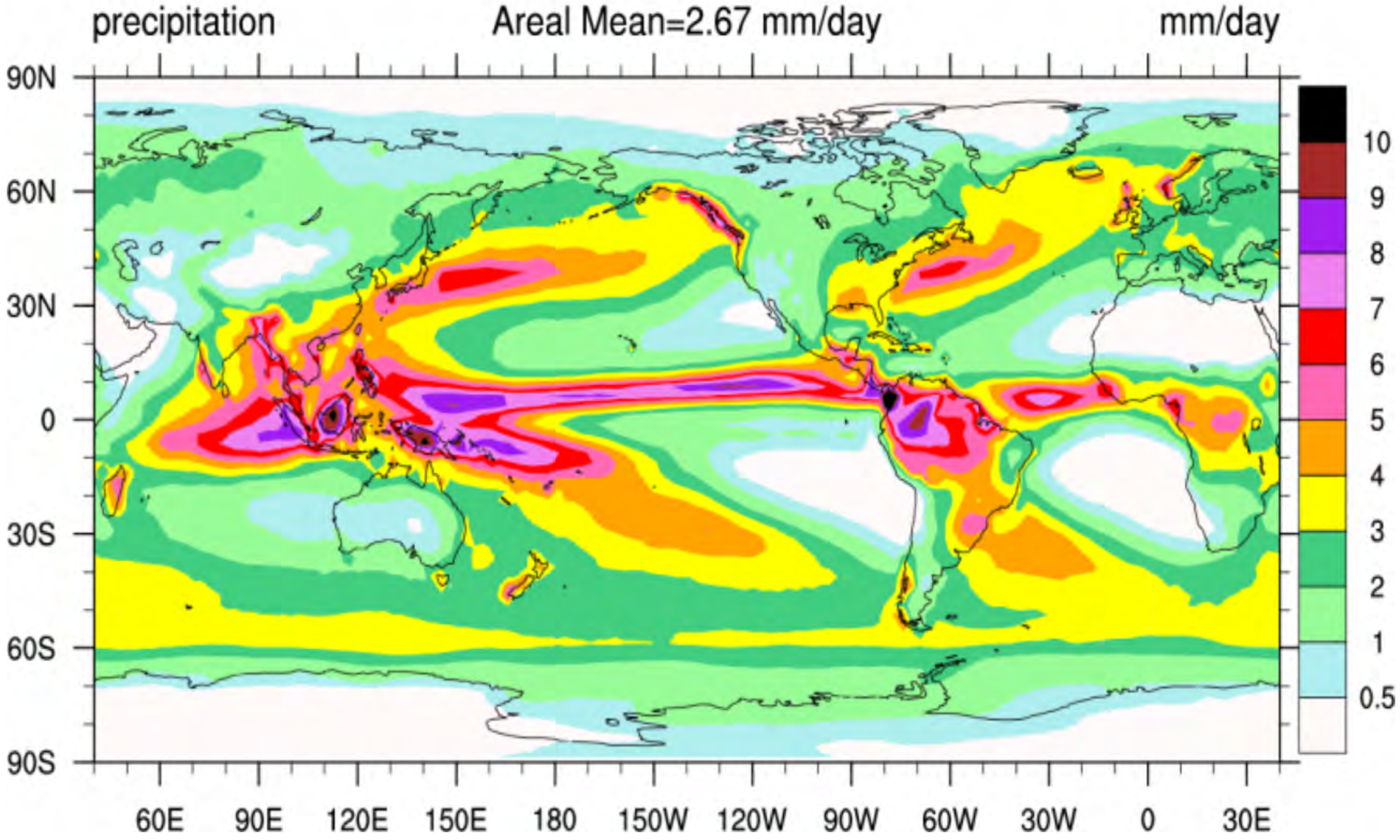
16
18
20
22
24
26
27
28
29

°C

TRMM GPCP: 1979-2010



MR3252: Tropical Meteorology

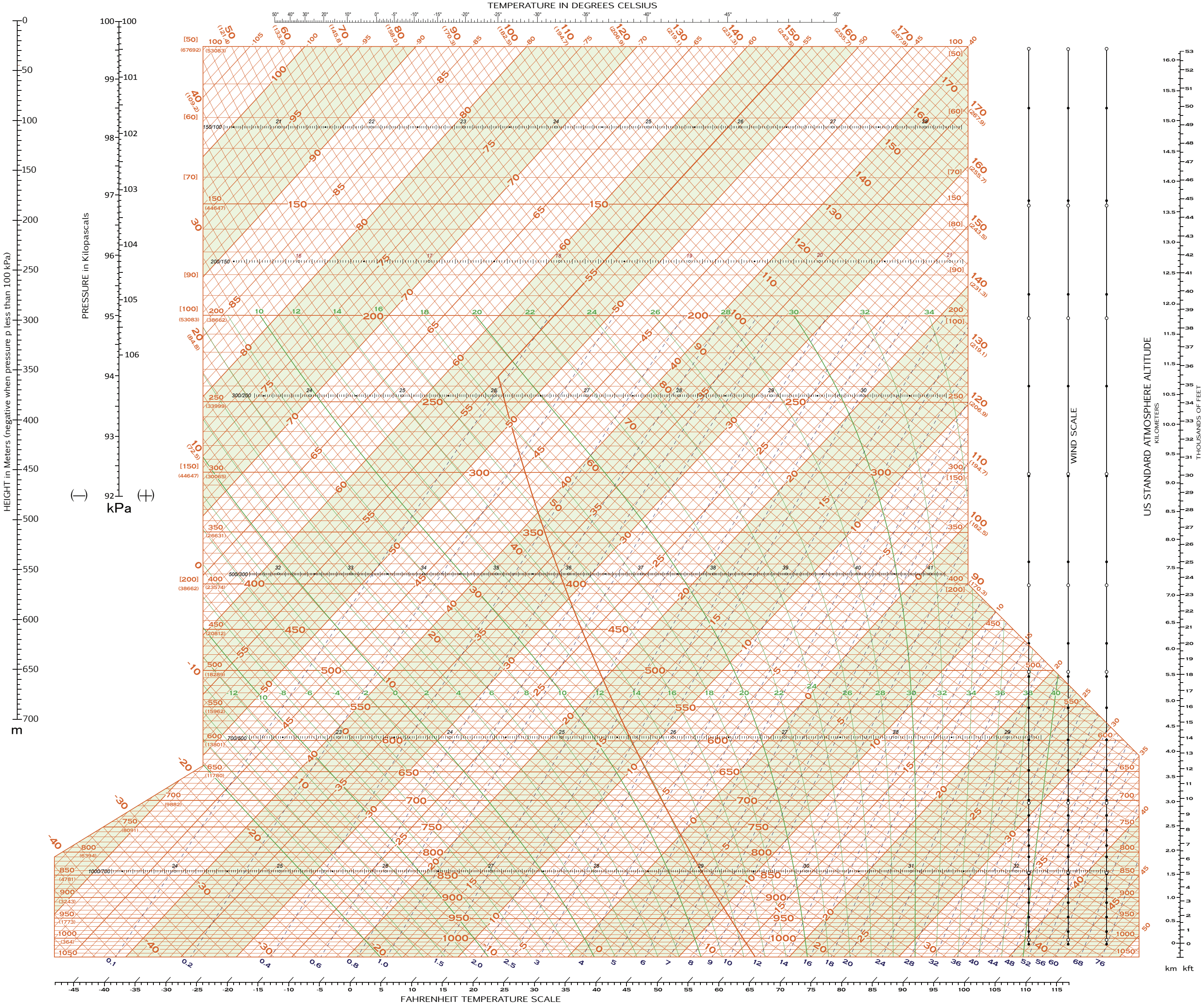
Thermodynamic Profiles and Important Quantities in the Tropical Atmosphere, Part I

Main Topics:

- Reading a Skew T-log P plot
- Potential temperature
- Equivalent potential temperature
- Diagnosing convective instability

Module 1.4

SKEW T ADIABATIC DIAGRAM



EXPLANATION

ISOBARS are straight horizontal brown lines. The heights in feet of the pressure surfaces in the U.S. Standard atmosphere are in parenthesis () below the pressure values on the left.

ISOTHERMS (°C) are the straight, equidistant brown lines running diagonally upward from the left to right.

DRY ADIABATS are the slightly curved brown lines that intersect the 1000 mb. isobar at intervals of 2°C, and run diaagonally upward from right to left. The Dry Adiabats for the folded portion of the pressure range are labeled with two (2) values. (See below).

SATURATED ADIABATS are the curved green lines that intersect the 1000 mb. isobar at intervals of 2°C, diverging upward and tending to become parallel to the dry adiabats.

SATURATION MIXING RATIO (in gm. per kg.) is represented by dashed green lines. Their values appear at the bottom of diagram.

THICKNESS (in hundreds of geopotential meters) of the layers between the levels 1000, 700, 500, 300, 200, 150, and 100 mb. is represented by numbers and a graduation along the middle of each layer. The thickness are obtained from the virtual temperature curve by the equal-area method, using any straight line as a dividing line.

HEIGHT in geopotential meters above mean sea level, or station level, of the 100 kPa surface is obtaind from the nomogram in the upper left-hand corner by drawing a straight line from the point on the temperature scale (°C) through the point p (mean sea level or station pressure) on the pressure scale, and reading the height on the height scale.

U.S. STANDARD ATMOSPHERE SOUNDING is indicated by a thick brown line.

The saturated adiabats and isopleths of saturation mising ratio are computed by use of vapor pressure over a plane water surface at all temperatures.

Extensions of chart to 50 mb. has been accomplished by overlap with pressure indicated in brackets, [200] at 400 mb., and [50] at 100 mb. Dry adiabats for the overlap are labeled in parenthesis. ( ).

APPROXIMATE VIRTUAL TEMPERATURE may be obtained from the formula T =T+ where T is virtual temperature in °C, T is free air temperature in °C, and w is the mixing ratio in grams/kilogram. For purposes of thickness computation, use the mean temperature of the layer for T and use the mean mixing ratio of the layer for w.

Black dots • along wind scale lines indicate the levels for which wind data is reported and plotted. The open circles ○ indicate the mandatory pressure level at which wind data is also entered.

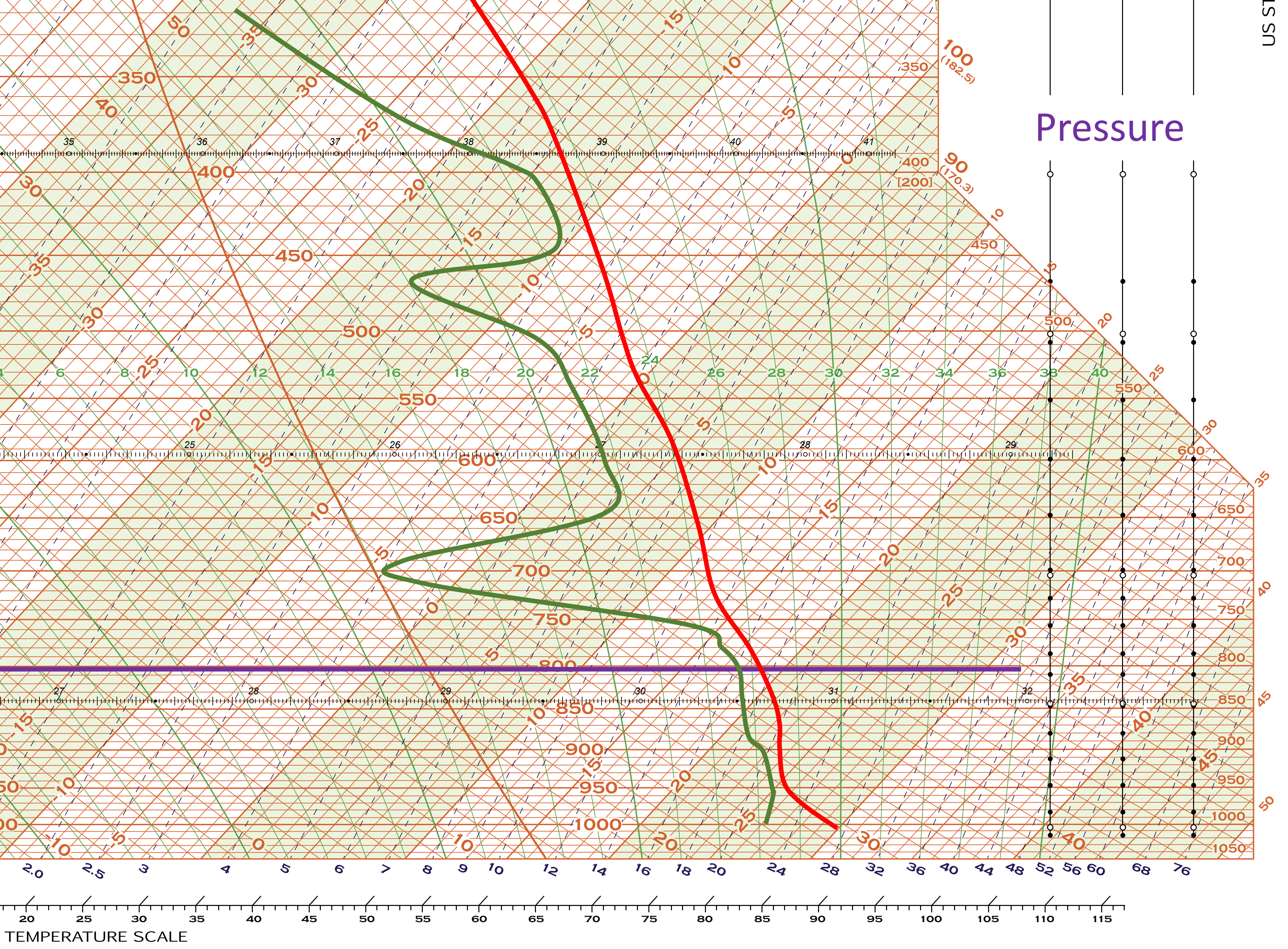
SKEW T ANALYSIS						
TIME				TIME		
AIRMASS ANALYSIS						
TYPE BOUNDARY ______ M. TYPE BOUNDARY ______ M. TYPE				______ M. ______ M.		
FREEZING LEVEL(S)						
INVERSIONS						
FRONTAL						
RADIATION						
SUBSIDENCE						
TROPOPAUSE						
L.C.L.						
C.C.L.						
L.F.C.						
SIGNIFICANT WIND						
MAX.						
MIN..						
LEVELS OF SHEAR						
STABILITY						
TO INDEX				TO INDEX		
TO				TO		
TO				TO		
CLOUDS						
TYPE						
AMOUNT						
BASES						
TOPS						
ICING						
TYPE						
SEVERITY						
BOUNDARIES						
CONTRAILS						
PERSISTENCE						
HEIGHT						
TURBULENCE						
DEGREE						
HEIGHT(S)						
MAX WIND GUSTS						
HAIL SIZE						
TEMPERATURES						
MAX.						
MIN.						
CUMULUS CLOUD FORMATION AT TEMP ______ TIME ______						
DISSIPATION OF LOW LEVEL INVERSION AT ______ TIME ______						

REMARKS

STATION ID	STATION ID	STATION ID
DATE	DATE	DATE
TIME (GMT)	TIME (GMT)	TIME (GMT)

CHART D-7

NOAA - National Weather Service 1977 Jan



Pressure

US ST
100
(182.5)
90
(170.3)
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[200]
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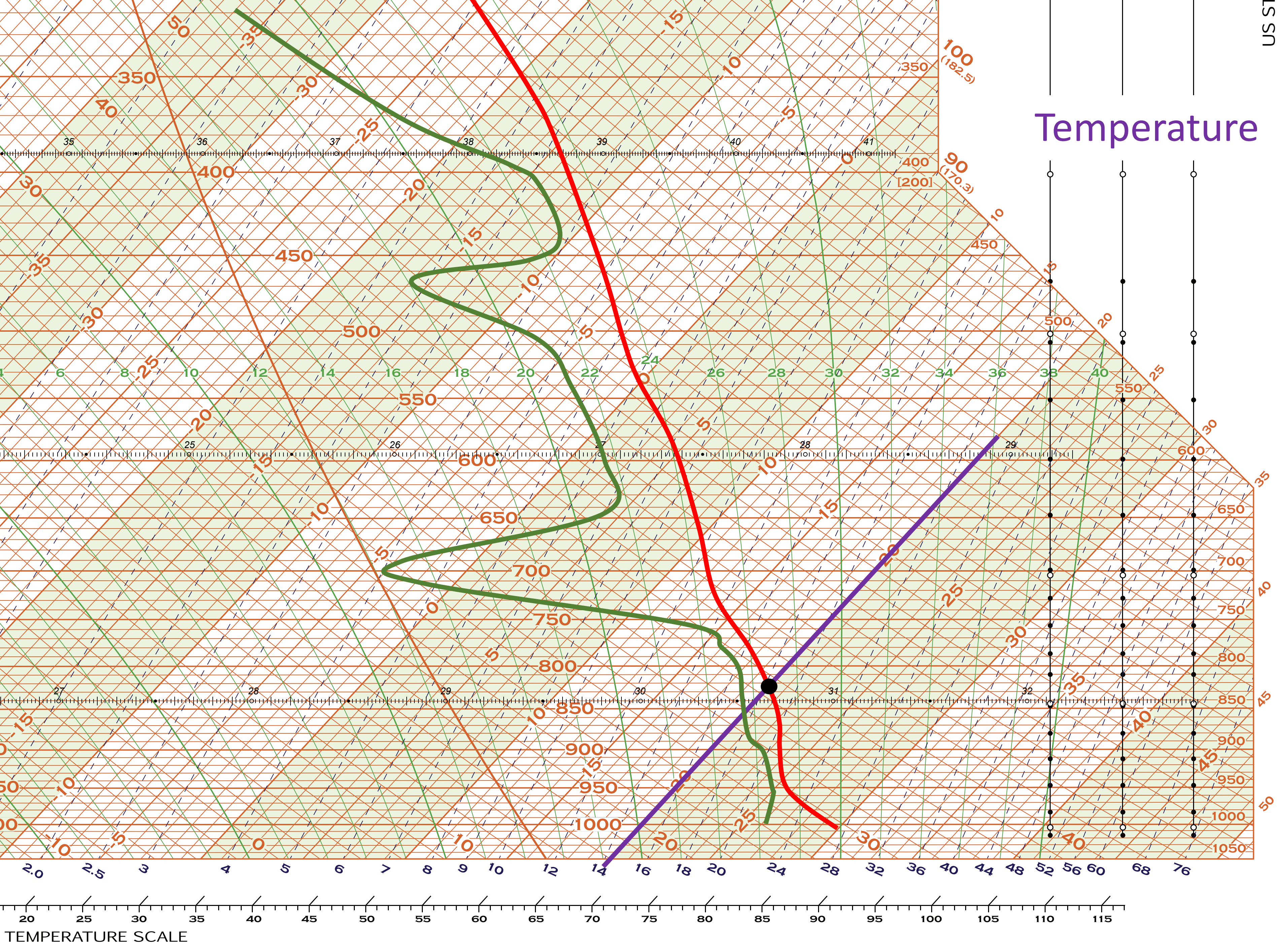
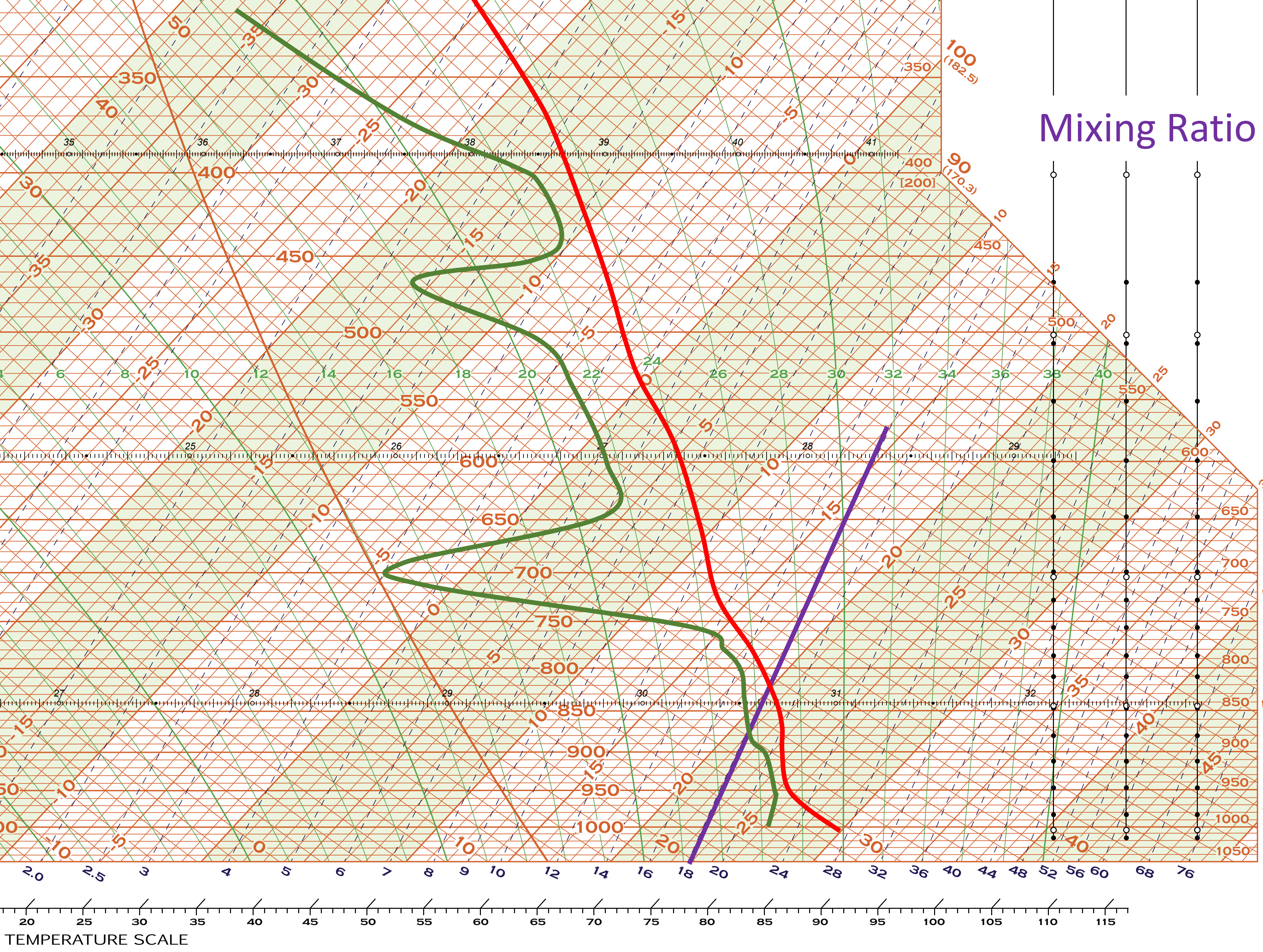
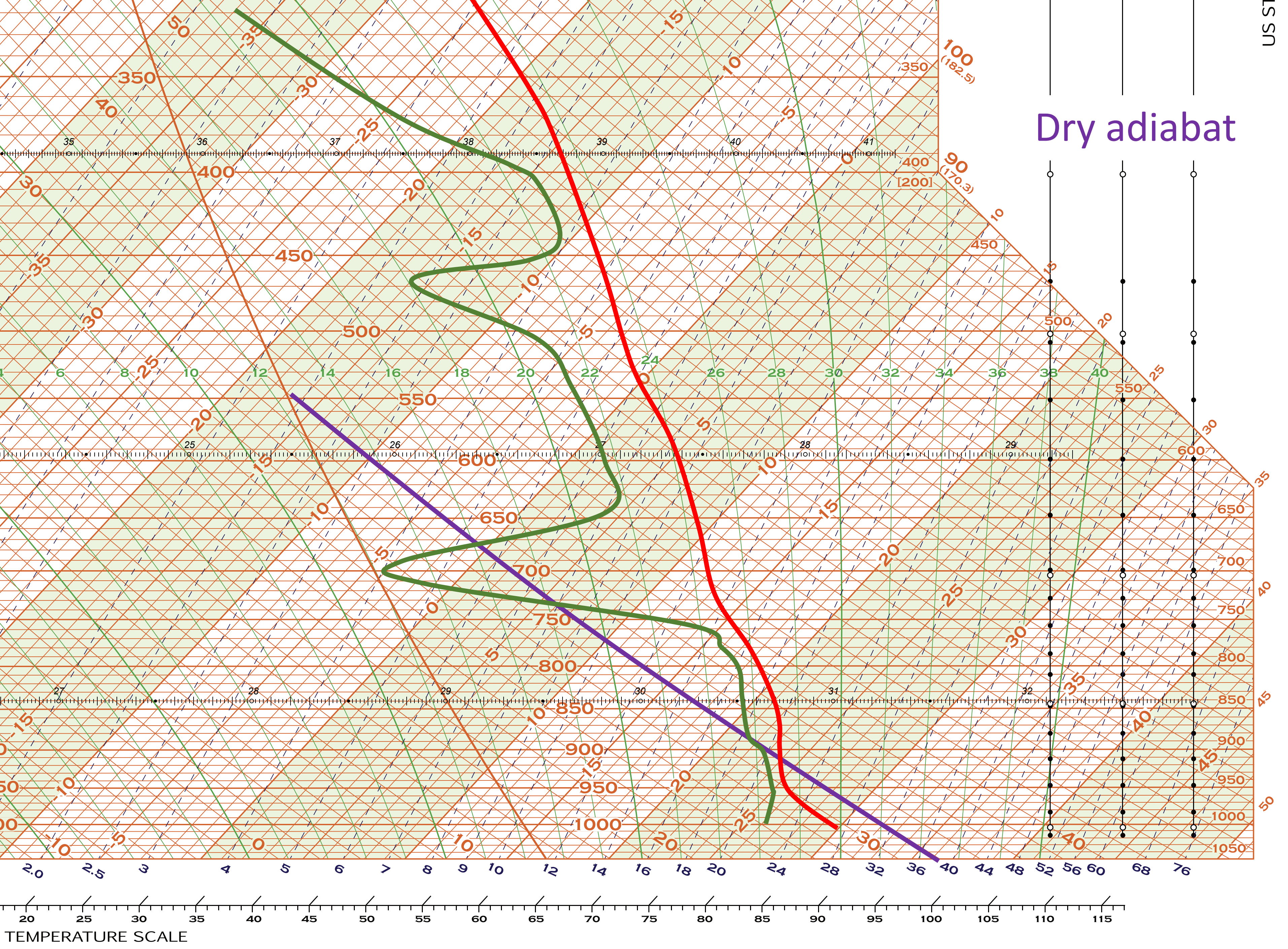
700
750
800
850
900
950
1000
1050
TEMPERATURE SCALE

