Welcome:

- Laboratory of Cryospheric Sciences:
- cryos.epfl.ch

Cryospheric Sciences

SLF

WSL

Welcome:

Michi Lehning

Anja Mödel

Dylan Reynolds

Léonard Lebrun

Arianna Bartolo

Lectures and Exercises (First 4 Weeks) on

Intro + Thermodynamics
Atmospheric Dynamics
Boundary Layer Dynamics
Weather

2 graded exercises

Hendrik Huwald   Brandon van Schaik
Alpole, hendrik.huwald@epfl.ch

PCA Lab course (travaux pratiques)

All relevant information can be found on the course Moodle page, first Section:

https://moodle.epfl.ch/course/view.php?id=13910

Participation is mandatory, deliverable: Lab report (counts 15% of final grade).

Looking forward to seeing you in the lab and outdoors!

Cryospheric Sciences

Atmospheric Physics and Chemistry

Welcome:

Julia Schmale

Berkay Dönmez

Yolanda Temel

Extreme Environments Research Laboratory
Associate Campus **Sion**

eerl.epfl.ch

Radiation (lecture 5,6)

Atmospheric Composition (lecture 7)

Stratospheric Chemistry (lecture 8)

Tropospheric Chemistry (lecture 9)

Introduction to Aerosols (lecture 10,11)

1 graded assignments

Please make sure to check moodle for the exact timing of lectures and exercises per week. There are holidays in between.

EPFL

Athanasios Nenes

Director of Laboratory of atmospheric processes and their impacts (LAPI)

http://lapi.epfl.ch

Topics covered in class:
Aerosol-Cloud interactions and Cloud Processes

Learning Principles

- Advanced in the Bachelor Curriculum
- Multi – Source Learning: Slides, Book, Blackboard, Problems, Additional Material – Organized on Moodle Site
- Listen to Explanations, Ask Questions, Go through Material
- **Apply Physical Principles to the Atmosphere**
  → Focus on Understanding not Calculating
- Practical Experience in TP (Hendrik Huwald)
- Observe the Atmosphere and your Environment

Recipe for Success

- Come to class and lab hours
- **Ask Questions**
- Recap the material early in the semester – revisit ZOOM recordings
- Watch video or read book/material BEFORE THE CLASS
- Watch recordings from previous years (if made available) before the class

<u>Aim:</u>

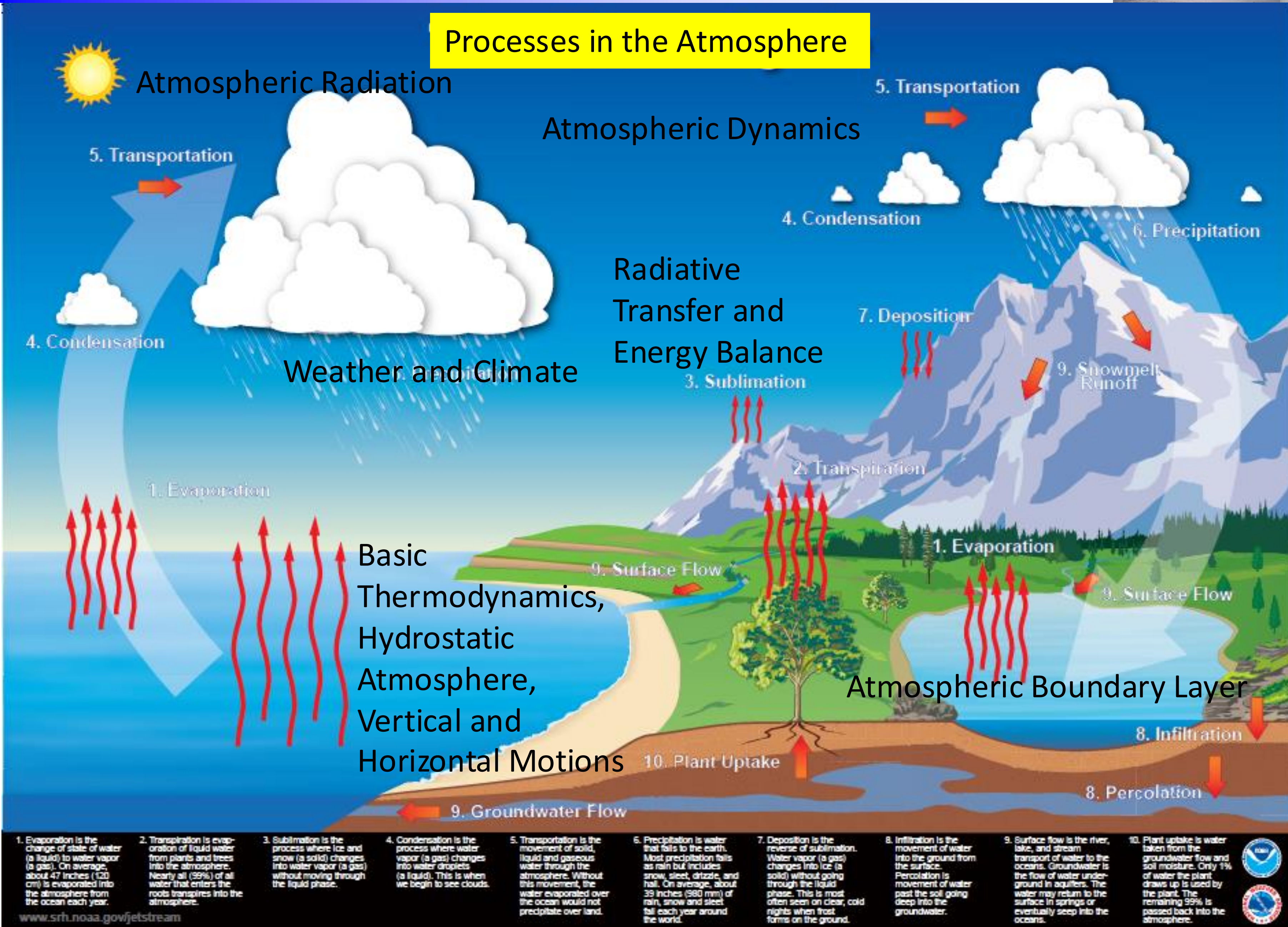
1) For you to understand the working principles of the atmosphere
2) To have some quantitative skills (calculate a geostrophic wind) and a lot of conceptual knowledge

<u>Exam:</u>

- Closed Book + Cheat Sheet; Questions on how processes work together

Overview Atmosphere, Thermodynamics, Motion

Overview on Course – Physics Part



Outline of First Block:

- What is Atmospheric Physics – An Introduction
- Thermodynamics
- Important Quantities and Definitions
- Vertical Motions and **Stability**
- Introduction to Horizontal Motions

Atmospheric Science

- From Magic / Art to Science
- A Basis for Warning from Natural Hazards
- Huge Economic Value – from Energy Trading to Agriculture
- The basis to understand our climate system
- Based on International Collaboration –
  Model Development and Data Gathering (Assimilation)

A View on Our Thin Atmosphere

NASA Gemini-4 photo:
Visible light scattered from atmospheric aerosols (red - white) and air molecules (blue)

How thick is what you see here?

A. 1 km
B. 10 km
C. 100 km
D. 1000 km

One Picture Summary: Cumulonimbus with Anvil Cloud

[Photograph courtesy of Rose Toomer and Bureau of Meteorology, Australia.]

How thick is what you see here?

P

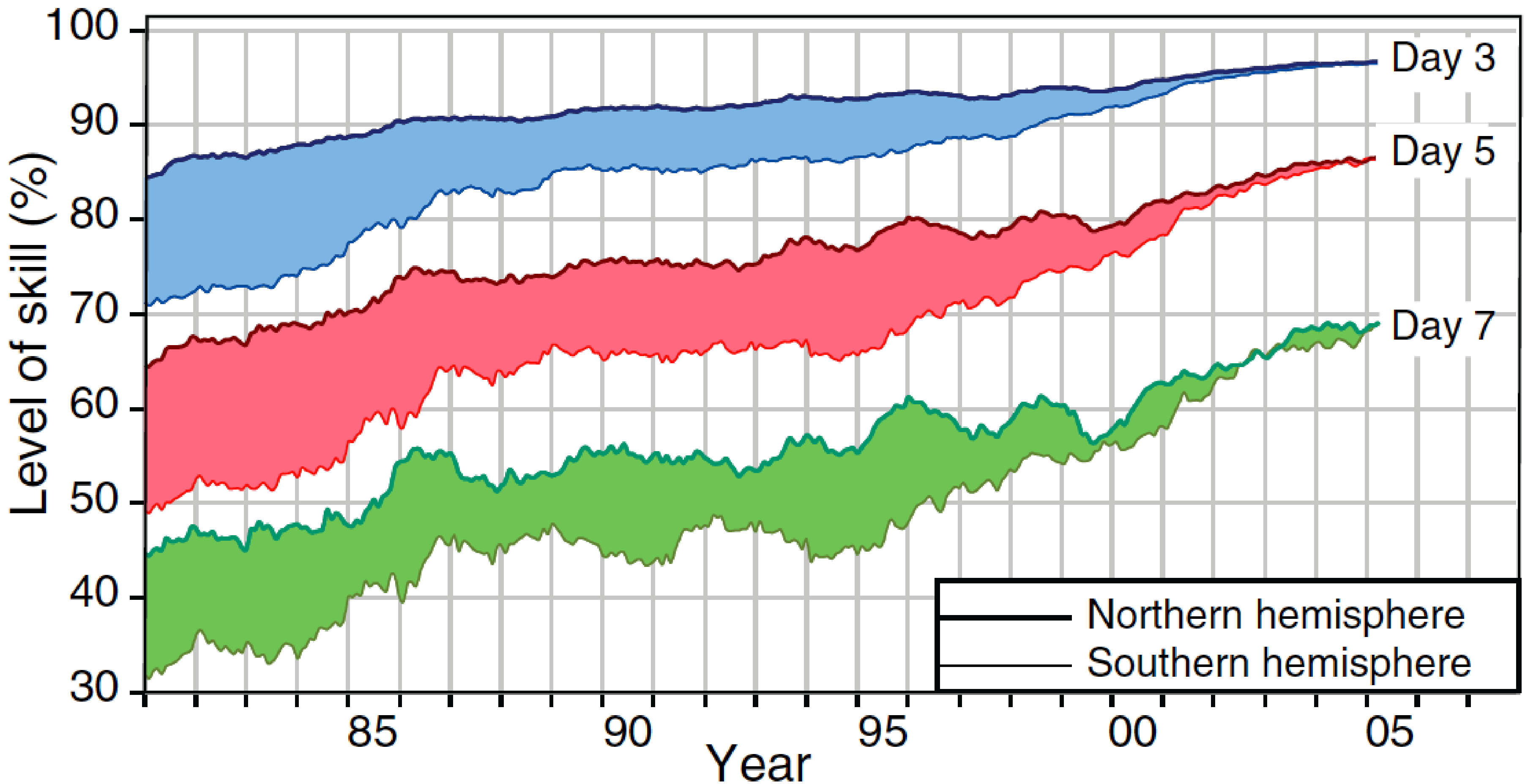
A. 1 km
B. 10 km
C. 100 km
D. 1000 km

AP between Meteorology ....

Hurricane Floyd
1999 on Florida

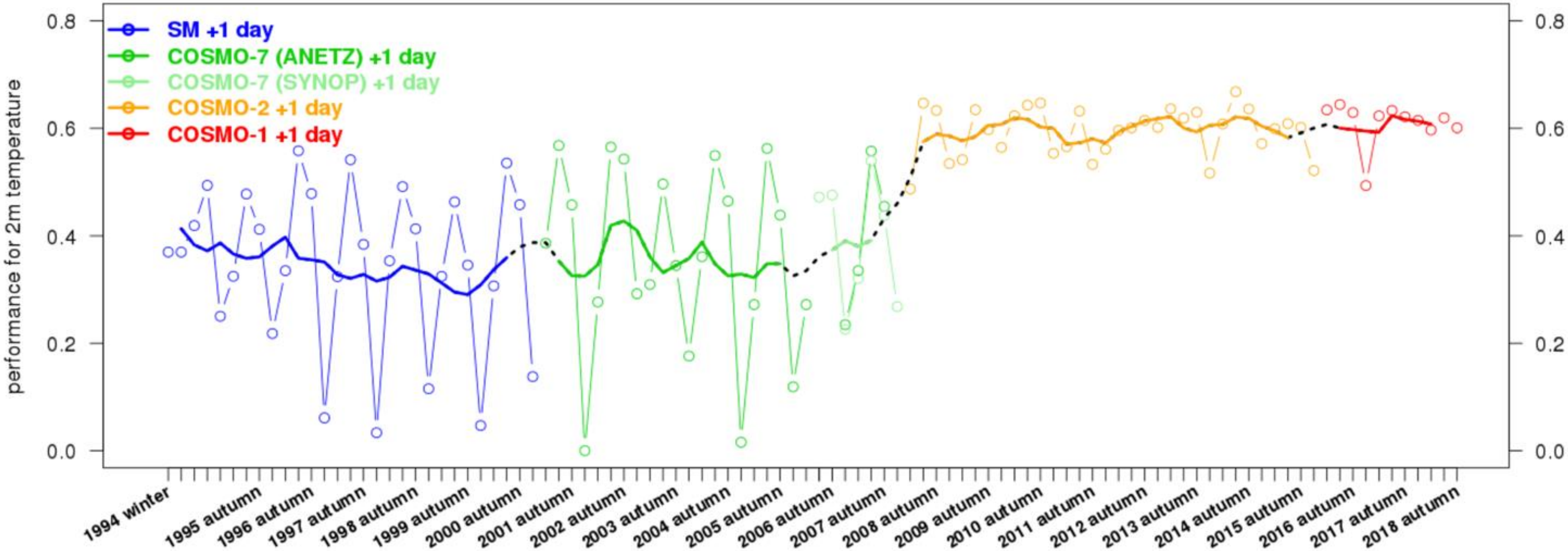
.......Weather Forecasting

Skill of hemispheric flow forecasts at 5 km. Convergence between hemispheres is because of increasing success in using satellite data and ocean observations for assimilation.