Temperature



Mixing Ratio




Dry adiabat
TEMPERATURE SCALE
100
350
400
450
500
550
600
650
700
750
800
850
900
950
1000
1050
2.0
2.5
3
4
5
6
7
8
9
10
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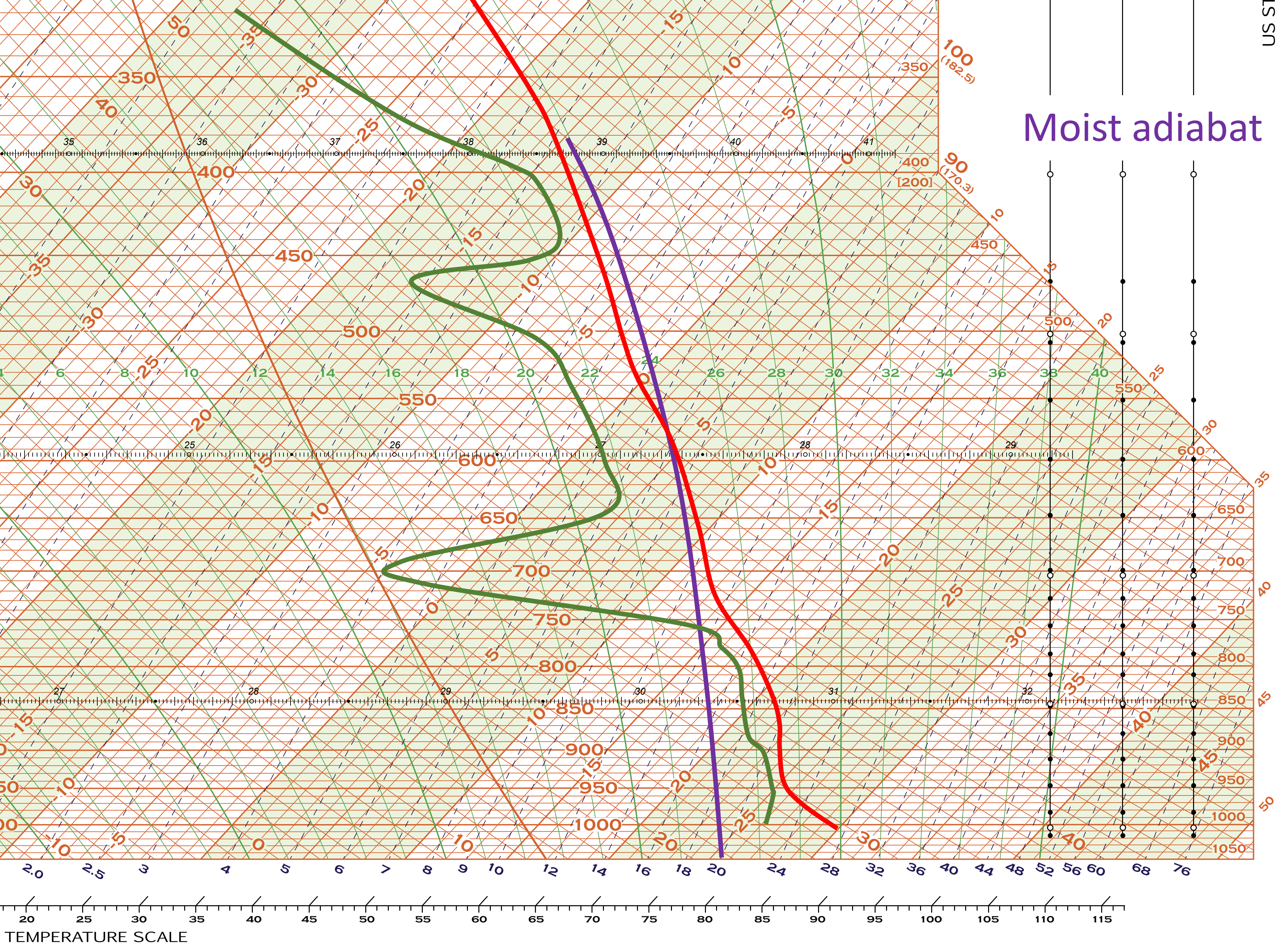
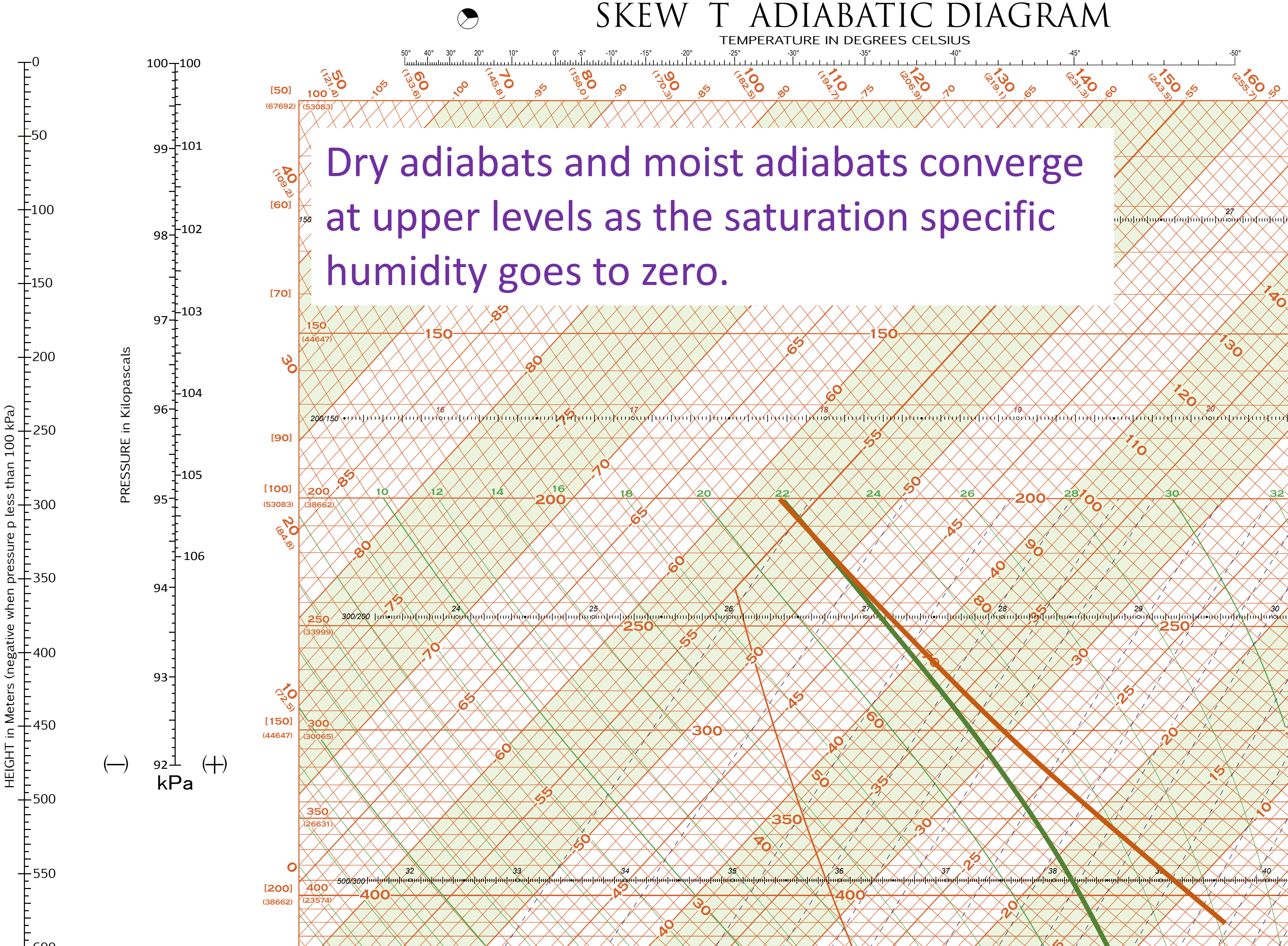
110
115


Moist adiabat
TEMPERATURE SCALE
US ST
100
350
400
450
500
550
600
650
700
750
800
850
900
950
1000
1050



Dry adiabats and moist adiabats converge at upper levels as the saturation specific humidity goes to zero.

Some quantities with which to be familiar:

Potential temperature (θ): Temperature of air dry adiabatically lifted or descended to 1000 hPa.

Equivalent potential temperature (θ_e): Temperature of air if lifted adiabatically until all water vapor is condensed, then descended dry adiabatically to 1000 hPa.

Saturation equivalent potential temperature (θ_{es}): Same as θ_e, but assuming air is initially saturated. $\theta_{es} \geq \theta_e$.

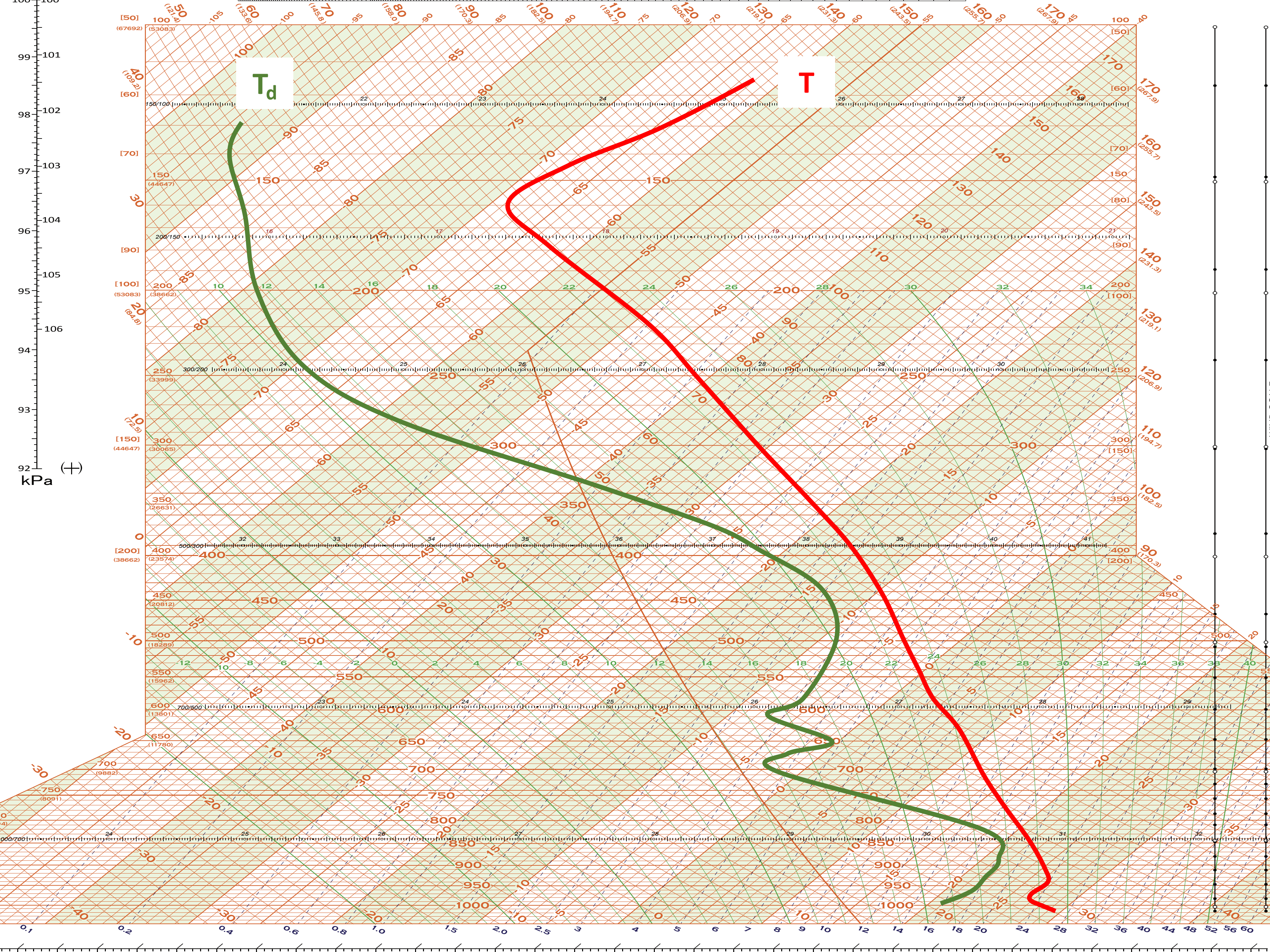
Virtual potential temperature (θ_v): Potential temperature of dry air that would have the same density as moist air. In the Tropics, may be O(1K) larger than θ.

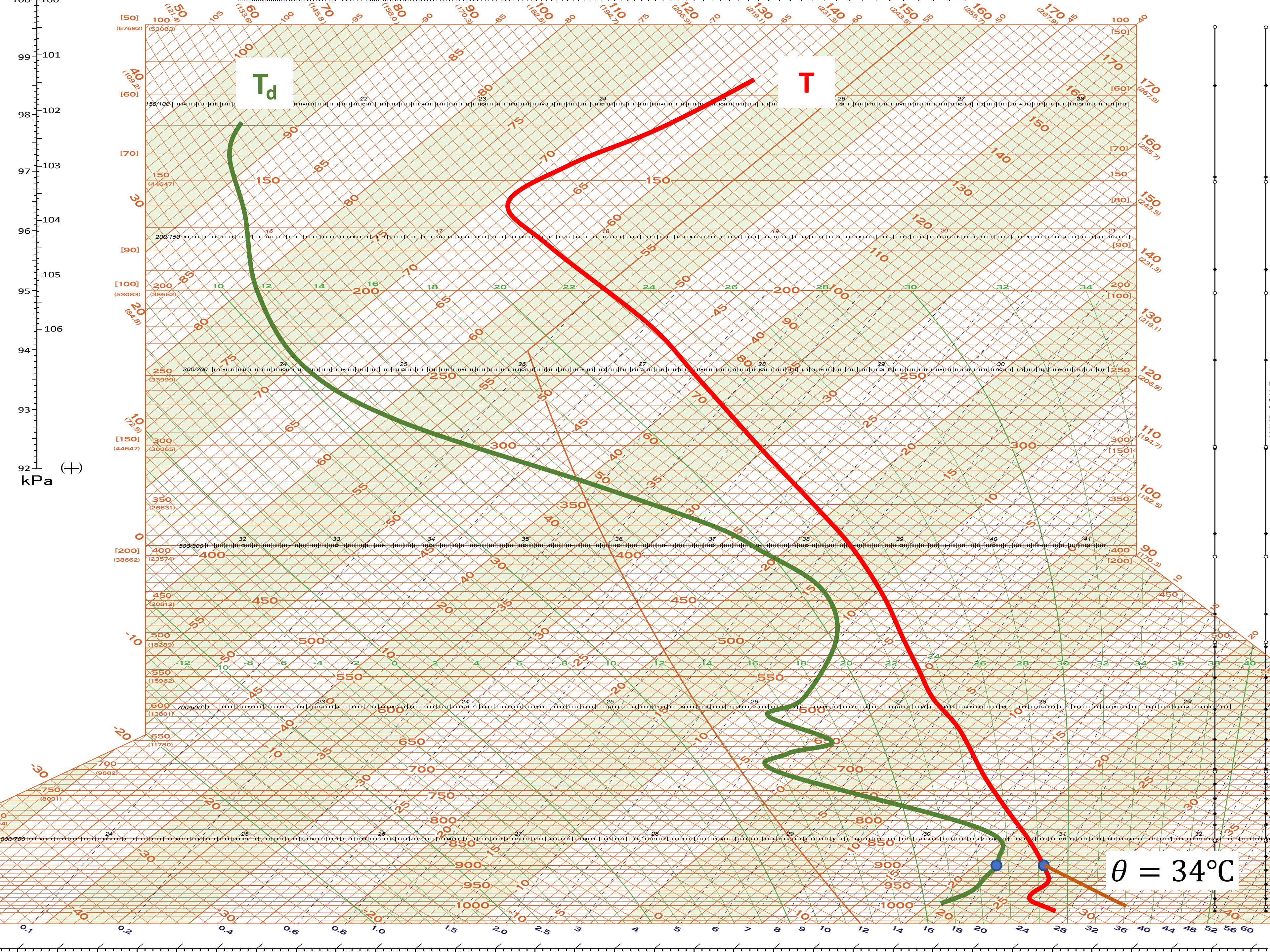
Environmental lapse rate: Observed change of temperature or a potential temperature variable with height.

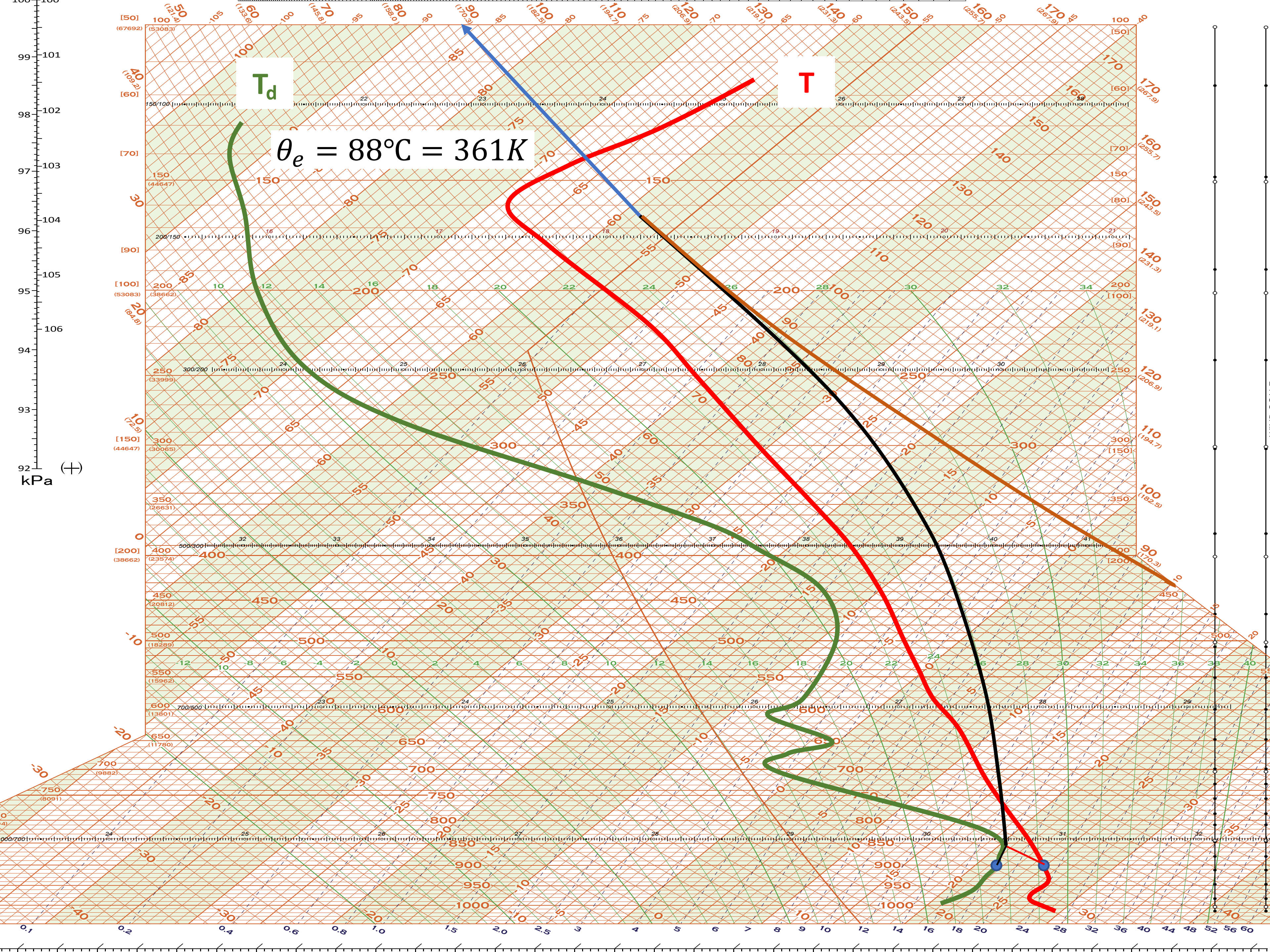
Dry adiabatic lapse rate: Adiabatic lapse rate that applies to sub-saturated air: $\Gamma_d = \frac{g}{c_p} \approx 9.81\mathrm{K/km}$

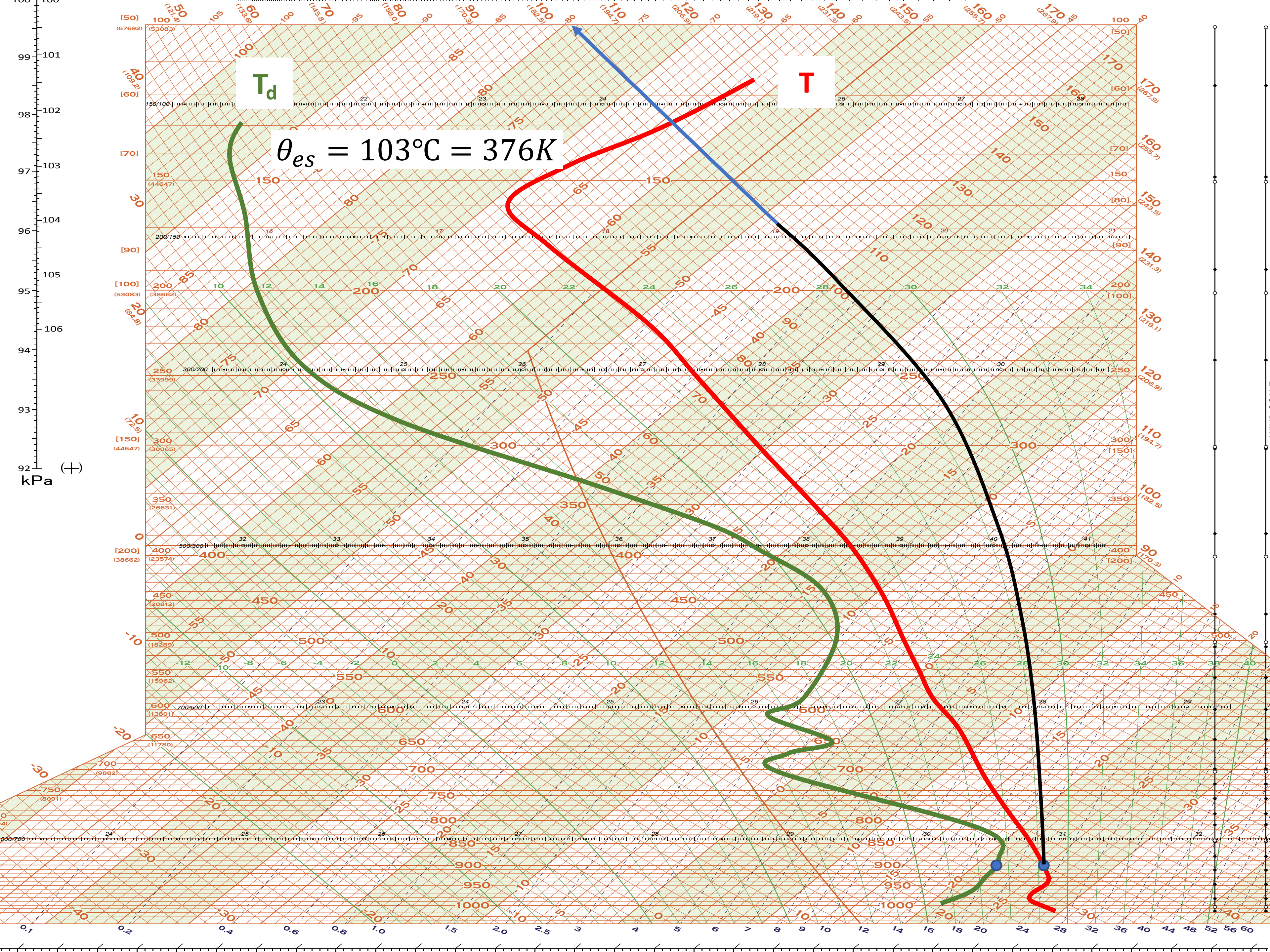
Moist adiabatic lapse rate: Adiabatic lapse rate that applies to saturated air:

$$\Gamma_m = g \frac{1 + \frac{L_v r}{RT}}{c_p + \frac{\epsilon L_v^2 r}{RT^2}}$$


Td
T
kPa
100
150
200
250
300
350
400
450
500
550
600
650
700
750
800
850
900
950
1000
[50]
[60]
[70]
[80]
[90]
[100]
[150]
[200]
150/100
200/150
300/200
500/300
700/500
1000/700


Td
T
θ = 34°C
kPa
100
150
200
250
300
350
400
450
500
550
600
650
700
750
800
850
900
950
1000


Td
T
$\theta_e = 88°C = 361K$
kPa



$\theta_{es} = 103°C = 376K$

Some quantities with which to be familiar:

Potential temperature (θ): Temperature of air dry adiabatically lifted or descended to 1000 hPa.

Equivalent potential temperature (θ_e): Temperature of air if lifted (dry) adiabatically until all water vapor is condensed, then descended (moist) adiabatically to 1000 hPa.

Saturated equivalent potential temperature (θ_{es}): Same as θ_e, but assuming air is initially saturated. $\theta_{es} \geq \theta_e$.

Virtual potential temperature (θ_v): Potential temperature of dry air that would have the same density as moist air. In the Tropics, may be O(1K) larger than θ.

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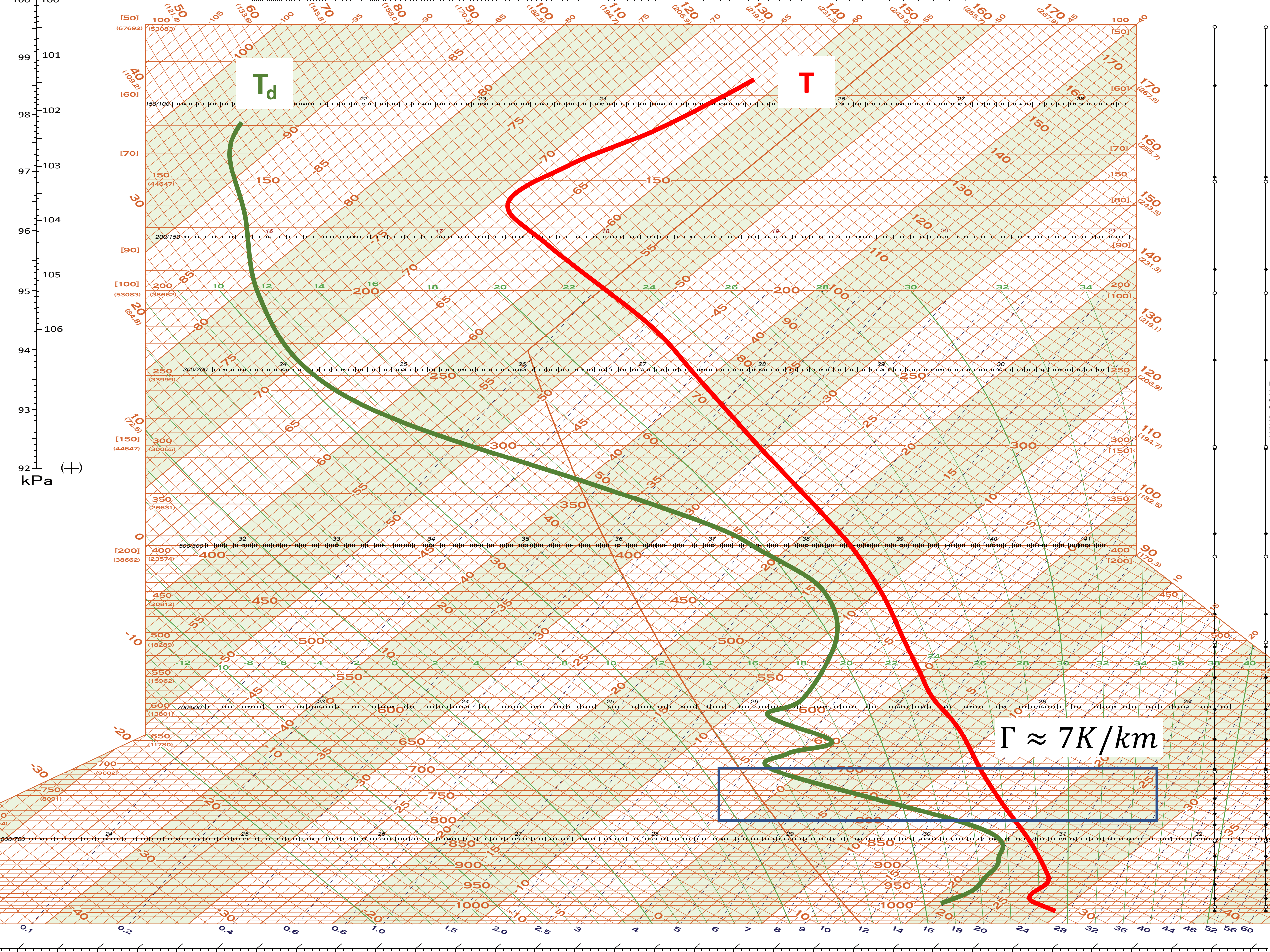
$$\Gamma_m = g\,\frac{1 + \frac{L_v r}{RT}}{c_p + \frac{\epsilon L_v^2 r}{RT^2}}$$



$\Gamma \approx 7K/km$

Stability Metrics

Dry Air (Module 1.2)

Temperature	**Potential Temperature**	**Stability**
$\Gamma > \Gamma_d$	$\frac{\partial\theta}{\partial z} < 0$	Unstable
$\Gamma = \Gamma_d$	$\frac{\partial\theta}{\partial z} = 0$	Neutral
$\Gamma < \Gamma_d$	$\frac{\partial\theta}{\partial z} > 0$	Stable

Moist air

Temperature	**Saturated Equivalent Potential Temperature**	**Stability**
$\Gamma > \Gamma_d$	$\frac{\partial\theta_{es}}{\partial z} < 0$	Unstable
$\Gamma_m \le \Gamma \le \Gamma_d$	$\frac{\partial\theta_{es}}{\partial z} < 0$	Conditionally Unstable
$\Gamma = \Gamma_m$	$\frac{\partial\theta_{es}}{\partial z} = 0$	Moist Neutral
$\Gamma < \Gamma_m$	$\frac{\partial\theta_{es}}{\partial z} > 0$	Stable

θ_e^* is just a different notation for θ_{es}.

Stability Metrics

Dry Air (Module 1.2)

Temperature	**Potential Temperature**	**Stability**
$\Gamma > \Gamma_d$	$\frac{\partial \theta}{\partial z} < 0$	Unstable
$\Gamma = \Gamma_d$	$\frac{\partial \theta}{\partial z} = 0$	Neutral
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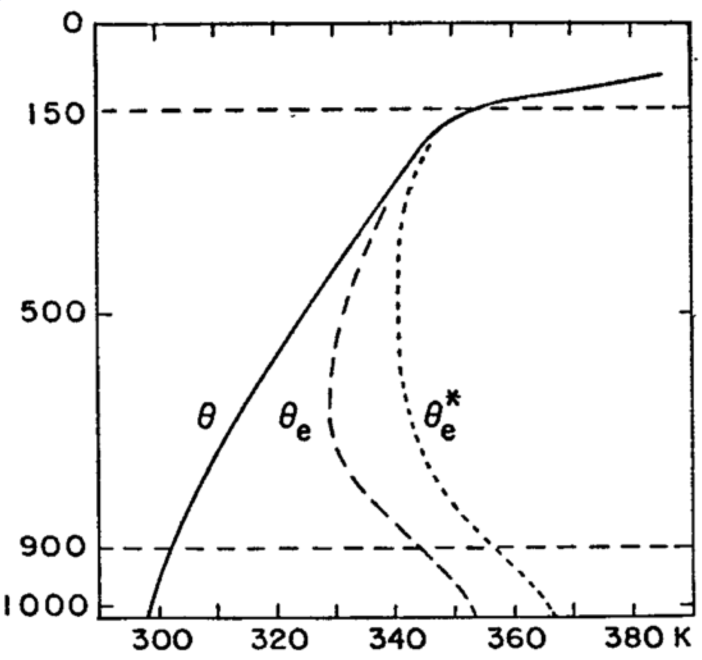


Moist air

Temperature	**Saturated Equivalent Potential Temperature**	**Stability**
$\Gamma > \Gamma_d$	$\frac{\partial \theta_{es}}{\partial z} < 0$	Unstable
$\Gamma_m \le \Gamma \le \Gamma_d$	$\frac{\partial \theta_{es}}{\partial z} < 0$	Conditionally Unstable
$\Gamma = \Gamma_m$	$\frac{\partial \theta_{es}}{\partial z} = 0$	Moist Neutral
$\Gamma < \Gamma_m$	$\frac{\partial \theta_{es}}{\partial z} > 0$	Stable

θ_e^* is the same as θ_{es}.

MR3252: Tropical Meteorology

Thermodynamic Profiles and Important Quantities in the Tropical Atmosphere, Part II

Main Topics:

- Interpretation of tropical thermodynamic profiles
- Moist static energy
- Convective available potential energy

Module 1.5

Some quantities with which to be familiar:

Total Precipitable Water (TPW): The depth of a hypothetical unit volume of liquid water that would be produced if all water vapor in the atmosphere were precipitated out

$$TPW = \frac{\int_{PSFC}^{0} r(p)dp}{\rho g}$$

Moist static energy (**MSE**, often *h*): Similar to θ_e.
Sort of conserved in adiabatic processes, but not exactly if we consider hydrometeors in *q*.

$$h \approx c_pT + \Phi + L_vq$$

Brackets often used to indicate "column-integrated" values. Here *s* is dry static energy.

↑

Gross moist stability (GMS, many variables used, *M* here**)**: A quantity that compares the lateral export of some conserved quantity (in adiabatic processes) to some measure of convective intensity. GMS is negative if MSE is imported (in net) into a column. See Raymond et al. (2009) and Inoue and Back (2015).

$$M = \frac{\nabla \cdot \langle h\mathbf{v} \rangle}{\nabla \cdot \langle s\mathbf{v} \rangle}$$

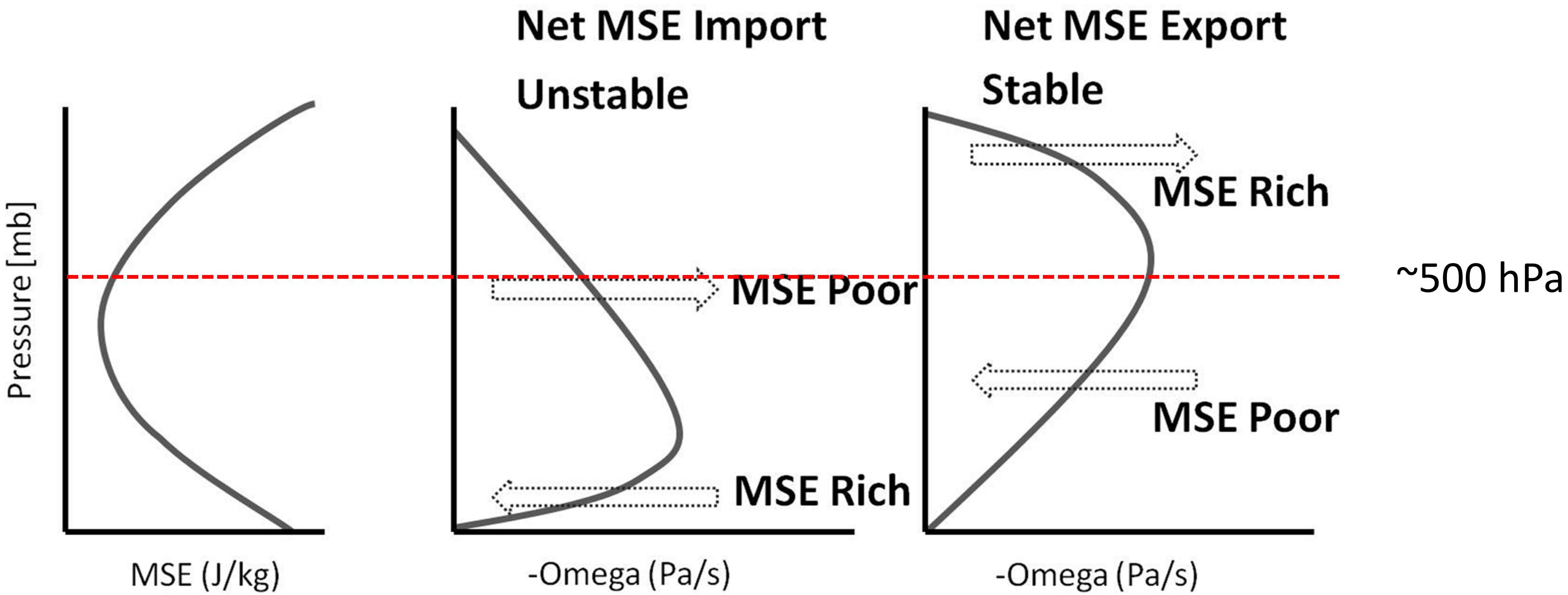
Convective available potential energy (**CAPE**): Maximum buoyancy of *undiluted* air parcel.

$$CAPE = \int_{P_{LFC}}^{P_{EL}} R\left(T_{v,p} - T_{v,e}\right) d\ln p$$

Typical MSE profile

Negative GMS

Positive GMS



Inoue and Back (2015)

Some quantities with which to be familiar:

Total Precipitable Water (TPW): The depth of a hypothetical unit volume of liquid water that would be produced if all water vapor in the atmosphere were precipitated out

$$TPW = \frac{\int_{PSFC}^{0} r(p)dp}{\rho g}$$

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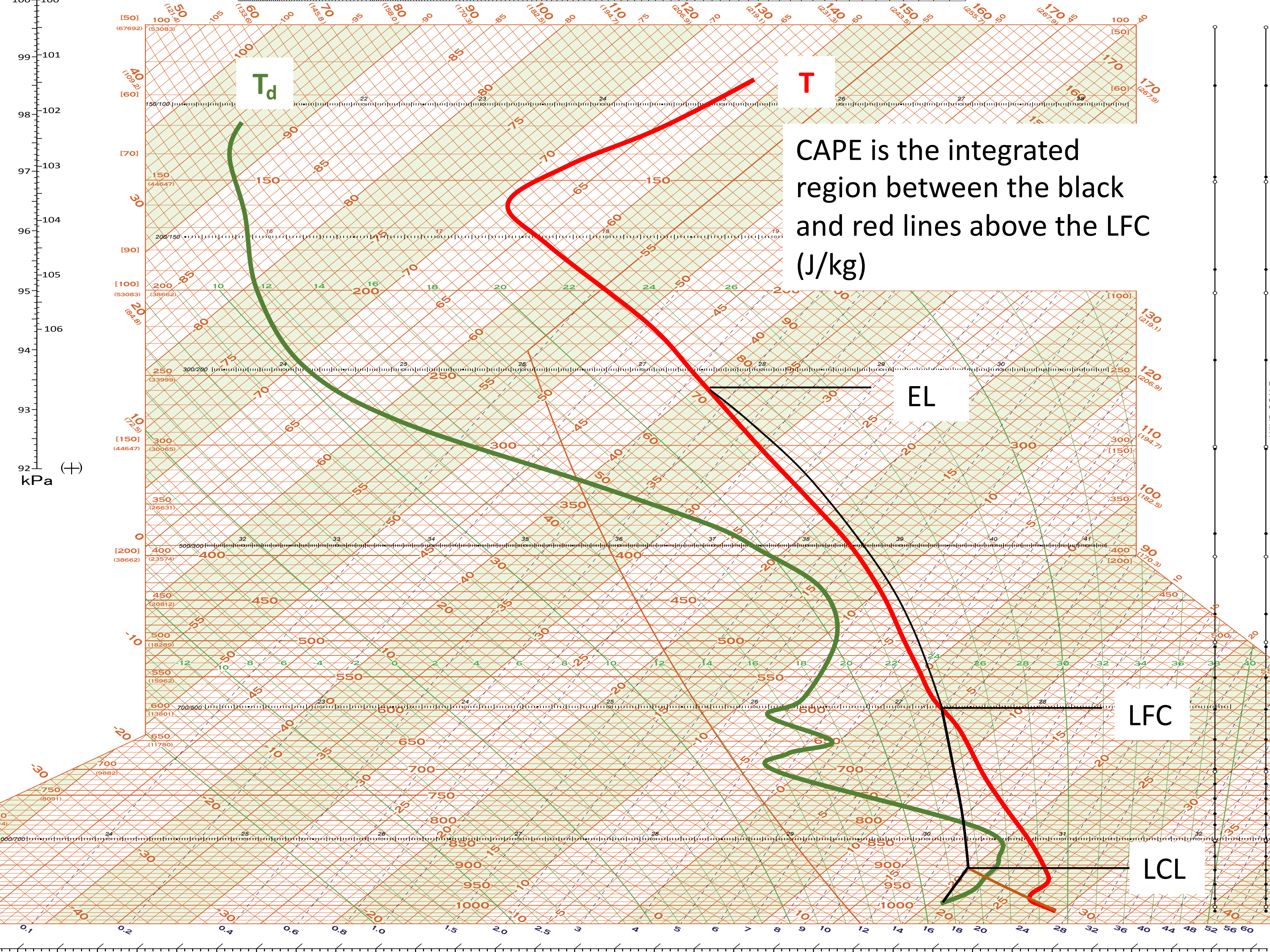
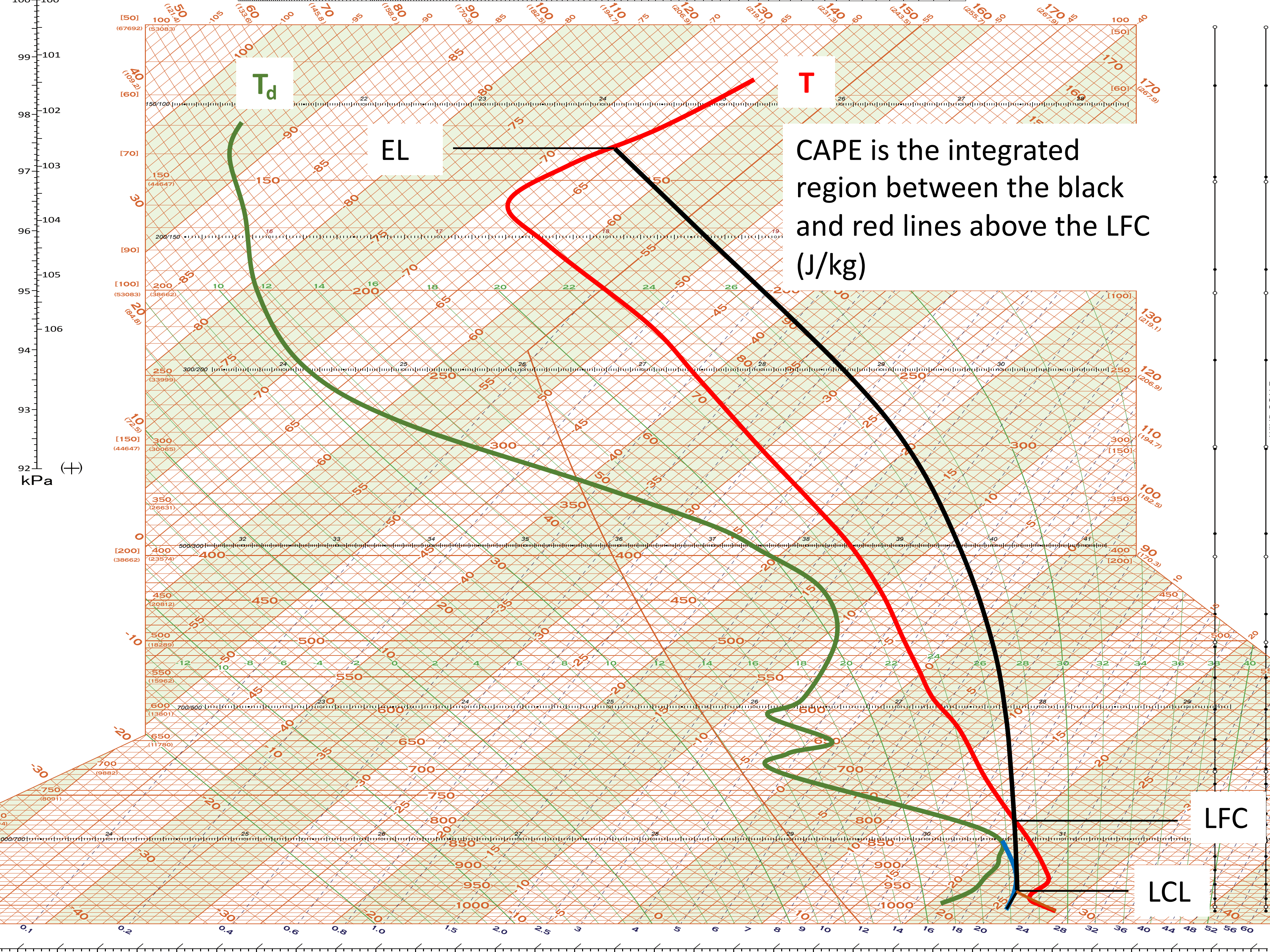
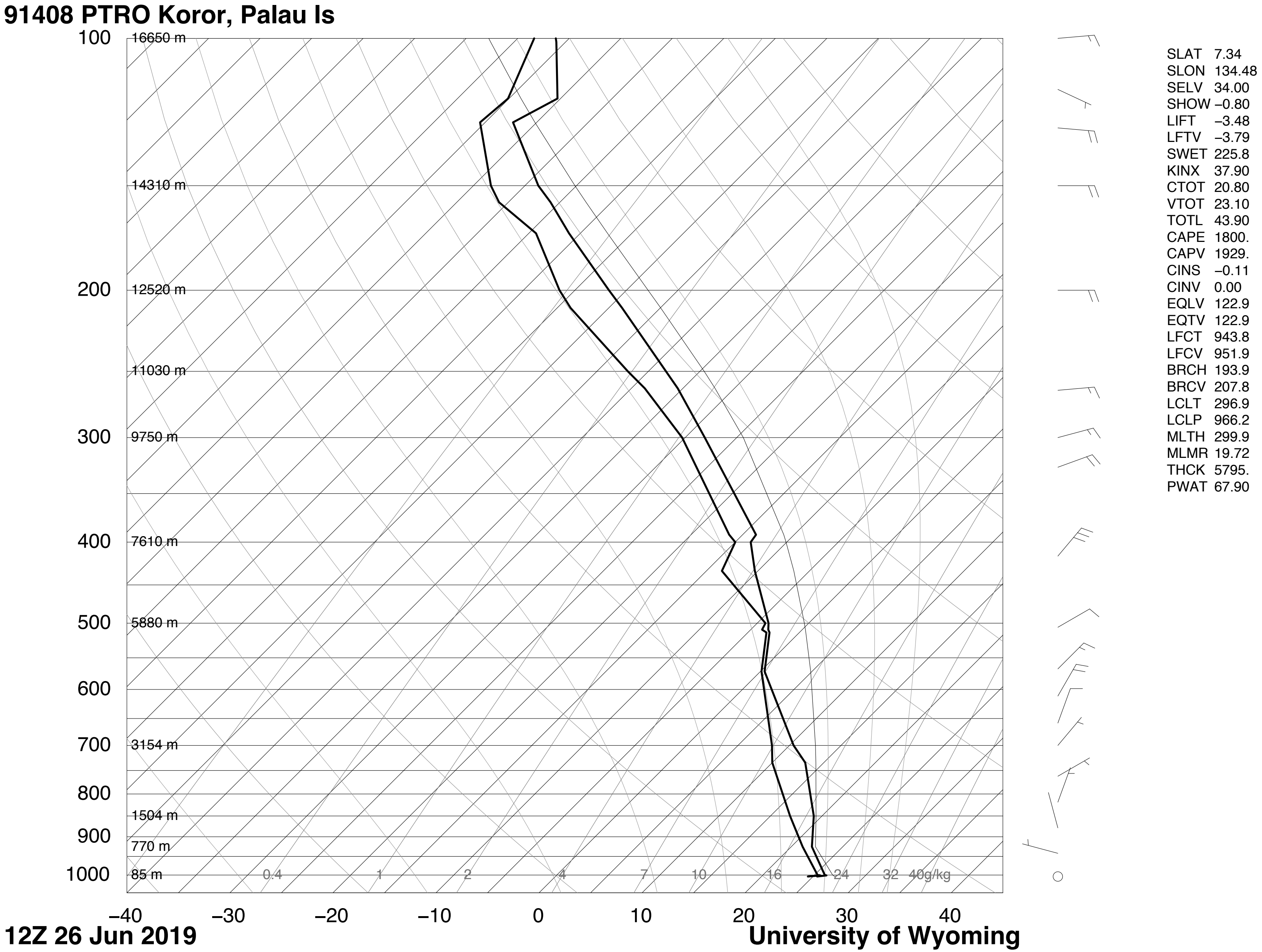


CAPE is the integrated region between the black and red lines above the LFC (J/kg)



CAPE is the integrated region between the black and red lines above the LFC (J/kg)

Tropical West Pacific: Near-saturated sounding, probably through organized MCS

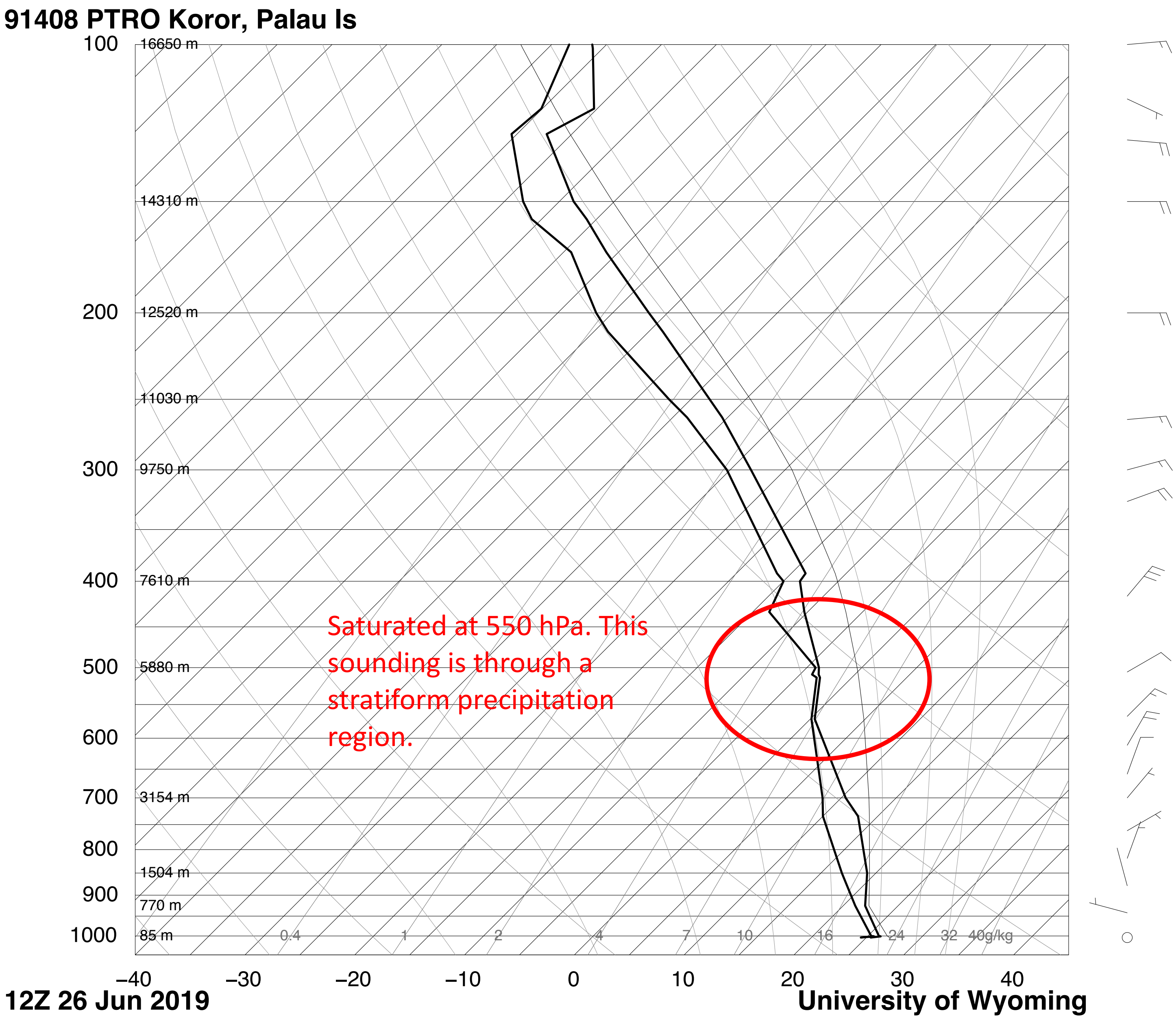


91408 PTRO Koror, Palau Is

SLAT	7.34
SLON	134.48
SELV	34.00
SHOW	−0.80
LIFT	−3.48
LFTV	−3.79
SWET	225.8
KINX	37.90
CTOT	20.80
VTOT	23.10
TOTL	43.90
CAPE	1800.
CAPV	1929.
CINS	−0.11
CINV	0.00
EQLV	122.9
EQTV	122.9
LFCT	943.8
LFCV	951.9
BRCH	193.9
BRCV	207.8
LCLT	296.9
LCLP	966.2
MLTH	299.9
MLMR	19.72
THCK	5795.
PWAT	67.90

12Z 26 Jun 2019 **University of Wyoming**

Tropical West Pacific: Near-saturated sounding, probably through organized MCS

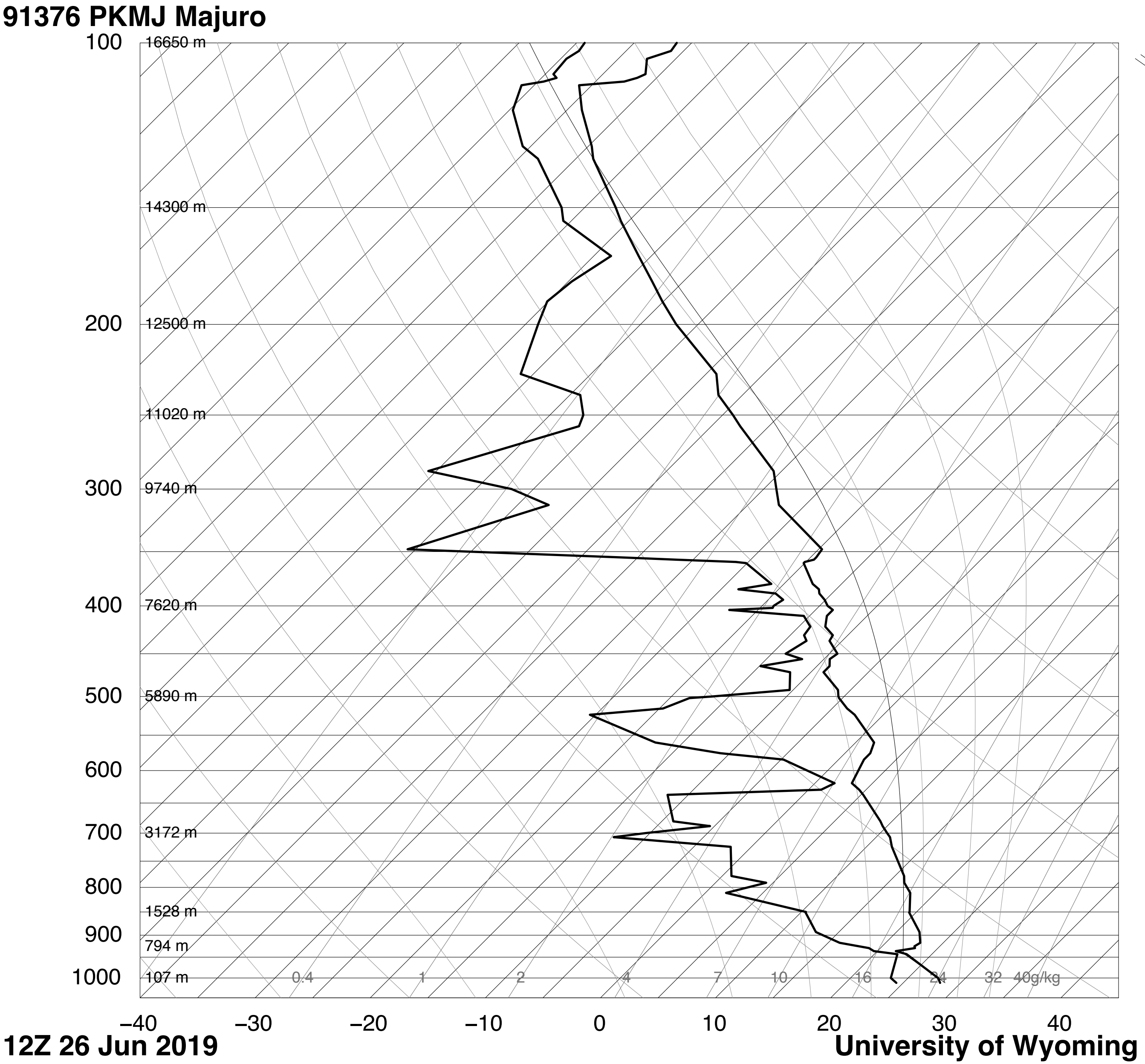


Saturated at 550 hPa. This sounding is through a stratiform precipitation region.

SLAT	7.34
SLON	134.48
SELV	34.00
SHOW	−0.80
LIFT	−3.48
LFTV	−3.79
SWET	225.8
KINX	37.90
CTOT	20.80
VTOT	23.10
TOTL	43.90
CAPE	1800.
CAPV	1929.
CINS	−0.11
CINV	0.00
EQLV	122.9
EQTV	122.9
LFCT	943.8
LFCV	951.9
BRCH	193.9
BRCV	207.8
LCLT	296.9
LCLP	966.2
MLTH	299.9
MLMR	19.72
THCK	5795.
PWAT	67.90

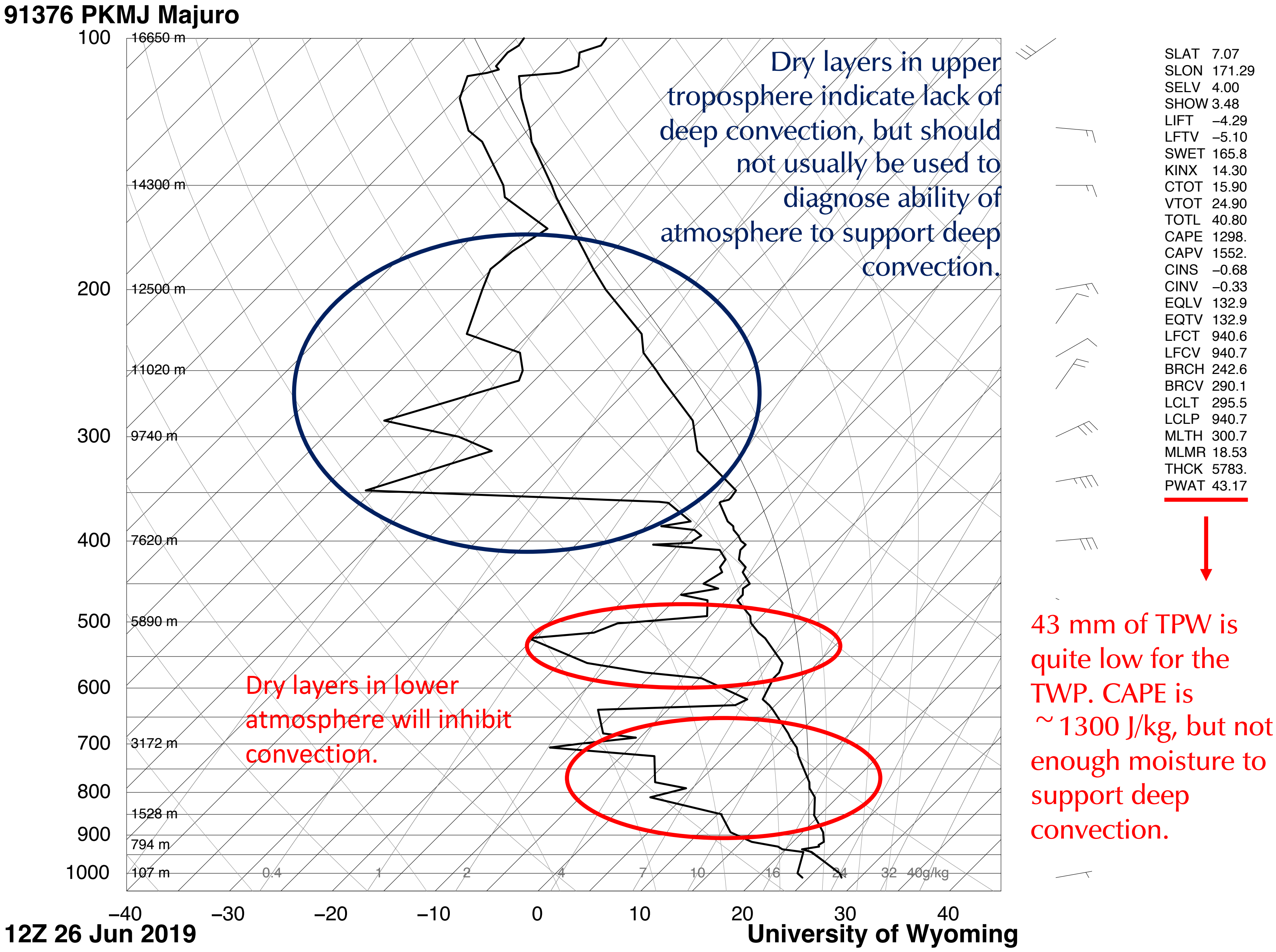
Very high TPW!

Tropical West Pacific: Drier marine sounding; supportive of shallow convection, but not deep, organized convection



SLAT	7.07
SLON	171.29
SELV	4.00
SHOW	3.48
LIFT	−4.29
LFTV	−5.10
SWET	165.8
KINX	14.30
CTOT	15.90
VTOT	24.90
TOTL	40.80
CAPE	1298.
CAPV	1552.
CINS	−0.68
CINV	−0.33
EQLV	132.9
EQTV	132.9
LFCT	940.6
LFCV	940.7
BRCH	242.6
BRCV	290.1
LCLT	295.5
LCLP	940.7
MLTH	300.7
MLMR	18.53
THCK	5783.
PWAT	43.17

Tropical West Pacific: Drier marine sounding; supportive of shallow convection, but not deep, organized convection



SLAT	7.07
SLON	171.29
SELV	4.00
SHOW	3.48
LIFT	−4.29
LFTV	−5.10
SWET	165.8
KINX	14.30
CTOT	15.90
VTOT	24.90
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CAPE	1298.
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CINS	−0.68
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EQLV	132.9
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LFCT	940.6
LFCV	940.7
BRCH	242.6
BRCV	290.1
LCLT	295.5
LCLP	940.7
MLTH	300.7
MLMR	18.53
THCK	5783.
PWAT	43.17

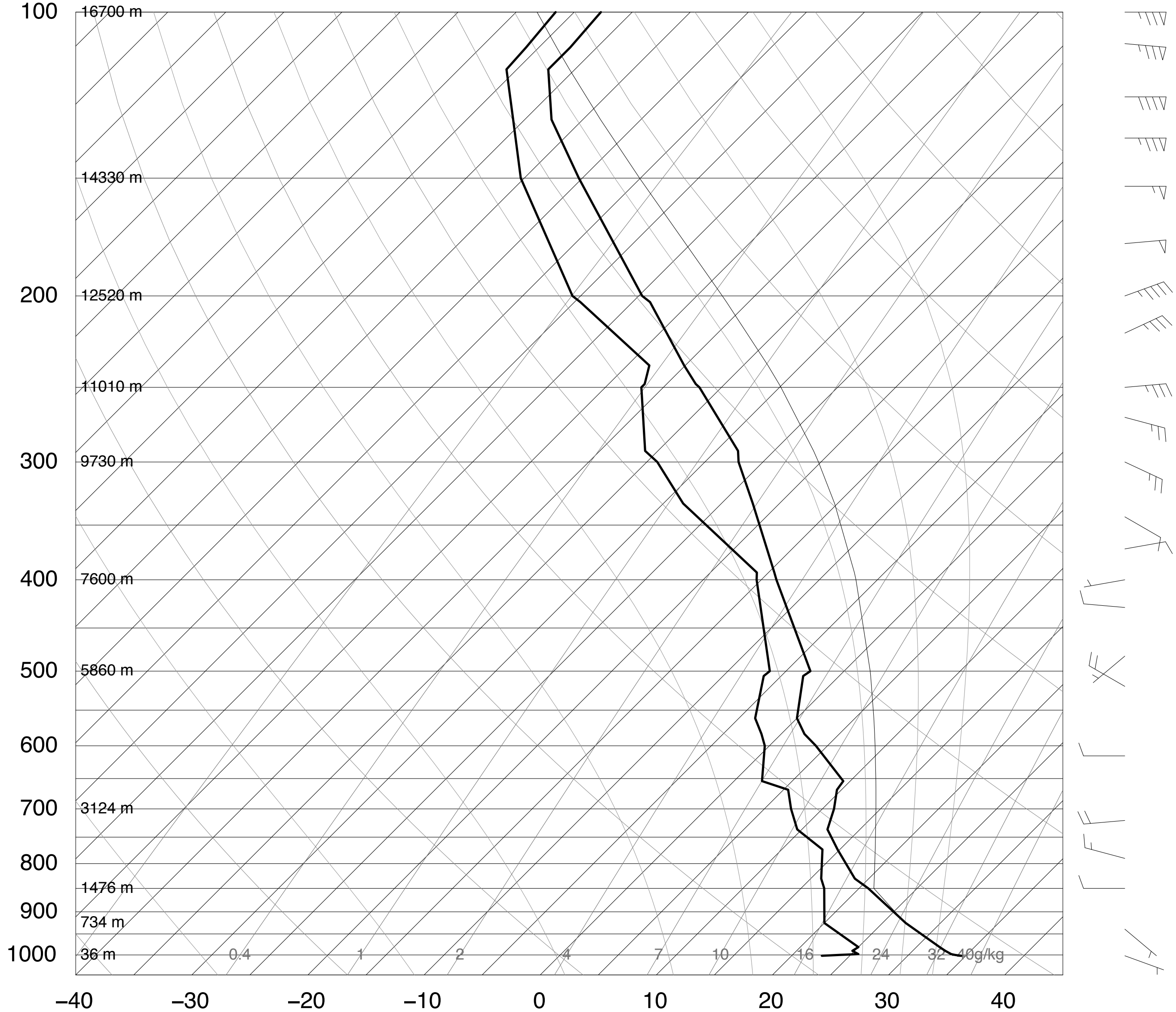
Dry layers in upper troposphere indicate lack of deep convection, but should not usually be used to diagnose ability of atmosphere to support deep convection.

Dry layers in lower atmosphere will inhibit convection.

43 mm of TPW is quite low for the TWP. CAPE is ~1300 J/kg, but not enough moisture to support deep convection.

South Asia coastal sounding: Lots of moisture and CAPE. PBL lapse rate almost dry adiabatic. High LCL/LFC though, forcing needed for deep convection.

43279 VOMM Madras

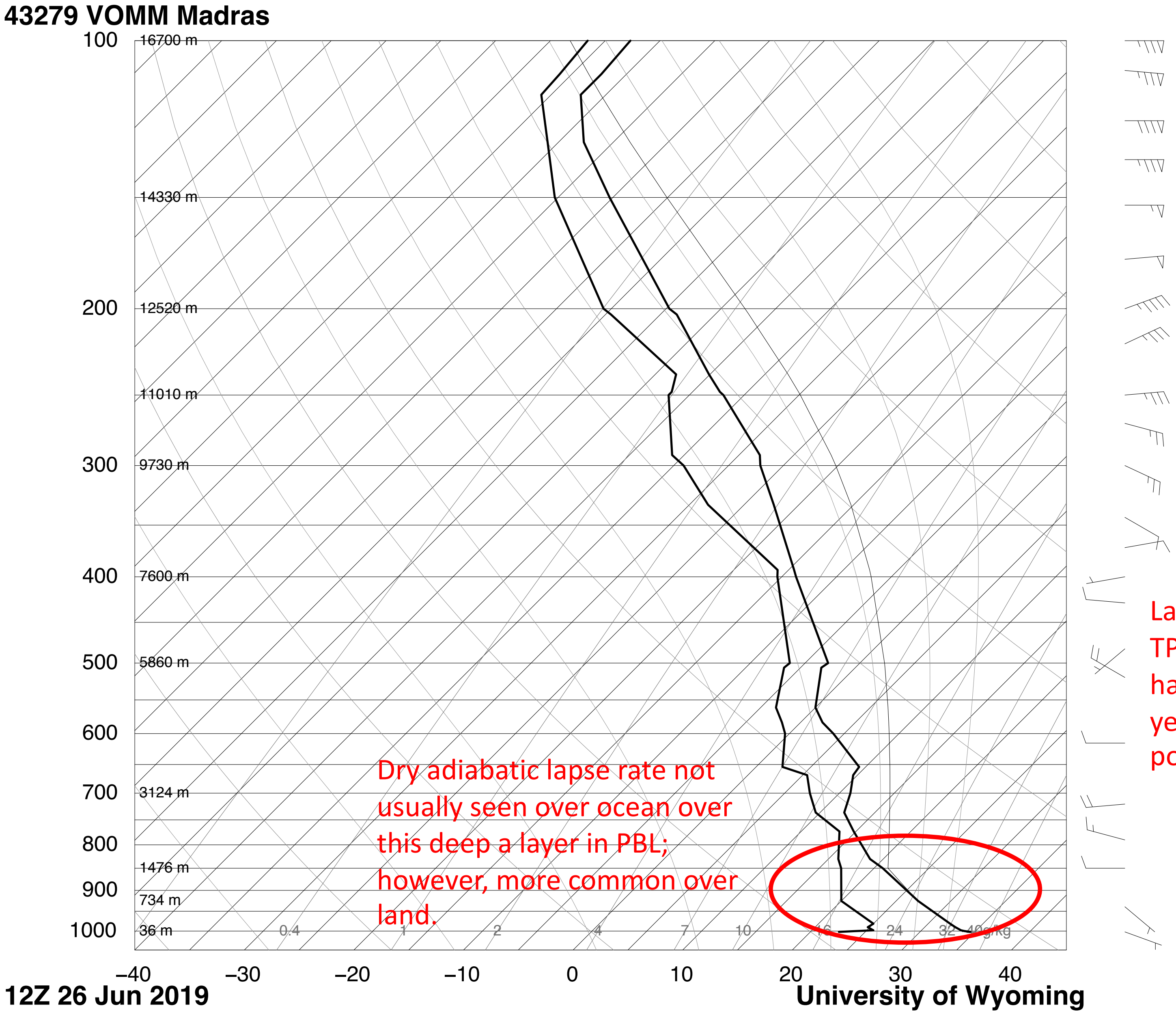


12Z 26 Jun 2019

University of Wyoming

SLAT	13.00
SLON	80.18
SELV	16.00
SHOW	−0.60
LIFT	−5.18
LFTV	−5.91
SWET	231.9
KINX	37.00
CTOT	19.90
VTOT	23.70
TOTL	43.60
CAPE	3231.
CAPV	3475.
CINS	−7.81
CINV	0.00
EQLV	111.3
EQTV	111.3
LFCT	860.2
LFCV	875.3
BRCH	68.26
BRCV	73.41
LCLT	295.4
LCLP	875.3
MLTH	306.8
MLMR	19.81
THCK	5824.
PWAT	64.34

South Asia coastal sounding: Lots of moisture and CAPE. PBL lapse rate almost dry adiabatic. High LCL/LFC though, forcing needed for deep convection.

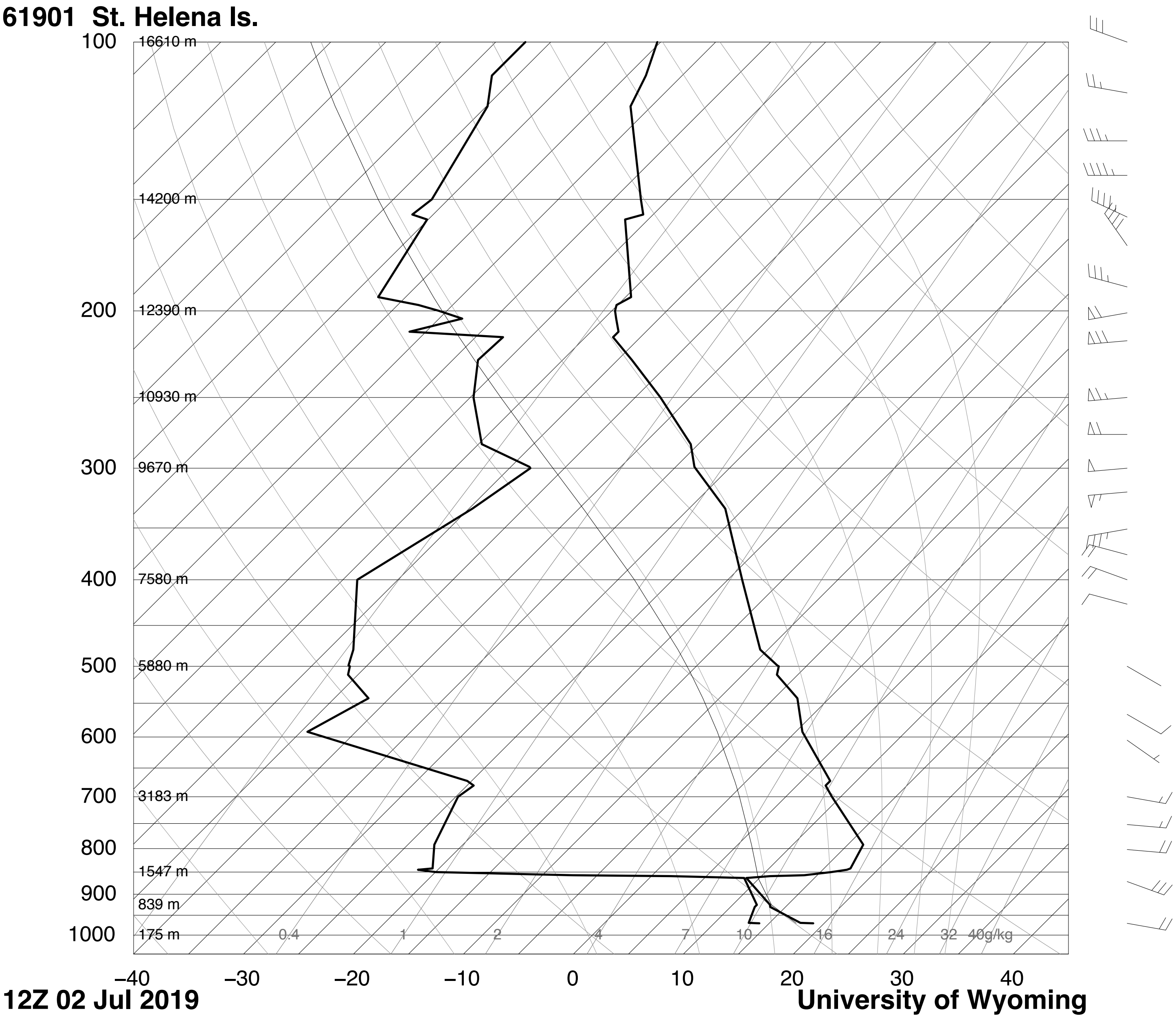


Dry adiabatic lapse rate not usually seen over ocean over this deep a layer in PBL; however, more common over land.

SLAT	13.00
SLON	80.18
SELV	16.00
SHOW	−0.60
LIFT	−5.18
LFTV	−5.91
SWET	231.9
KINX	37.00
CTOT	19.90
VTOT	23.70
TOTL	43.60
CAPE	3231.
CAPV	3475.
CINS	−7.81
CINV	0.00
EQLV	111.3
EQTV	111.3
LFCT	860.2
LFCV	875.3
BRCH	68.26
BRCV	73.41
LCLT	295.4
LCLP	875.3
MLTH	306.8
MLMR	19.81
THCK	5824.
PWAT	64.34

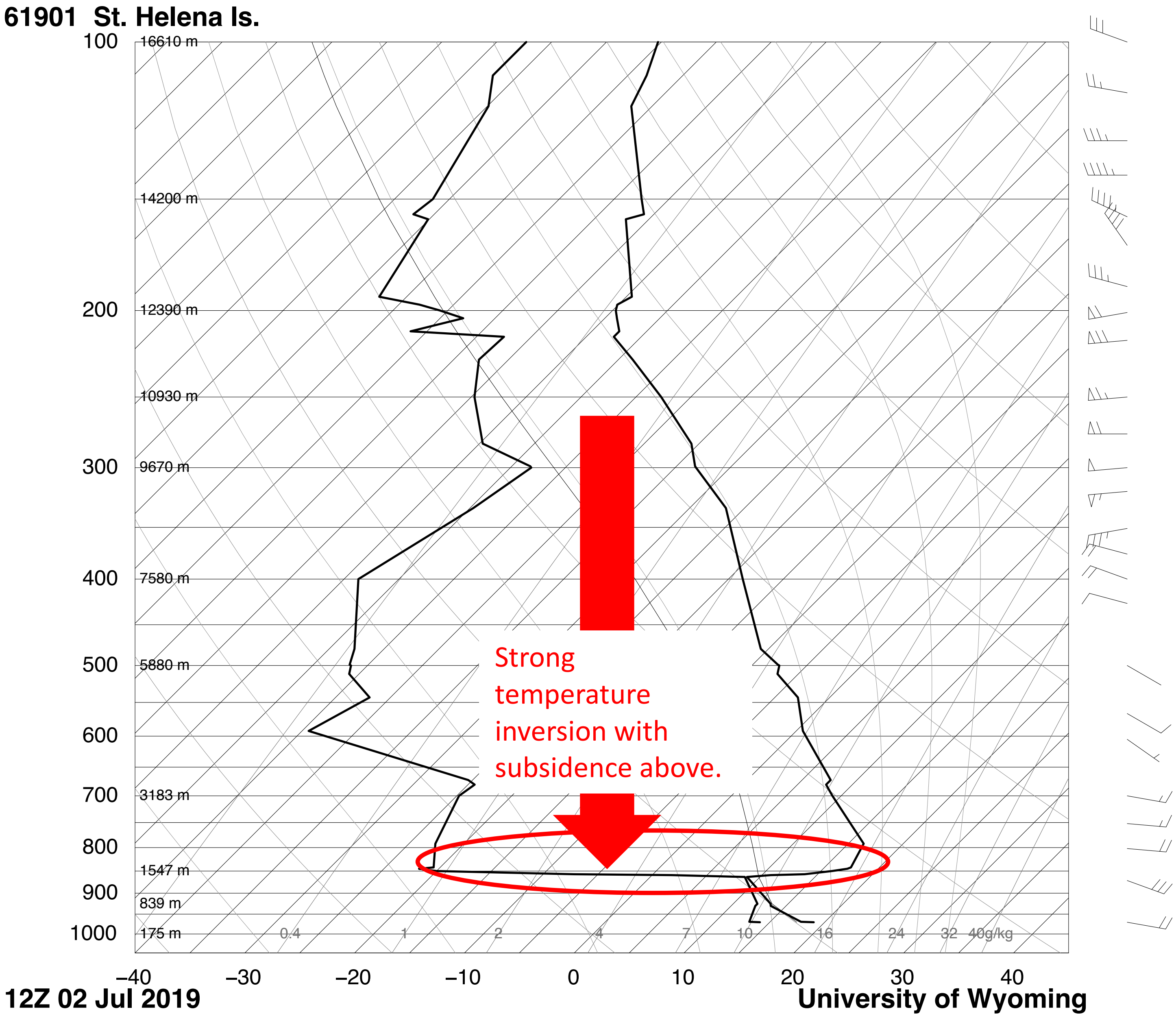
Large CAPE and TPW. Convection has not occurred yet, but it very possible.

Tropical East Atlantic sounding: Strong inversion and subsidence drying above PBL; stratocumulus common



SLAT	−15.93
SLON	−5.66
SELV	436.0
SHOW	17.41
LIFT	8.33
LFTV	8.00
SWET	54.99
KINX	−30.5
CTOT	−12.5
VTOT	23.50
TOTL	11.00
CAPE	5.52
CAPV	7.43
CINS	−0.27
CINV	0.00
EQLV	861.0
EQTV	860.1
LFCT	900.4
LFCV	903.3
BRCH	1.18
BRCV	1.60
LCLT	285.0
LCLP	906.6
MLTH	293.1
MLMR	9.74
THCK	5705.
PWAT	12.77

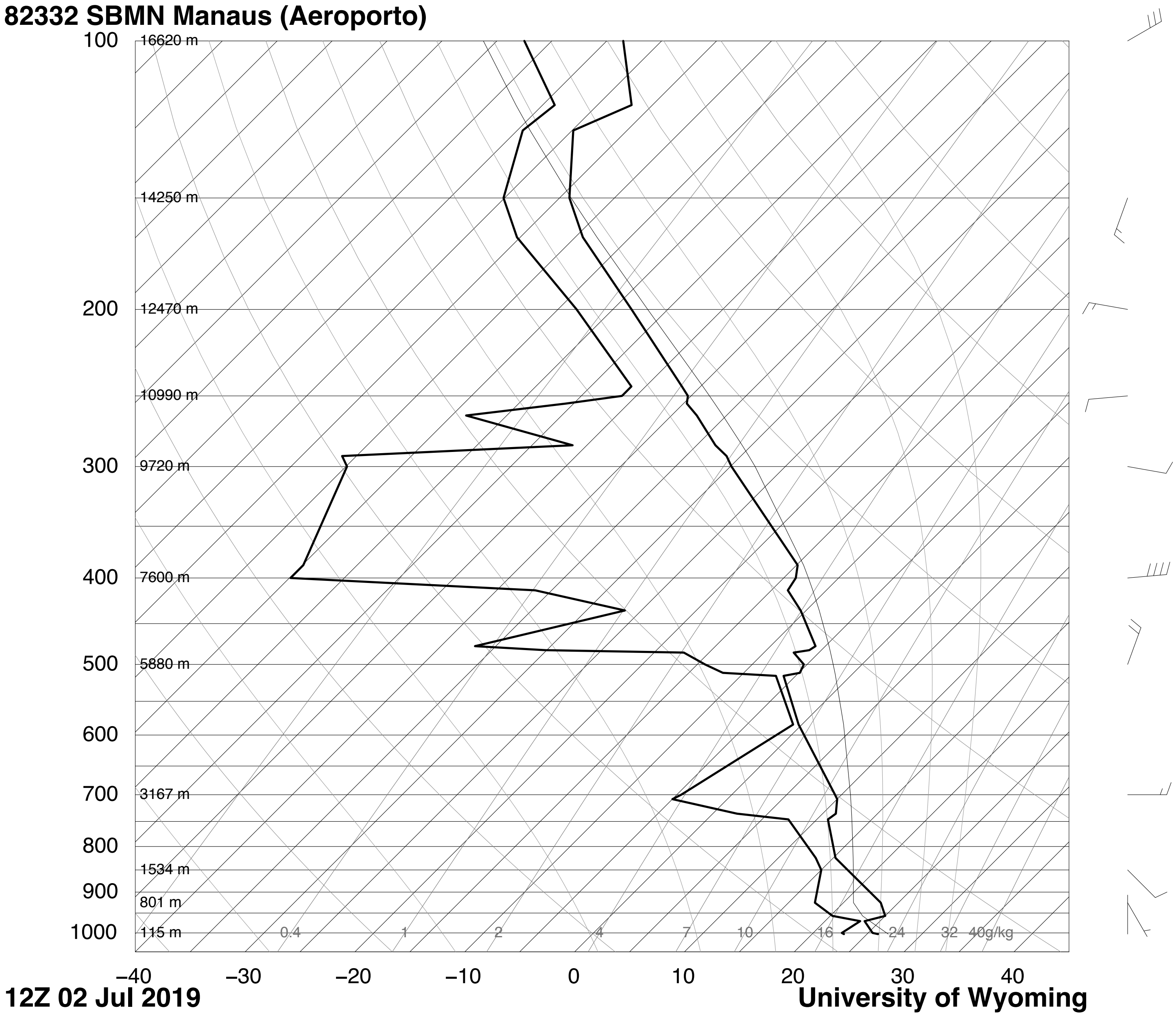
Tropical East Atlantic sounding: Strong inversion and subsidence drying above PBL; stratocumulus common



Strong temperature inversion with subsidence above.

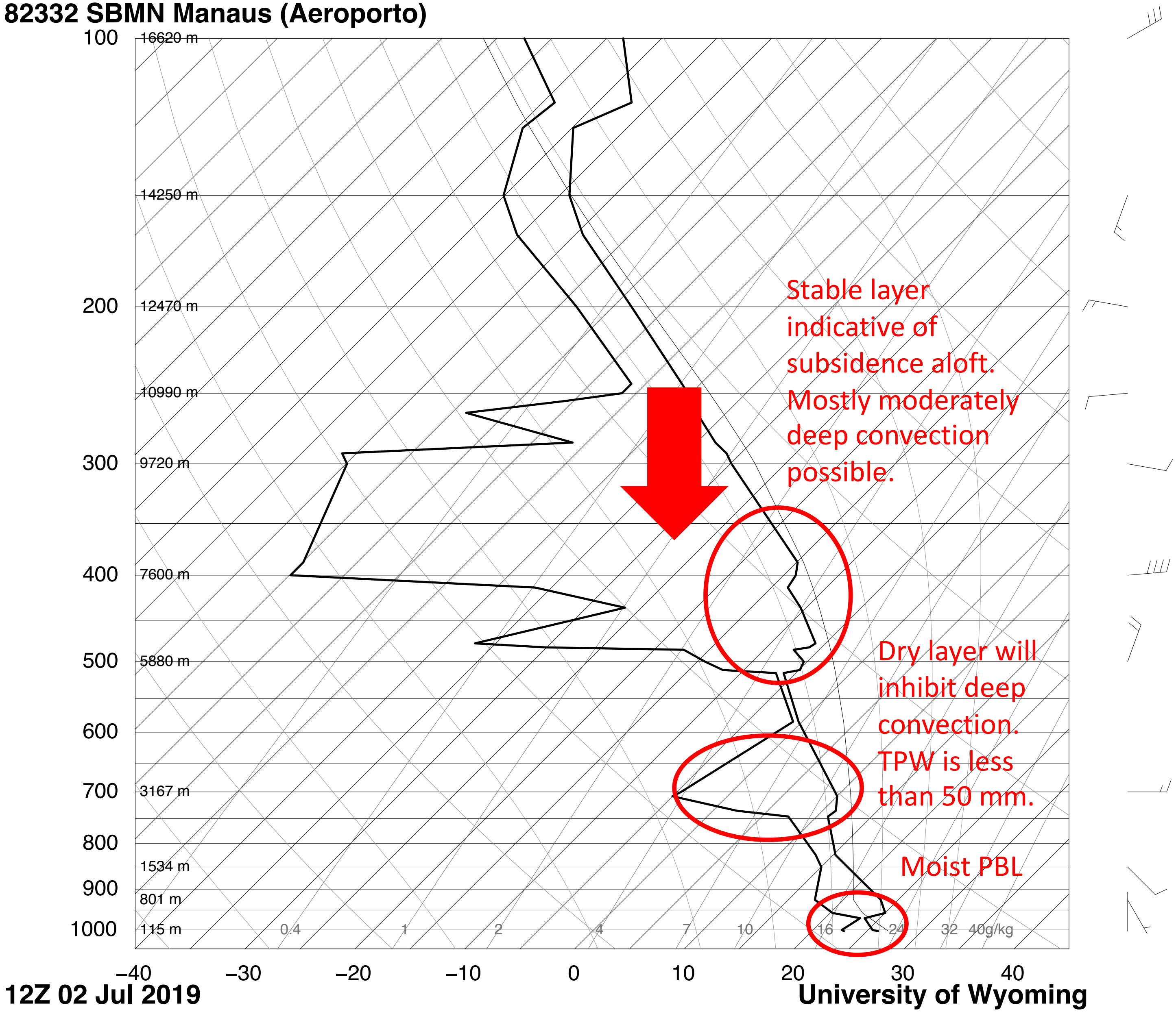
SLAT	−15.93
SLON	−5.66
SELV	436.0
SHOW	17.41
LIFT	8.33
LFTV	8.00
SWET	54.99
KINX	−30.5
CTOT	−12.5
VTOT	23.50
TOTL	11.00
CAPE	5.52
CAPV	7.43
CINS	−0.27
CINV	0.00
EQLV	861.0
EQTV	860.1
LFCT	900.4
LFCV	903.3
BRCH	1.18
BRCV	1.60
LCLT	285.0
LCLP	906.6
MLTH	293.1
MLMR	9.74
THCK	5705.
PWAT	12.77

Amazonian sounding: Can be much moister than what is shown below. Rainforest acts like “green ocean” to keep PBL moist.



SLAT	−3.15
SLON	−59.98
SELV	84.00
SHOW	0.52
LIFT	−2.68
LFTV	−3.31
SWET	220.0
KINX	23.70
CTOT	20.30
VTOT	22.70
TOTL	43.00
CAPE	988.5
CAPV	1180.
CINS	−43.1
CINV	−29.7
EQLV	149.0
EQTV	149.0
LFCT	862.5
LFCV	875.1
BRCH	44.40
BRCV	53.01
LCLT	294.6
LCLP	941.0
MLTH	299.8
MLMR	17.56
THCK	5765.
PWAT	48.64

Amazonian sounding: Can be much moister than what is shown below. Rainforest acts like “green ocean” to keep PBL moist.



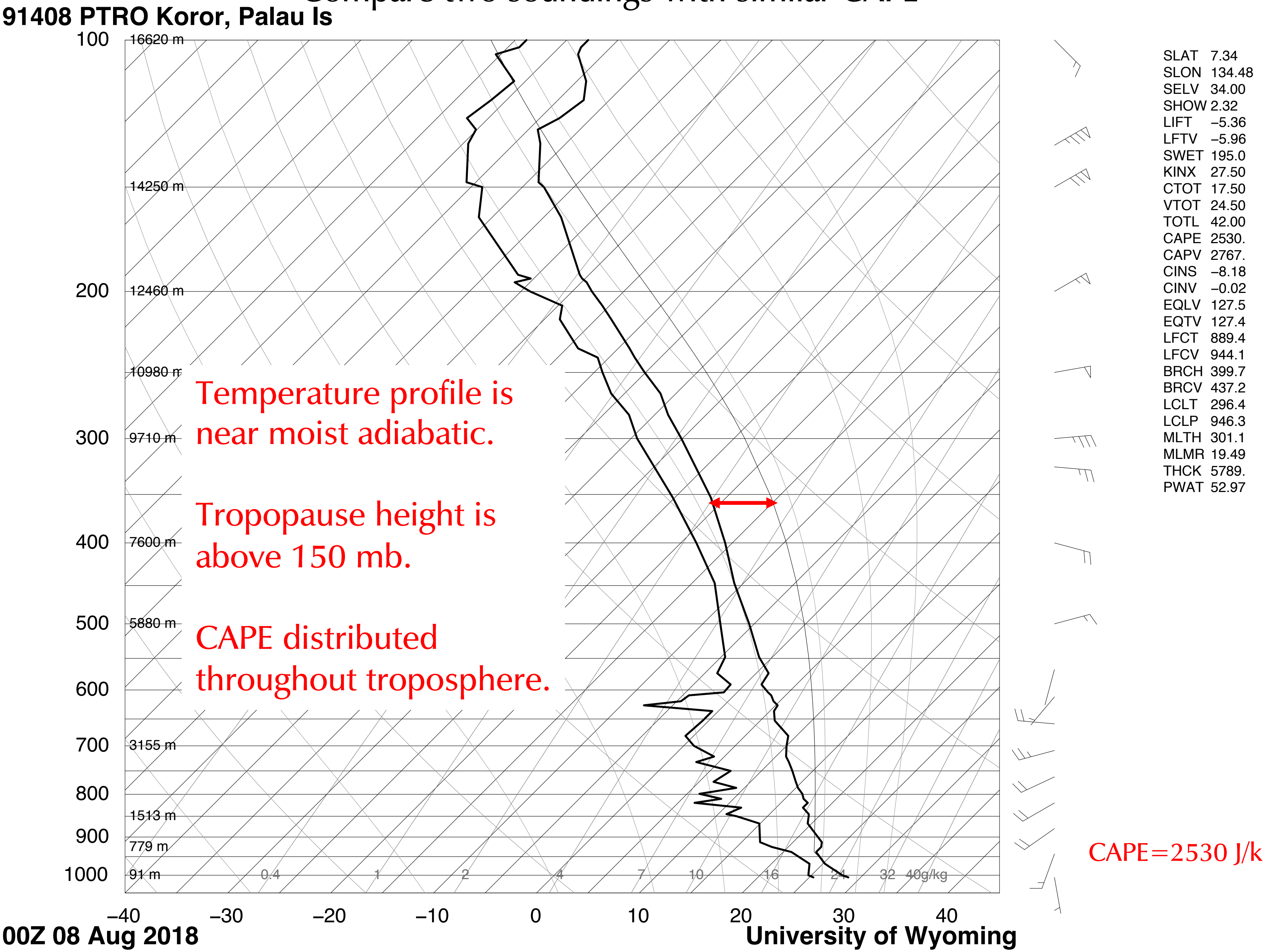
Stable layer indicative of subsidence aloft. Mostly moderately deep convection possible.

Dry layer will inhibit deep convection. TPW is less than 50 mm.

Moist PBL

SLAT	−3.15
SLON	−59.98
SELV	84.00
SHOW	0.52
LIFT	−2.68
LFTV	−3.31
SWET	220.0
KINX	23.70
CTOT	20.30
VTOT	22.70
TOTL	43.00
CAPE	988.5
CAPV	1180.
CINS	−43.1
CINV	−29.7
EQLV	149.0
EQTV	149.0
LFCT	862.5
LFCV	875.1
BRCH	44.40
BRCV	53.01
LCLT	294.6
LCLP	941.0
MLTH	299.8
MLMR	17.56
THCK	5765.
PWAT	48.64

Compare two soundings with similar CAPE



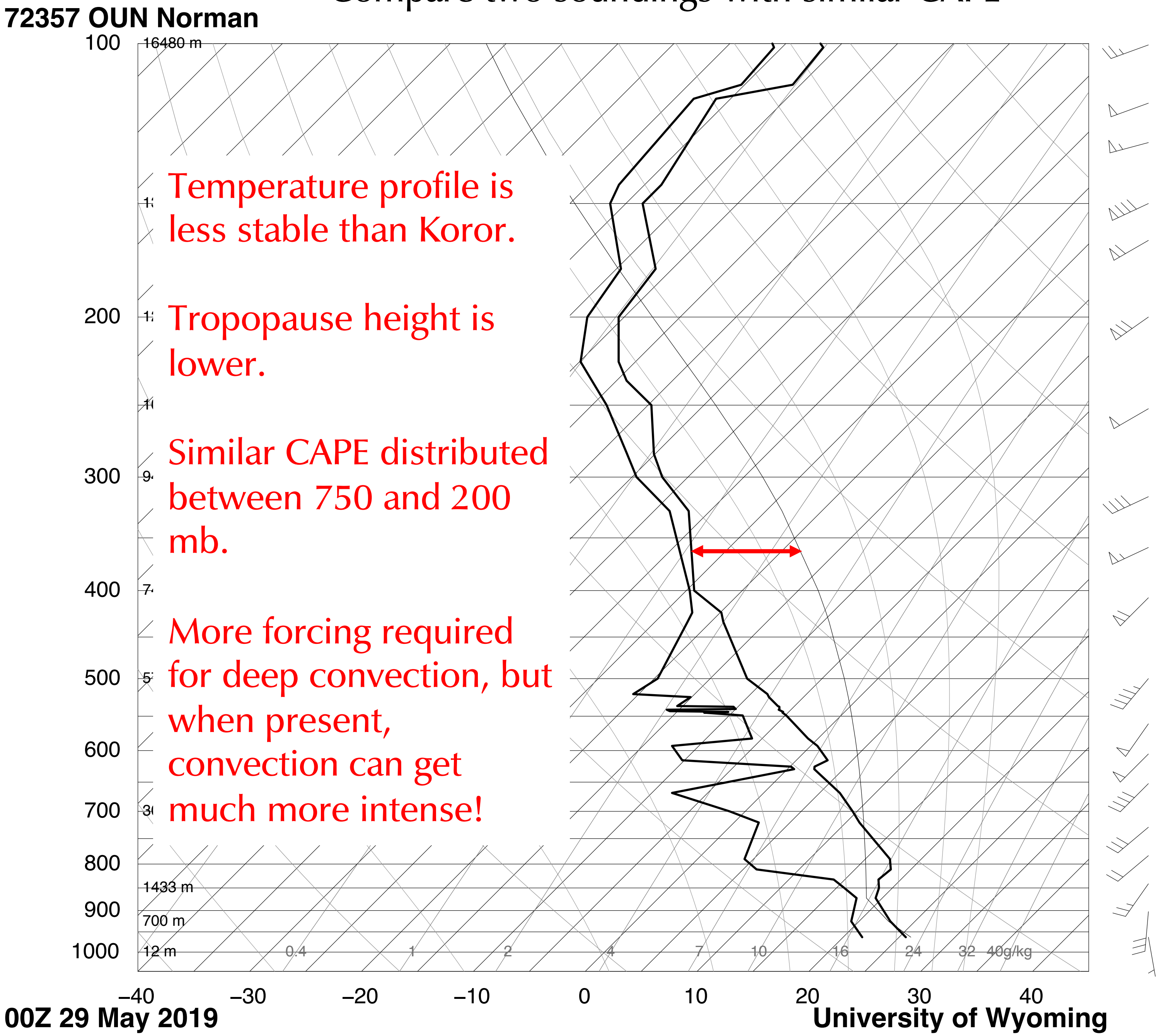
Temperature profile is near moist adiabatic.

Tropopause height is above 150 mb.

CAPE distributed throughout troposphere.

CAPE=2530 J/kg

Compare two soundings with similar CAPE



Temperature profile is less stable than Koror.

Tropopause height is lower.

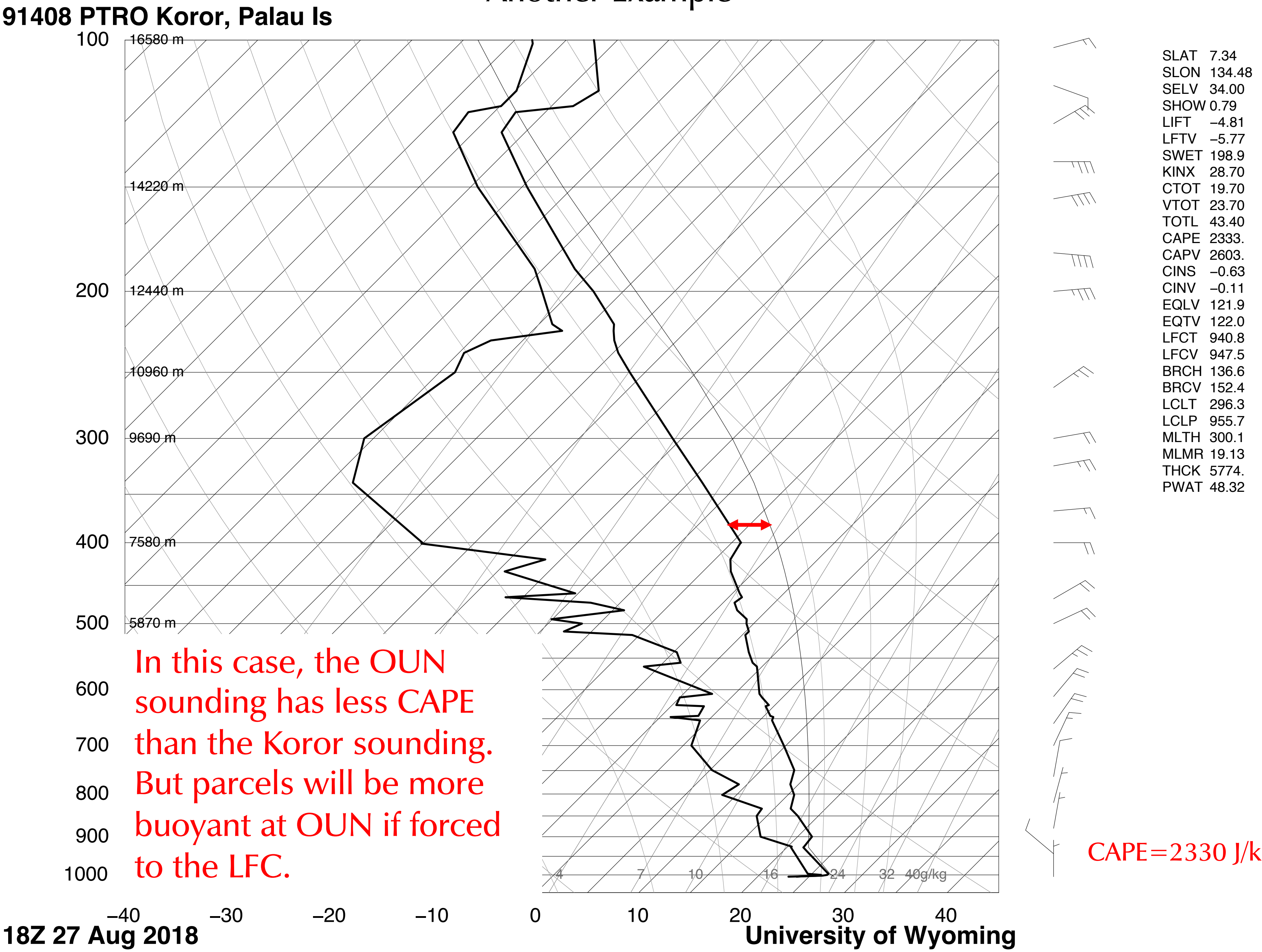
Similar CAPE distributed between 750 and 200 mb.

More forcing required for deep convection, but when present, convection can get much more intense!

SLAT	35.18
SLON	−97.44
SELV	345.0
SHOW	−7.23
LIFT	−8.71
LFTV	−9.46
SWET	519.7
KINX	35.20
CTOT	27.40
VTOT	30.50
TOTL	57.90
CAPE	2581.
CAPV	2743.
CINS	−79.7
CINV	−39.0
EQLV	188.7
EQTV	188.6
LFCT	730.1
LFCV	761.2
BRCH	39.11
BRCV	41.56
LCLT	292.4
LCLP	888.4
MLTH	302.4
MLMR	16.10
THCK	5758.
PWAT	39.40

CAPE=2581 J/kg

Another Example

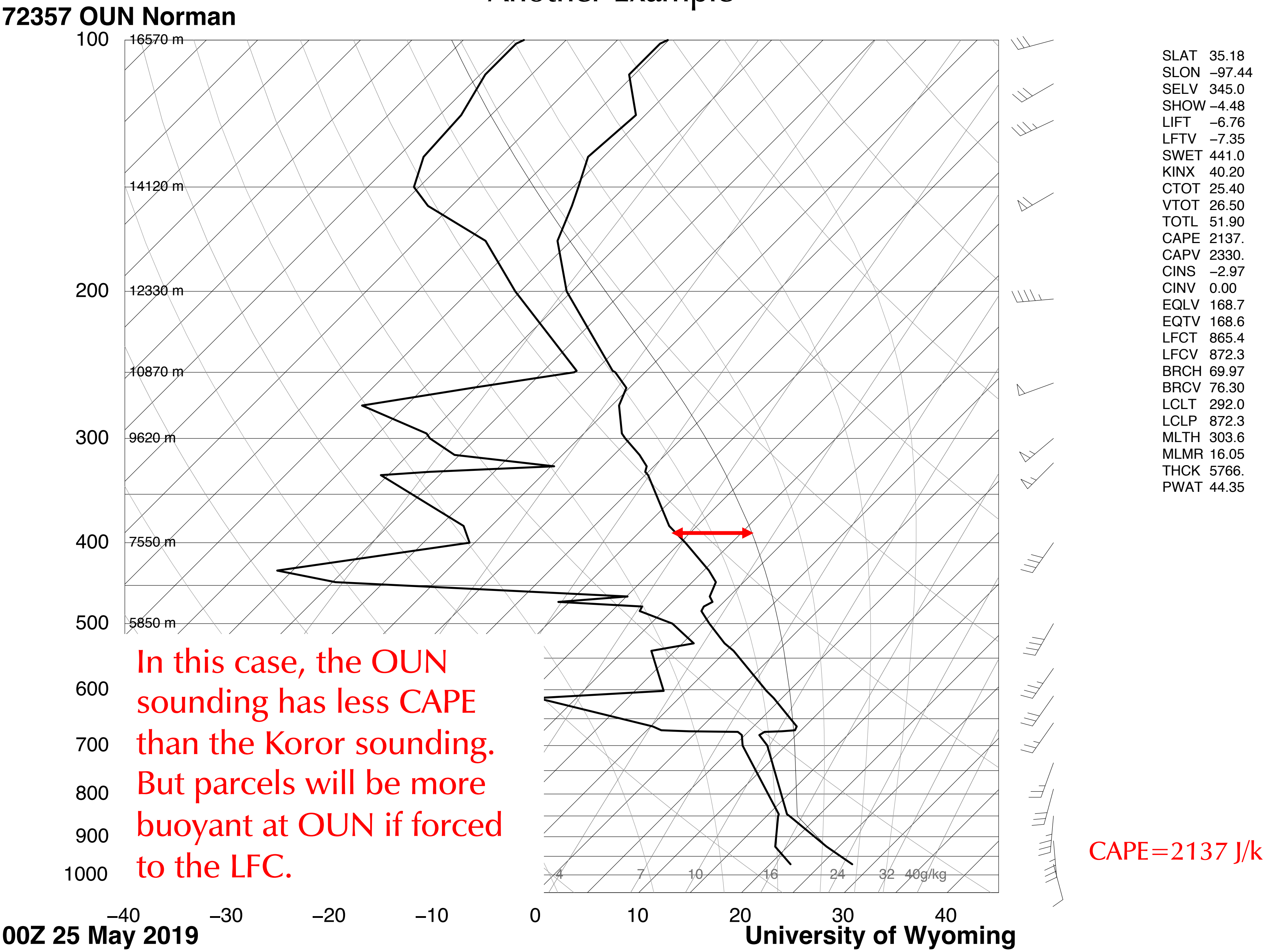


SLAT	7.34
SLON	134.48
SELV	34.00
SHOW	0.79
LIFT	−4.81
LFTV	−5.77
SWET	198.9
KINX	28.70
CTOT	19.70
VTOT	23.70
TOTL	43.40
CAPE	2333.
CAPV	2603.
CINS	−0.63
CINV	−0.11
EQLV	121.9
EQTV	122.0
LFCT	940.8
LFCV	947.5
BRCH	136.6
BRCV	152.4
LCLT	296.3
LCLP	955.7
MLTH	300.1
MLMR	19.13
THCK	5774.
PWAT	48.32

In this case, the OUN sounding has less CAPE than the Koror sounding. But parcels will be more buoyant at OUN if forced to the LFC.

CAPE=2330 J/kg

Another Example



SLAT	35.18
SLON	−97.44
SELV	345.0
SHOW	−4.48
LIFT	−6.76
LFTV	−7.35
SWET	441.0
KINX	40.20
CTOT	25.40
VTOT	26.50
TOTL	51.90
CAPE	2137.
CAPV	2330.
CINS	−2.97
CINV	0.00
EQLV	168.7
EQTV	168.6
LFCT	865.4
LFCV	872.3
BRCH	69.97
BRCV	76.30
LCLT	292.0
LCLP	872.3
MLTH	303.6
MLMR	16.05
THCK	5766.
PWAT	44.35

In this case, the OUN sounding has less CAPE than the Koror sounding. But parcels will be more buoyant at OUN if forced to the LFC.

CAPE=2137 J/kg

MR3252: Tropical Meteorology

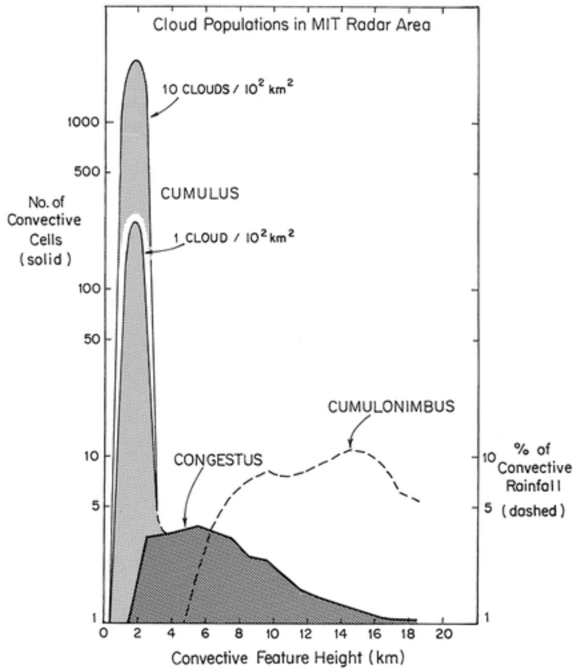
Modes of Cumuliform Convection

Main Topics:

- Shallow cumulus, congestus, deep convection
- Radiative-convective equilibrium
- Large-scale subsidence

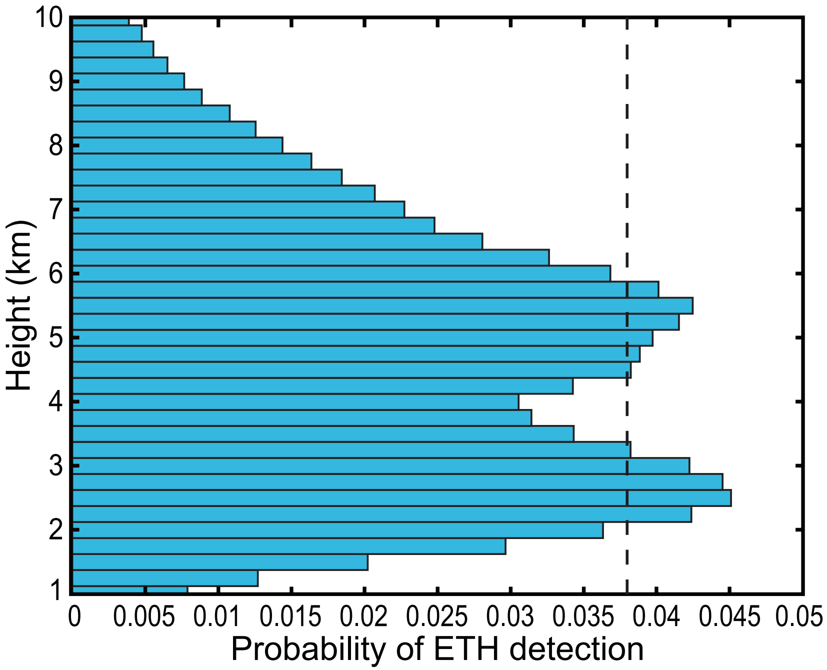
Trimodal distribution of convection

Radar data during TOGA COARE



Johnson et al. (1999)

In TRMM data over TWP: Distribution of 20 dBZ Echo Top Heights (Moderate and Deep modes visible)

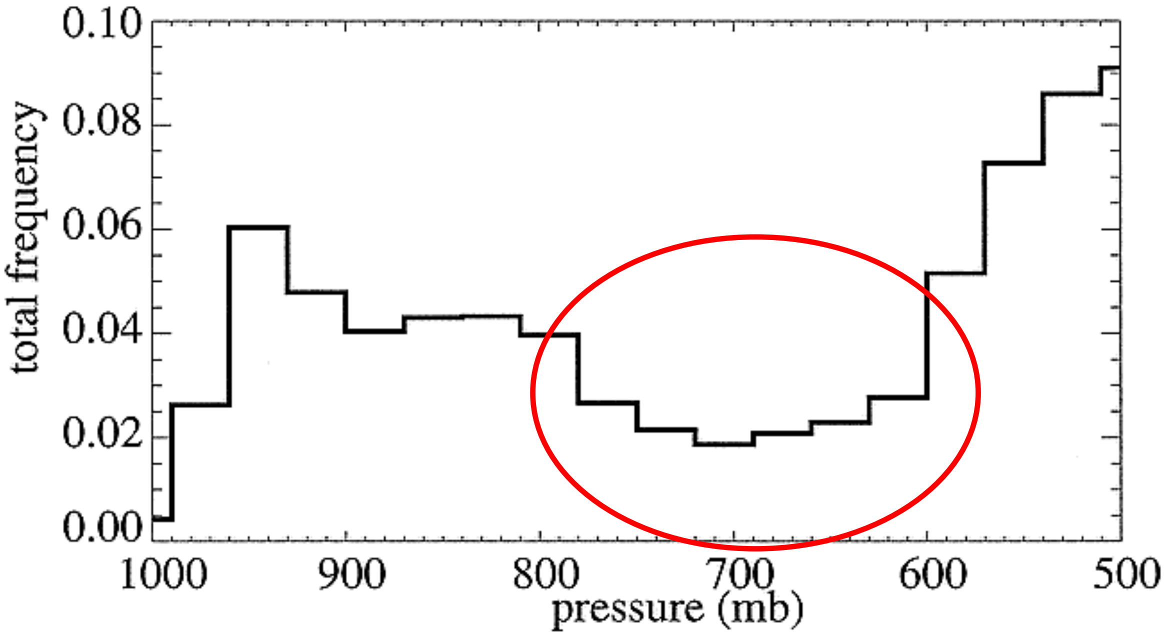


Powell and Houze (2015)

Shallow convection

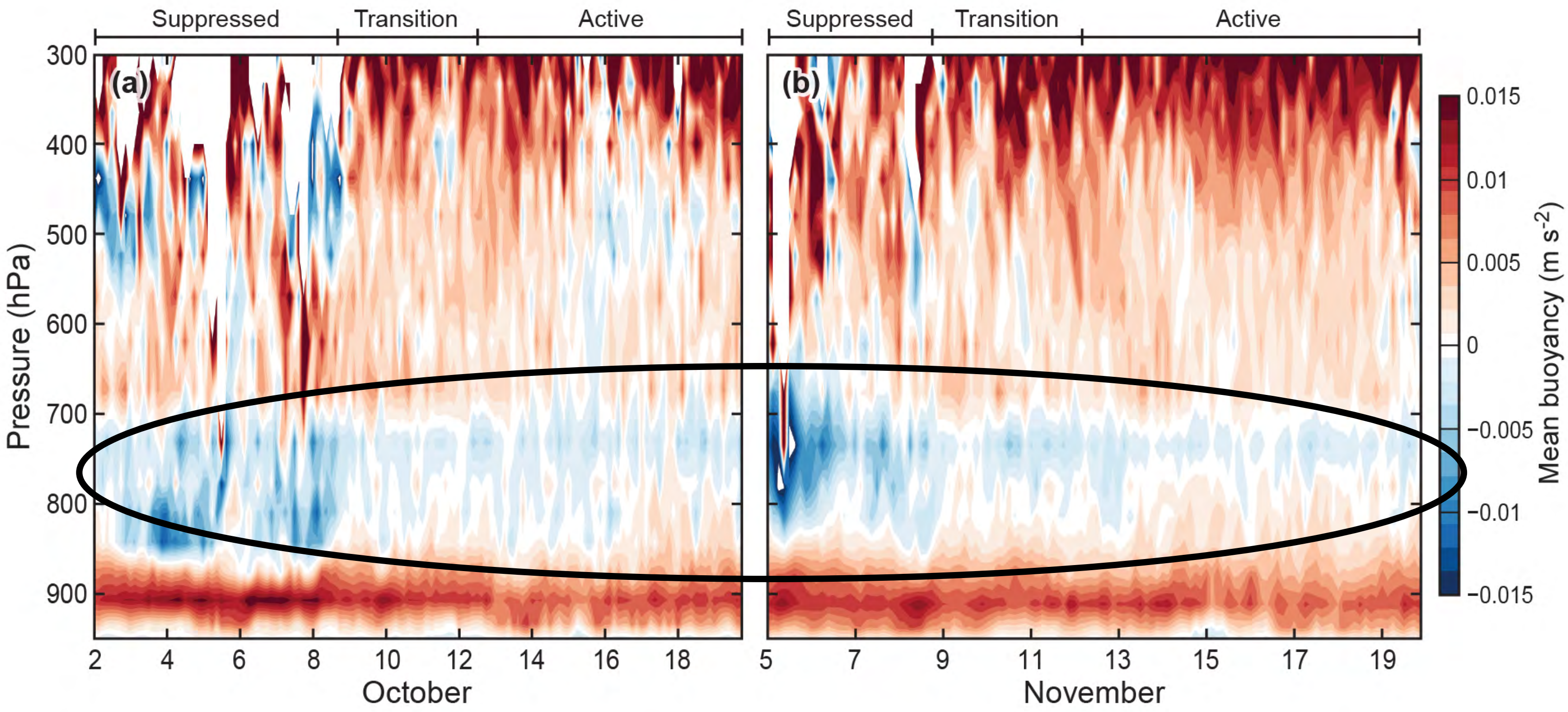
- Common when free troposphere is dry
- In some locations, often seen with large subsidence signal aloft, indicating an inversion often seen in the trade wind regime. Thus, sometimes these are called “trade cumuli”

Zuidema (1998): estimated using soundings



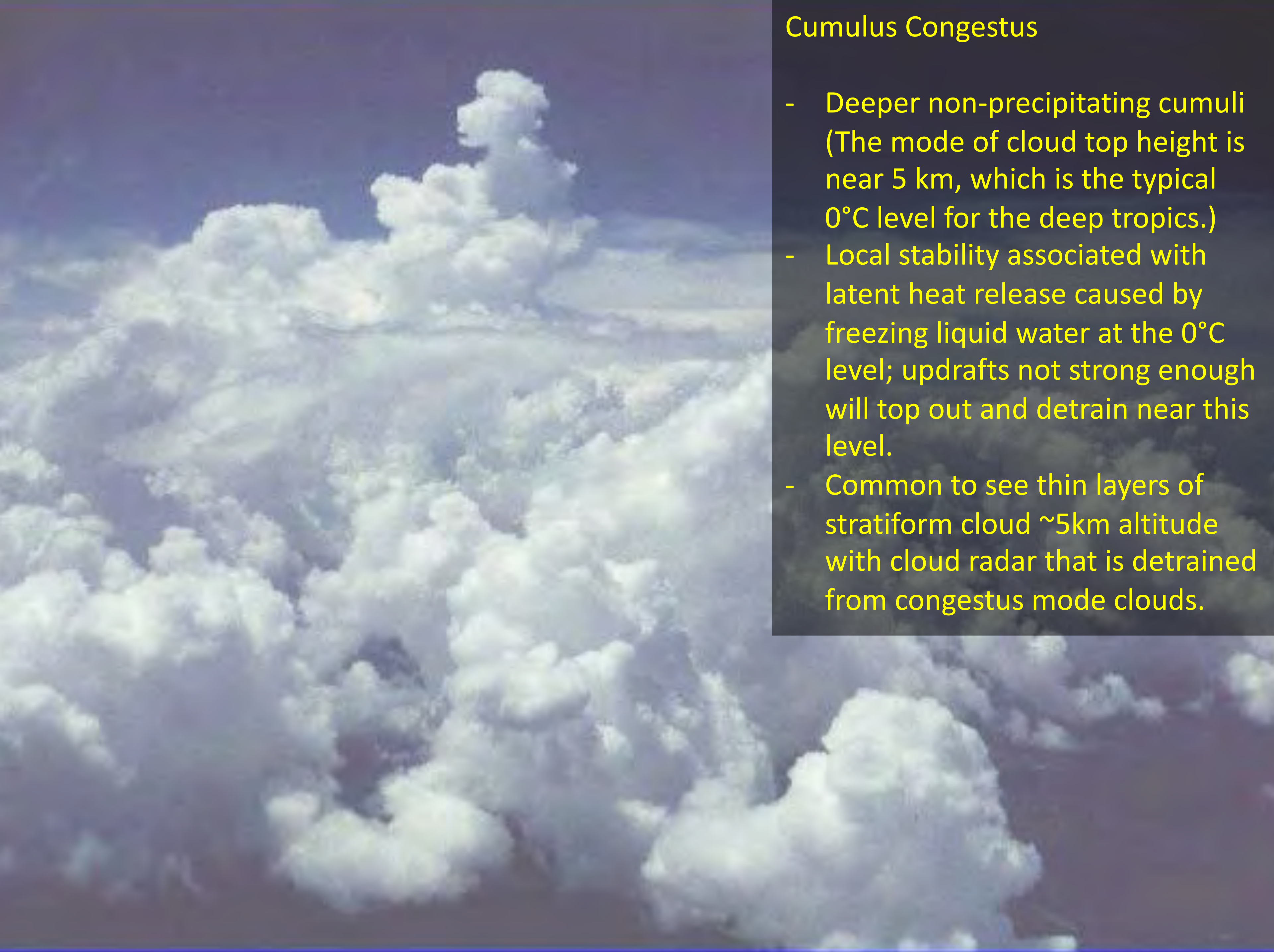
Minimum of clouds above boundary layer

Powell (2016): model of warm pool convection



Negative buoyancy typically present between about 700 and 850 mb.

$$B \approx g\left(\frac{T^*}{\underbrace{T_e}_{\text{Temperature}}} - \frac{\overbrace{p^*}^{\text{Pressure}}}{p_e} + \underbrace{0.608\left(w^*\right)}_{\text{Vapor}} - \overbrace{w_H}^{\text{Hydrometeor}}\right)$$



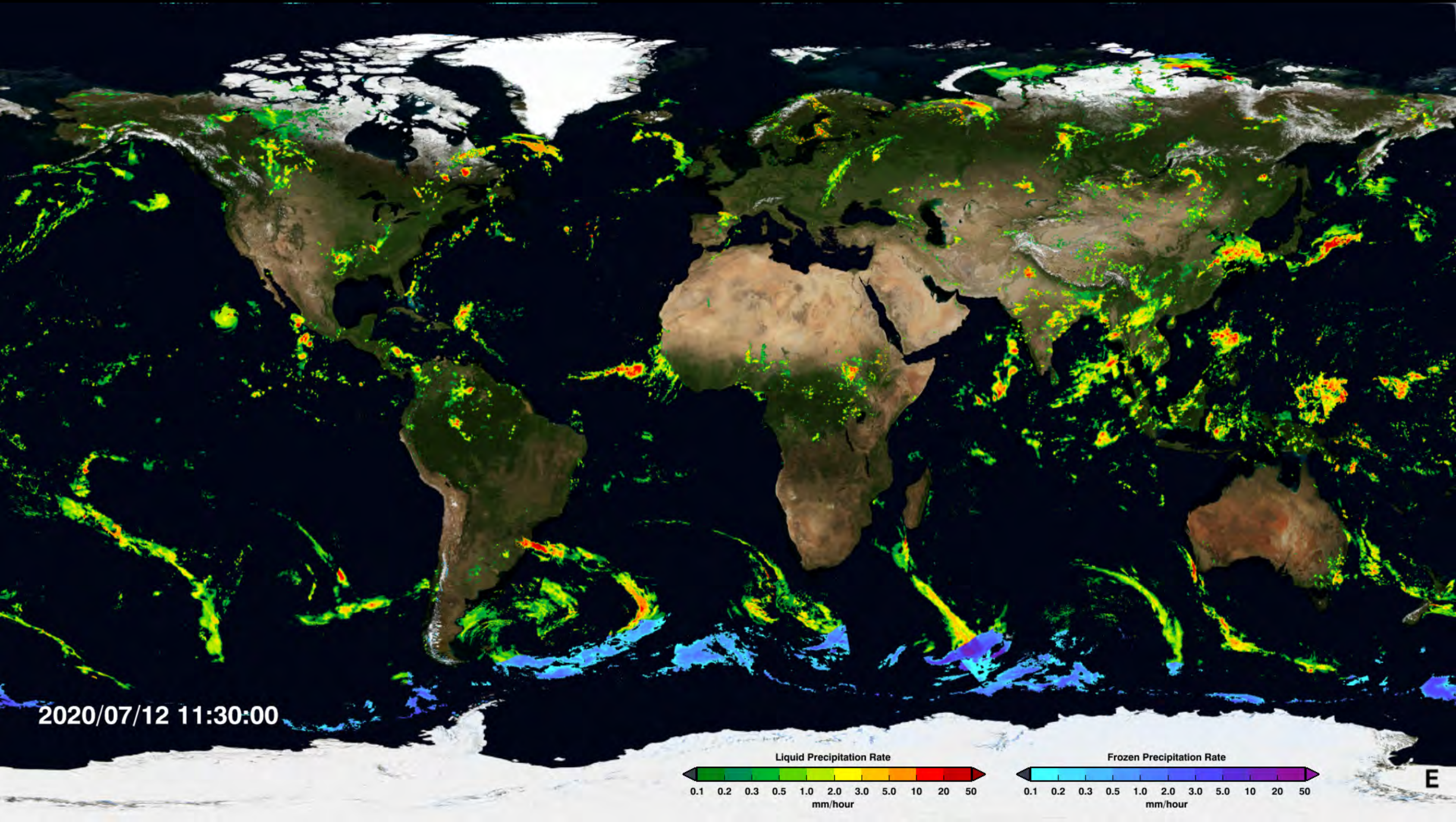
Cumulus Congestus

- Deeper non-precipitating cumuli (The mode of cloud top height is near 5 km, which is the typical 0°C level for the deep tropics.)
- Local stability associated with latent heat release caused by freezing liquid water at the 0°C level; updrafts not strong enough will top out and detrain near this level.
- Common to see thin layers of stratiform cloud ~5km altitude with cloud radar that is detrained from congestus mode clouds.

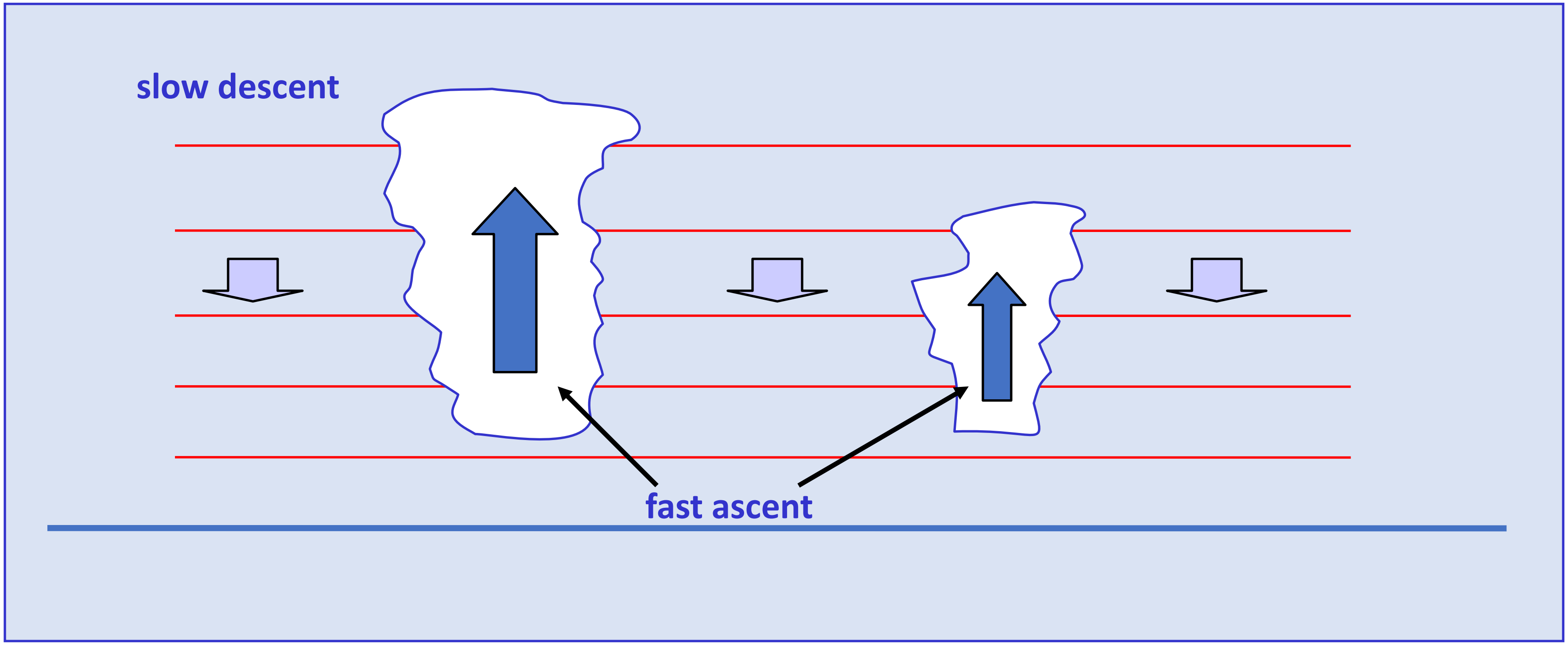
Deep Convection

- Either has strong vertical motion exiting boundary layer and/or develops in a favorable environment.
- In the deep Tropics, can sometimes reach altitudes of up to 20 km, but ultimately capped by stratospheric inversion.

At any given time, rain occurs only over a small area!



Subsidence occurs in clear-air outside clouds

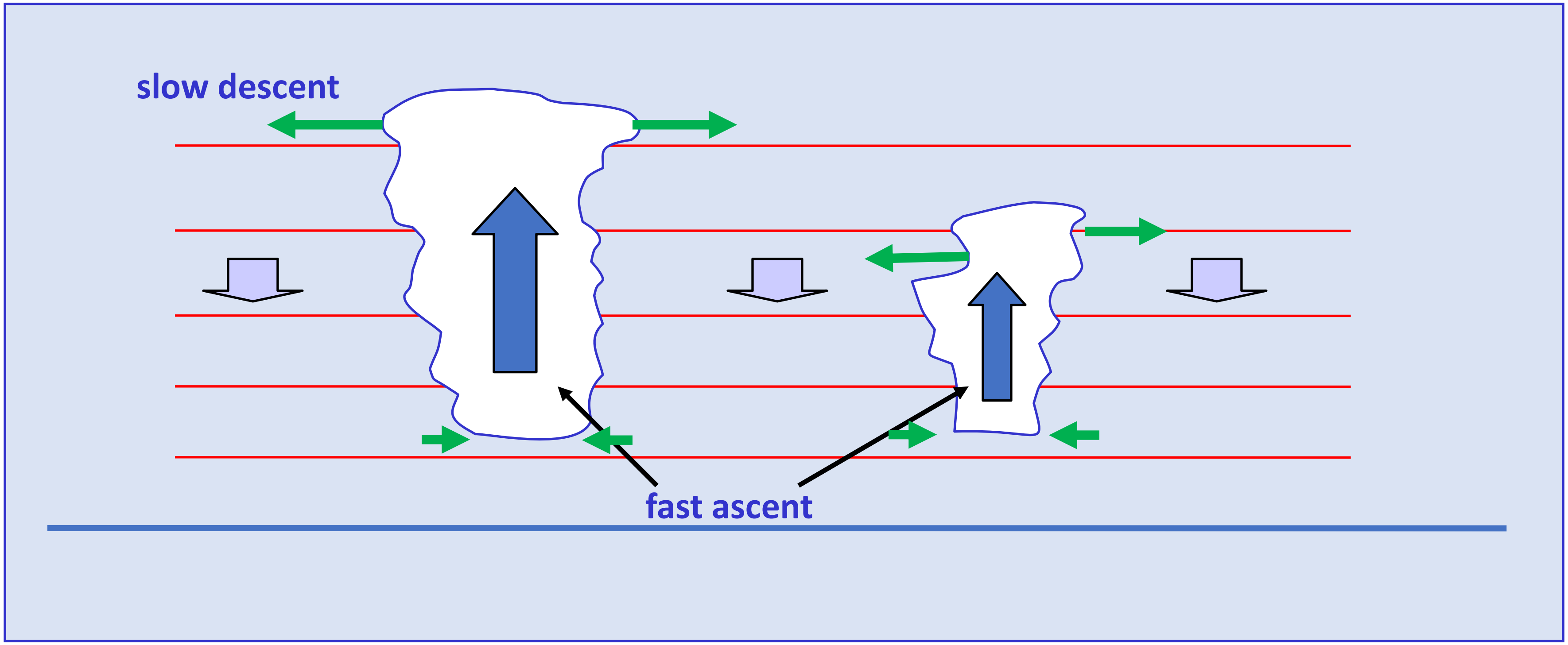


In subsiding regions, adiabatic heating occurs. Because downward motion dries the column, more radiative cooling occurs.

In convection, adiabatic cooling occurs, but is countered by latent heat release. The radiative impacts of clouds are more complicated to quantify.

Statistically on time scales longer than that of individual clouds, convection and the environment are in approximate **radiative-convective equilibrium.**

Subsidence occurs in clear-air outside clouds



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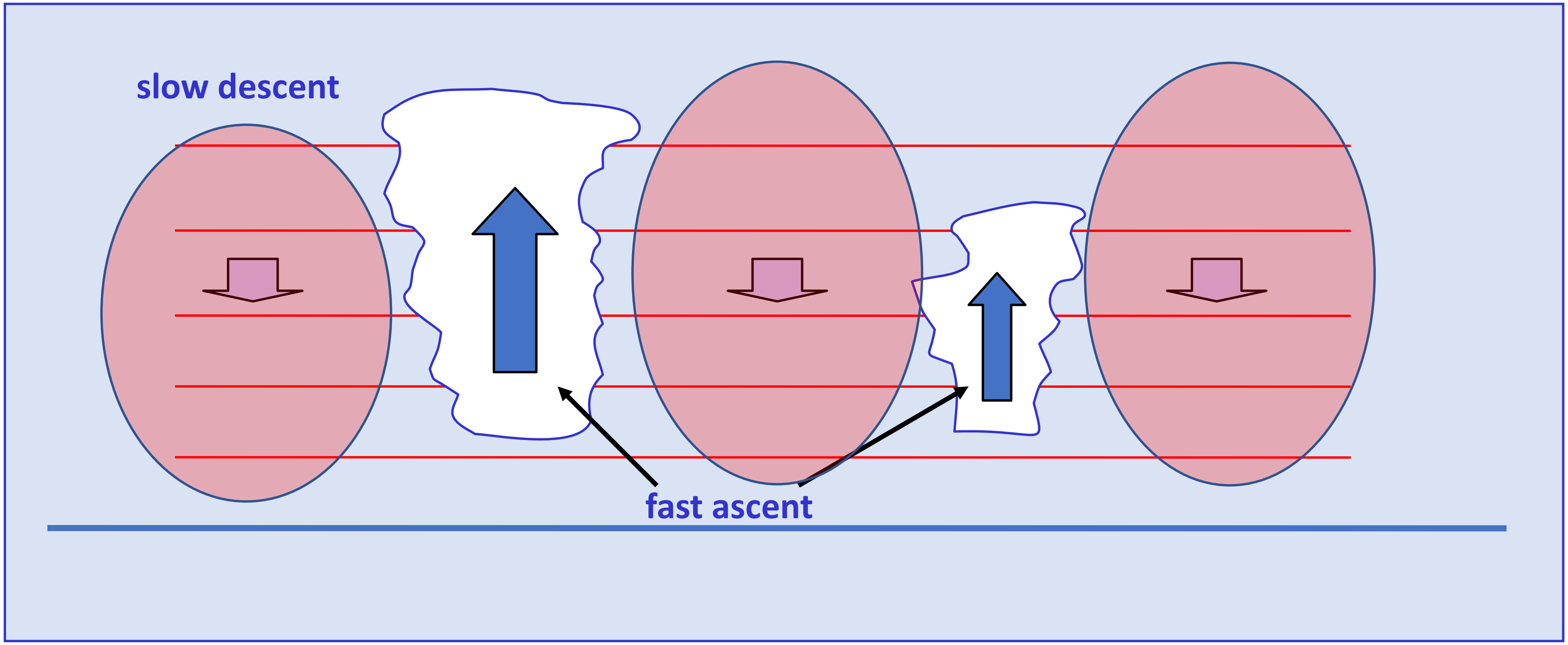

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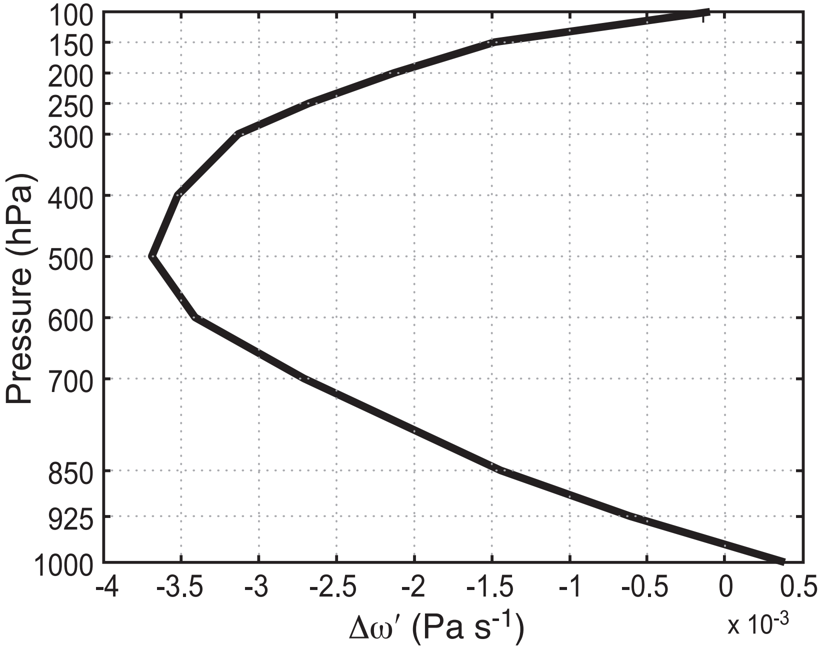
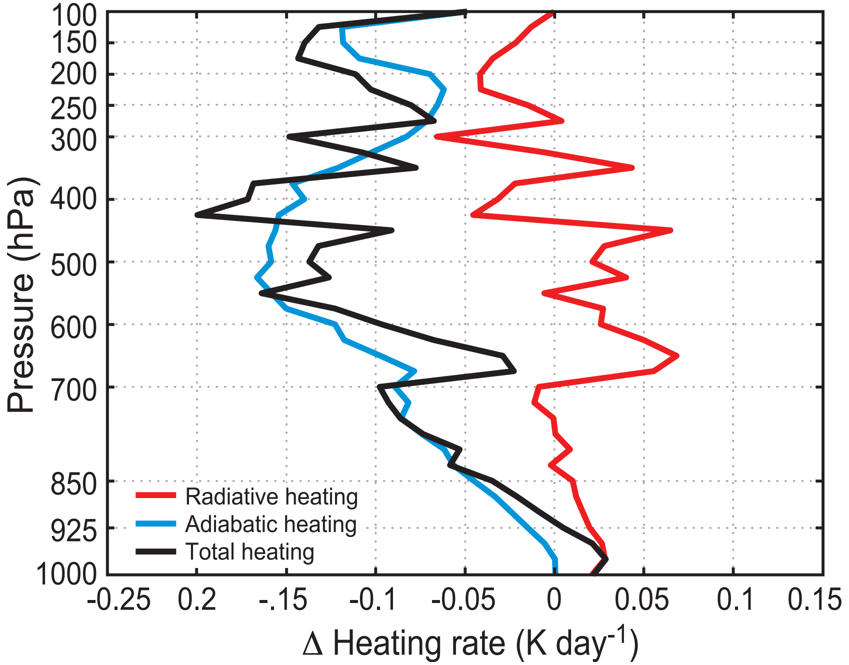
Statistically on time scales longer than that of individual clouds, convection and the environment are in approximate **radiative-convective equilibrium.**

Reducing subsidence can make large-scale environment more conducive to convection.



Change in large-scale vertical motion



Corresponding rate of change in temperature due to adiabatic and radiative processes.

However, do not automatically interpret large-scale upward motion as favorable for convection! It generally means that convection is already present (the upward motion is in the convection itself), and convection reduces conditional instability in the environment.

MR3252: Tropical Meteorology

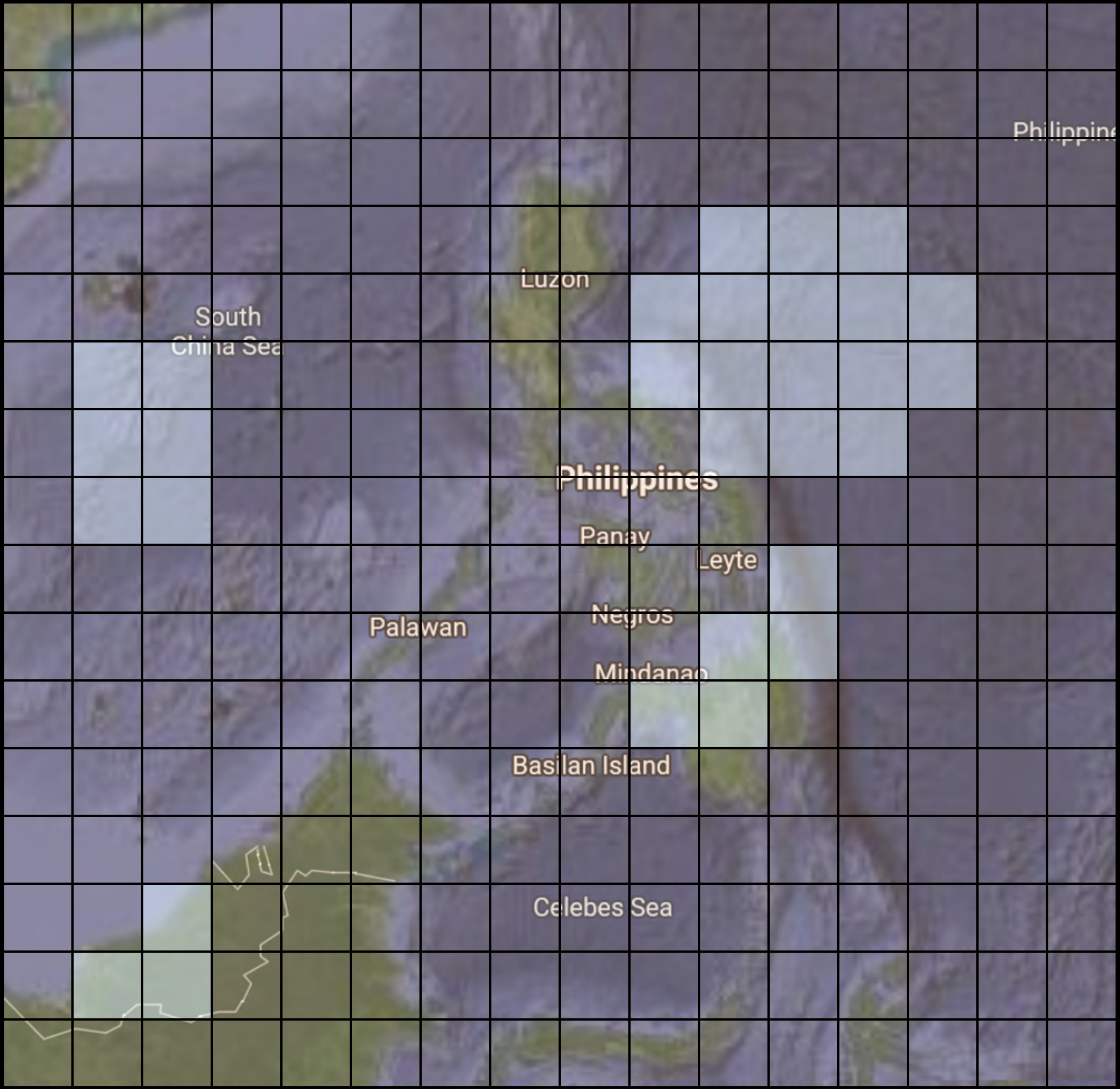
Large-Scale Tropical Heating and Moisture Budgets

Main Topics:

- Large-scale heat sources and moisture sinks
- Relationship of large-scale heating to precipitation and fluxes

Module 1.7


Philippine
Luzon
South
China Sea
Philippines
Panay
Leyte
Negros
Palawan
Mindanao
Basilan Island
Celebes Sea


Philippine
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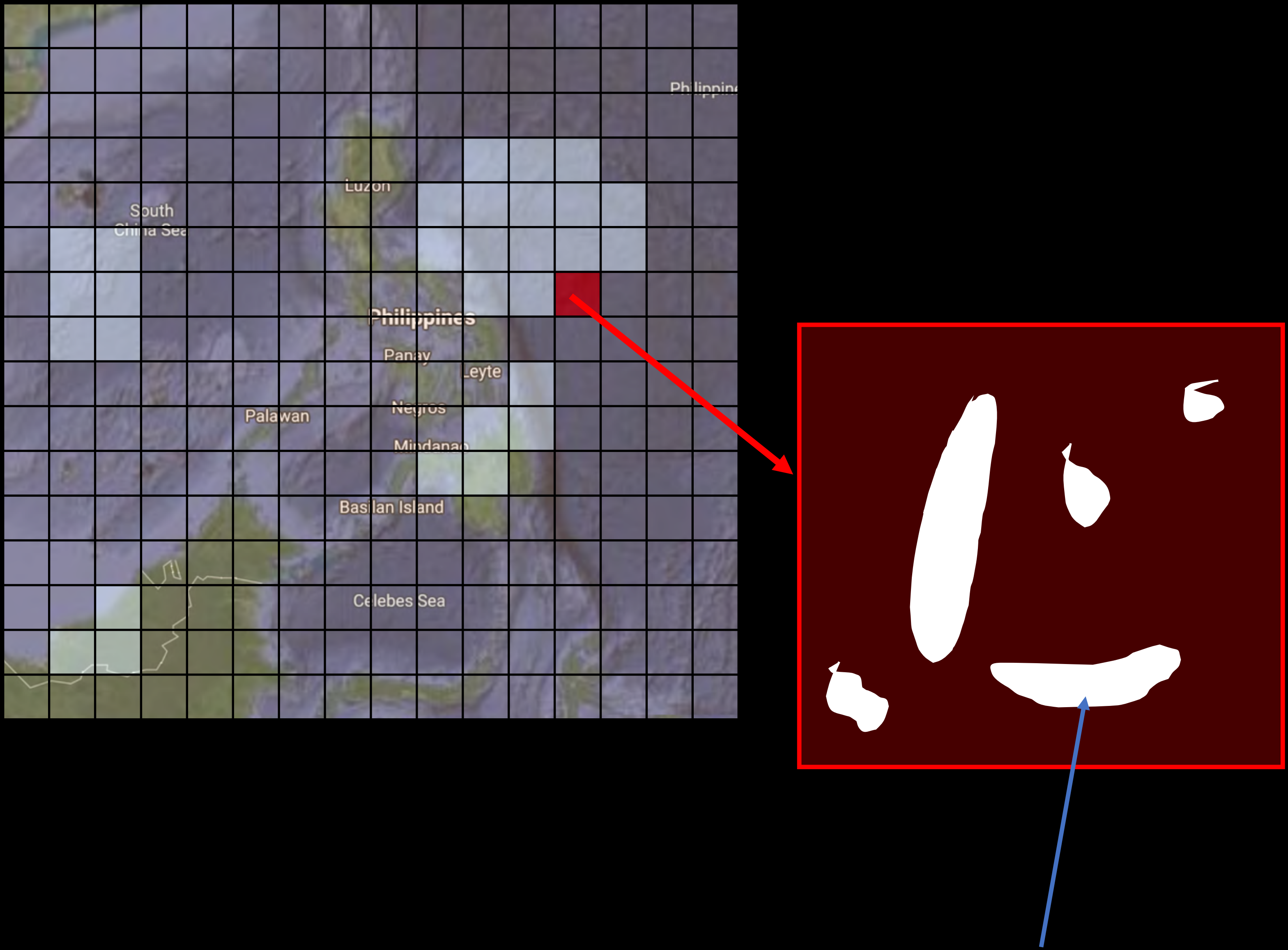
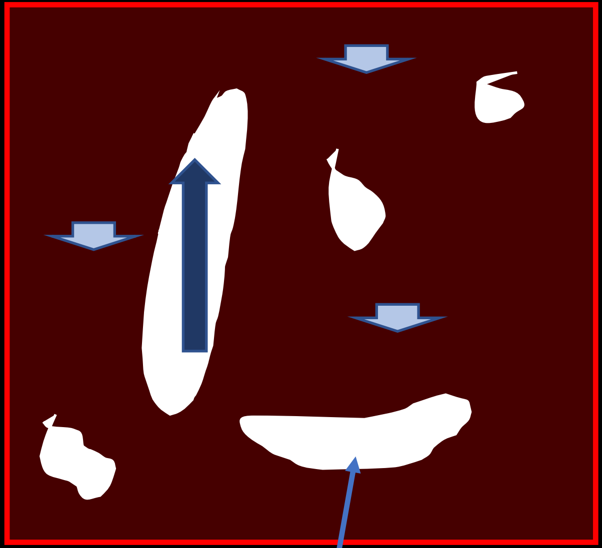


Slices through clouds

$$w = \overline{w} + w'$$

w', q', T' $\overline{w}, \overline{q}, \overline{T}$



Slices through clouds

Large-scale budget equations

$$\frac{\partial \bar{s}}{\partial t} + \overline{\nabla \cdot s\mathbf{V}} + \frac{\partial \overline{s\omega}}{\partial p} = Q_R + Q_c$$

$$Q_C = L_v(\bar{c} - \bar{e})$$

s = DSE, h = MSE

$$\frac{\partial \bar{q}}{\partial t} + \overline{\nabla \cdot \mathrm{q}\mathbf{V}} + \frac{\partial \overline{q\omega}}{\partial p} = \bar{e} - \bar{c}$$

rate of condensation of water vapor

$$\overline{\nabla \cdot \mathbf{V}} + \frac{\partial \bar{\omega}}{\partial p} = 0$$

rate of evaporation of liquid water

Eddy terms describe cloud-scale processes

Apparent heat source (Q_1)

$$Q_1 \equiv \frac{\partial \bar{s}}{\partial t} + \bar{\mathbf{V}} \cdot \Delta \bar{s} + \frac{\partial (\bar{s}\bar{\omega})}{\partial p} = Q_R + Q_c + \frac{\partial}{\partial p}(\overline{\omega' s'})$$

Apparent moisture sink (Q_2)

$$Q_2 \equiv -L_v\left(\frac{\partial \bar{q}}{\partial t} + \bar{\mathbf{V}} \cdot \Delta \bar{q} + \frac{\partial (\bar{q}\bar{\omega})}{\partial p}\right) = Q_c + L_v\frac{\partial}{\partial p}(\overline{\omega' q'})$$

*-L_v

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*$-L_v$

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Notice separation of variables beneath mean when defining Q_1 and Q_2.

$$Q_1 \equiv \frac{\partial \bar{s}}{\partial t} + \overline{\boldsymbol{V}} \cdot \Delta \bar{s} + \frac{\partial (\bar{s}\bar{\omega})}{\partial p} = Q_R + Q_c + \frac{\partial}{\partial p}\left(\overline{\omega' s'}\right)$$

$$Q_2 \equiv -L_v \left(\frac{\partial \bar{q}}{\partial t} + \overline{\mathbf{V}} \cdot \Delta \bar{q} + \frac{\partial (\bar{q}\bar{\omega})}{\partial p} \right) = Q_c + L_v \frac{\partial}{\partial p}\left(\overline{\omega' q'}\right)$$

Subtract Q_2 from Q_1:

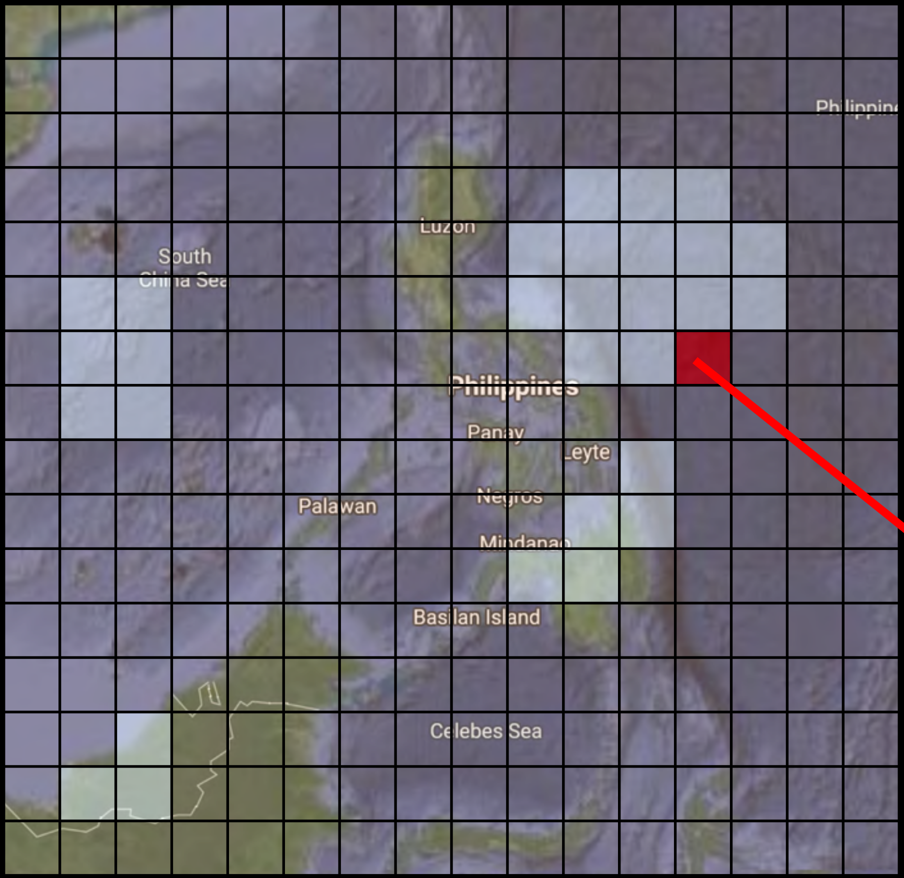
$$Q_1 - Q_2 = Q_R - \frac{\partial}{\partial p}\left(\overline{\omega' h'}\right)$$

and

$$Q_1 - Q_2 - Q_R = -\frac{\partial}{\partial p}\left(\overline{\omega' h'}\right) \longrightarrow$$

Vertical divergence of the vertical eddy transport of MSE (i.e. mostly in-cloud vertical transport of MSE)

If there is no convection in the area of interest, then $Q_1 - Q_2 = Q_R$



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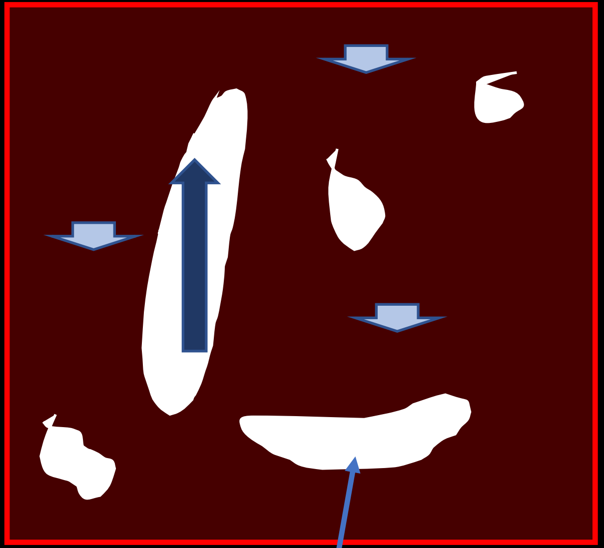

Slices through clouds

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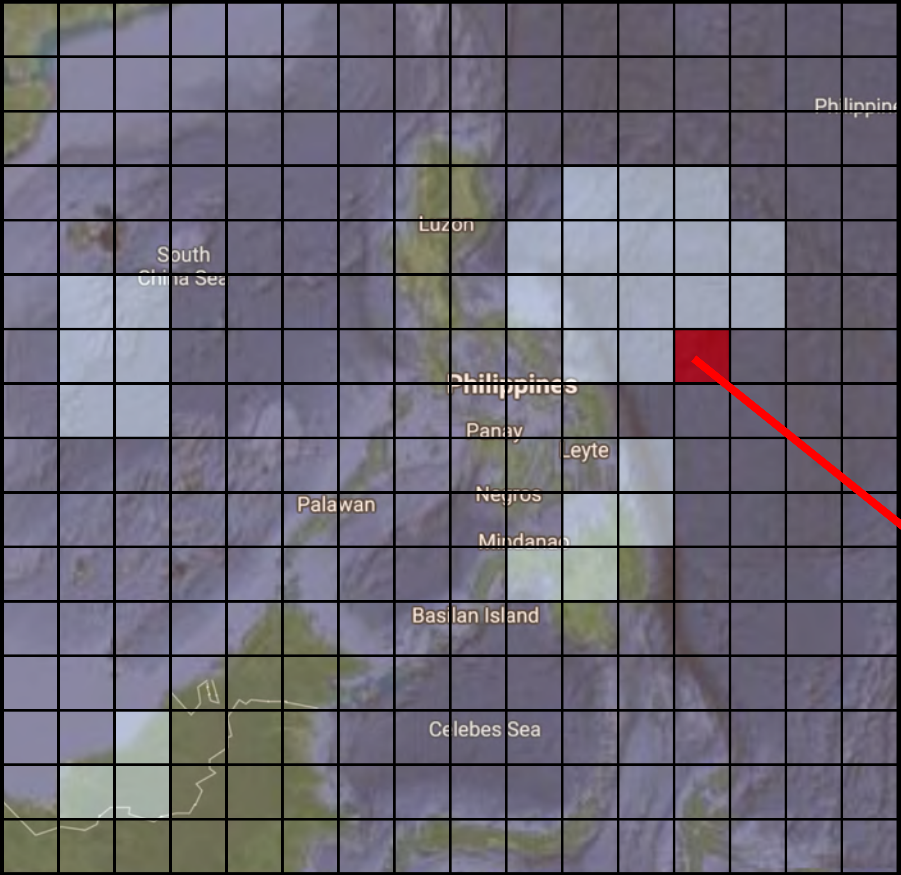

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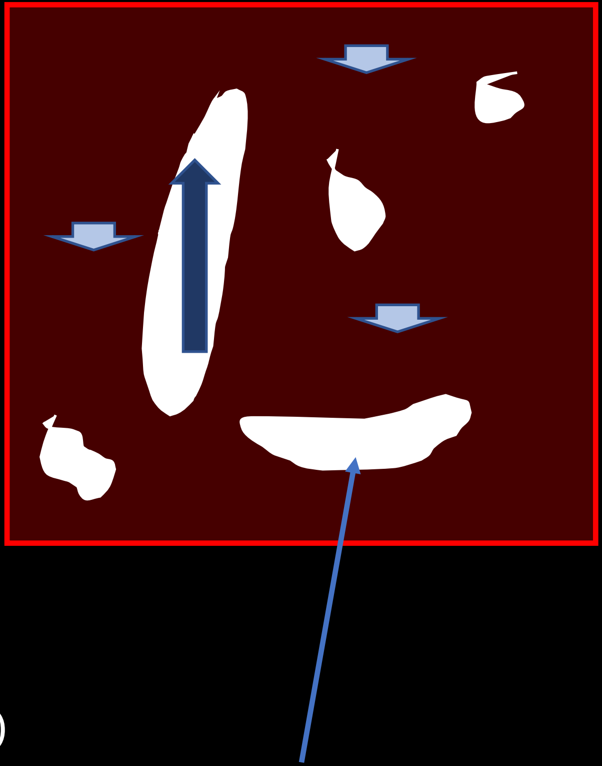


Slices through clouds

$$Q_2 \equiv \frac{\partial \overline{q}}{\partial t} + \overline{\boldsymbol{V}} \cdot \Delta \overline{q} + \frac{\partial(\overline{q}\,\overline{\omega})}{\partial p} = Q_c + L_v \frac{\partial}{\partial p}(\overline{\omega' q'})$$

Vertically-integrated heat and moisture budgets

$$Q_1 - Q_2 - \bar{Q}_R = -\frac{\partial}{\partial p}\left(\overline{h'\omega'}\right)$$

Integrate $Q_1 - Q_2 - Q_R = -\frac{\partial}{\partial p}(\overline{\omega' h'})$ **down from the level of cloud tops,** p_{top}

⇒

$$\overline{h'\omega'} = \int_{p_{top}}^{p} (Q_1 - Q_2 - Q_R)dp = -gF$$

F **is the vertical flux of total heat.**

At the surface, F_s = SHF + LHF.

Integrate $Q_1 = Q_R + L_v(\bar{e} - \bar{c}) + \frac{\partial}{\partial p}(\overline{\omega' s'})$ from p_{top} to p_{sfc} (the pressure at the sea surface)

⇒

$$\frac{1}{g}\int_{p_{top}}^{p_{sfc}} (Q_1 - Q_R)dp = \frac{L_v}{g}\int_{p_{top}}^{p_{sfc}} (\bar{c} - \bar{e})dp - \frac{1}{g}(\overline{s'\omega'})|_{p=p_{sfc}}$$

$$\approx L_v P + \rho_0 c_p(\overline{T'\omega'})|_{p=p_{sfc}}$$

$$= L_v P + SHF$$

Column-integrated radiative heating is the column-integrated heat source minus latent heating due to condensation minus the source of sensible heat from the surface.

P = Precipitation

SHF = Surface sensible heat flux

Integrate $Q_2 = L_v(\bar{e} - \bar{c}) - L_v \frac{\partial}{\partial p}(\overline{\omega' q'})$ from p_{top} to p_{sfc}

$$\frac{1}{g}\int_{p_{top}}^{p_{sfc}}(Q_2)dp = \frac{L_v}{g}\int_{p_{top}}^{p_{sfc}}(\bar{c} - \bar{e})dp - \frac{1}{g}(\overline{q'\omega'})|_{p=p_{sfc}}$$

$$\approx L_v P - \rho_0 L_v(\overline{q'\omega'})|_{p=p_{sfc}}$$

$$= L_v(P - E)$$

The rate of evaporation (E) from the surface, related to latent heat flux

The four equations:

$$\langle \overline{h'\omega'} \rangle = \int_{p_{top}}^{p} (Q_1 - Q_2 - Q_R)dp = -gF_s$$

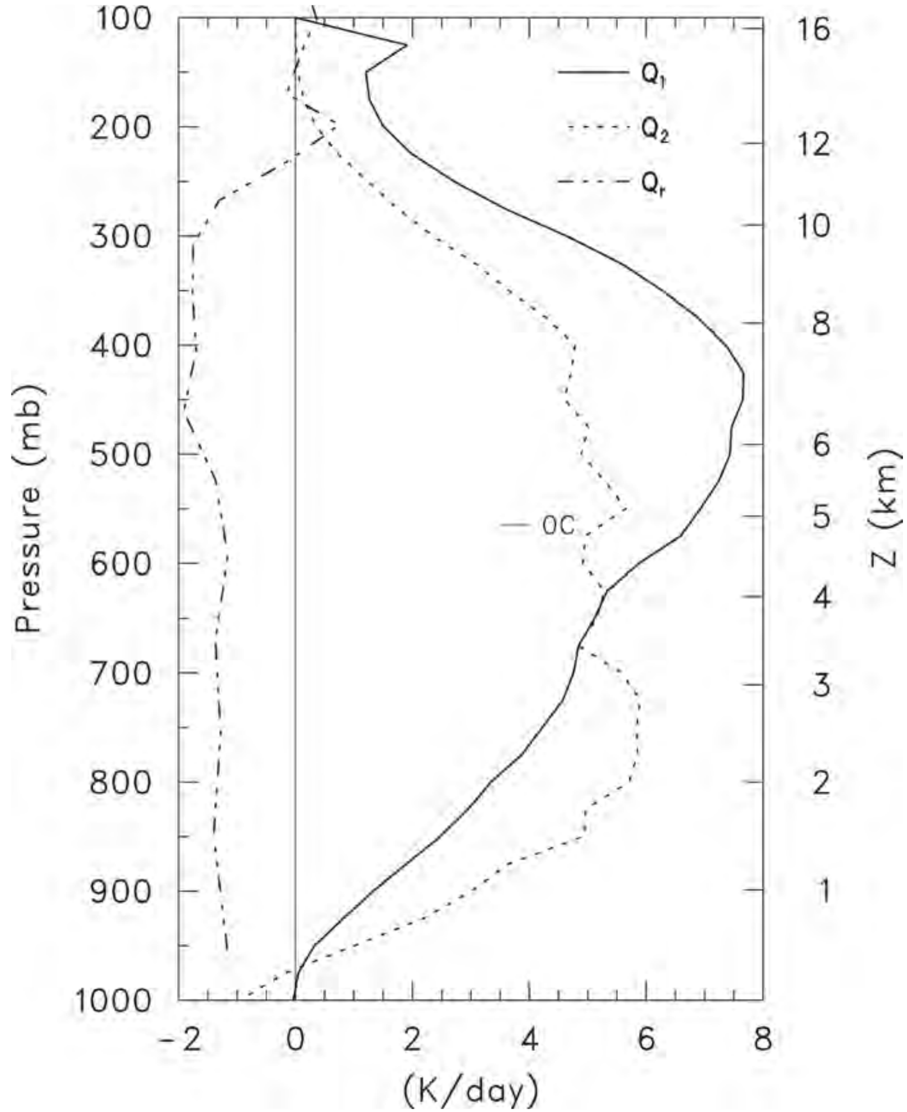
$$\frac{1}{g}\int_{p_{top}}^{p_{sfc}} (Q_1 - Q_R)dp = L_v P + SHF$$

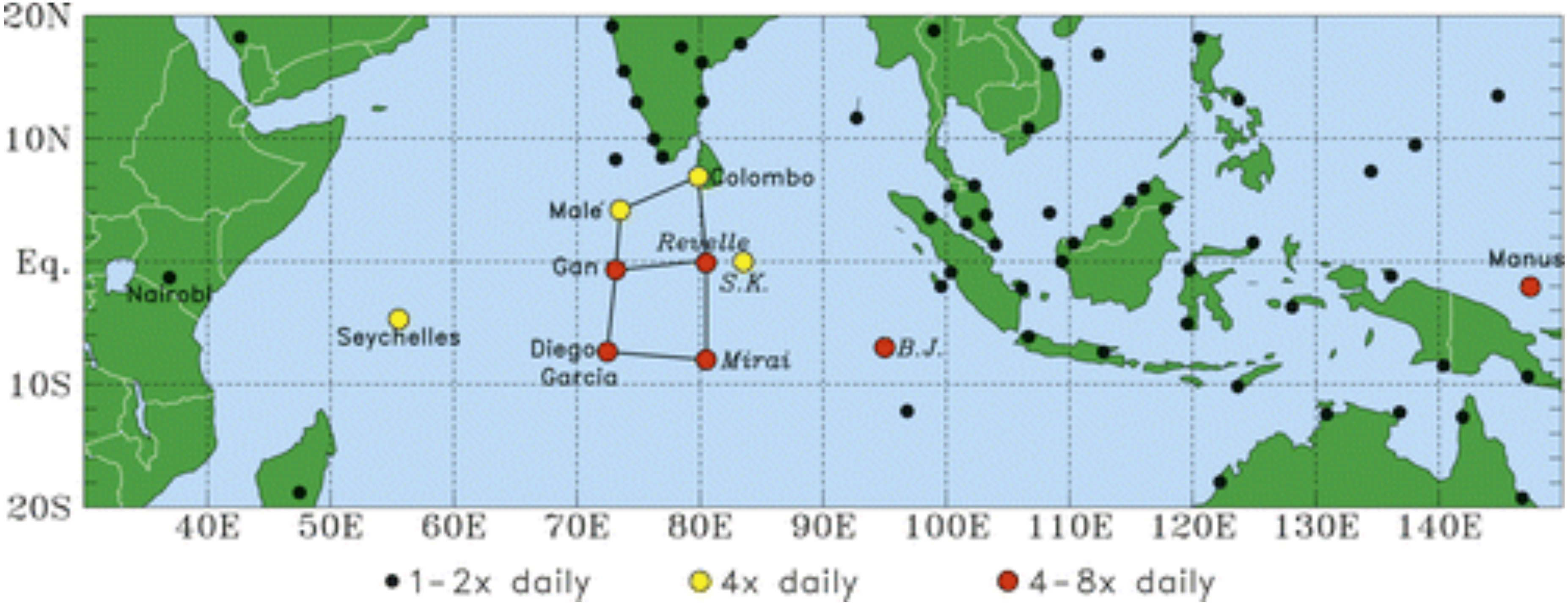
$$\frac{1}{g}\int_{p_{top}}^{p_{sfc}} (Q_2)dp = L_v(P - E)$$

$$F_s = SHF + L_v E$$

show how large-scale heating and precipitation are **constrained by surface fluxes.**

L'Ecuyer and Stephens (2003)





Johnson et al. (2013)

Q_1 and Q_2 together provide insight for cloud processes occurring on large-scale.

Powell and Houze (2015)