Swiss Weather (Temperature) Forecasting

MeteoSwiss has to report its forecast skill to the parliament. Below you find a performance measure for one-day (very short term) temperature forecast and how it relates to the introduction of the numerical weather forecast models.



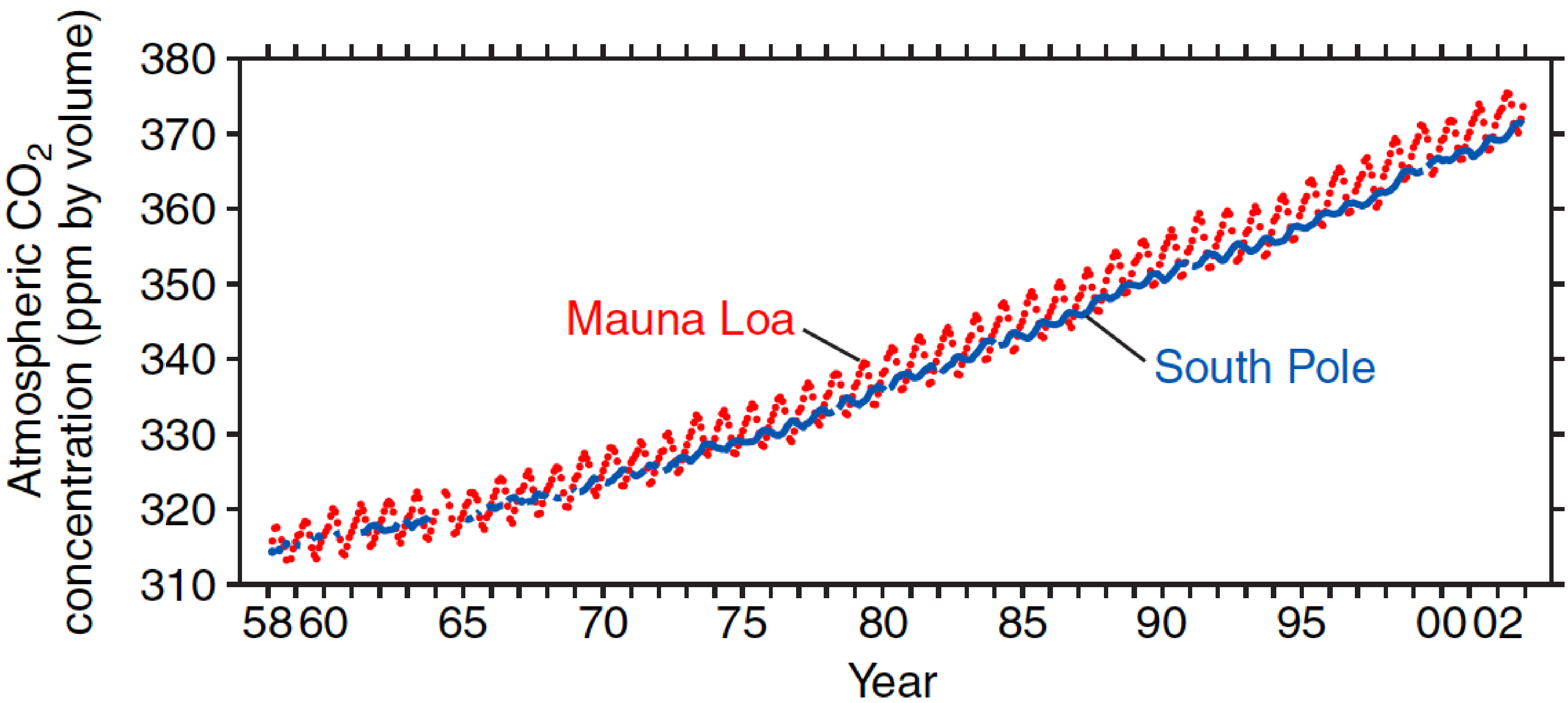
What is your personal 3-day skill score you give to MeteoSwiss?

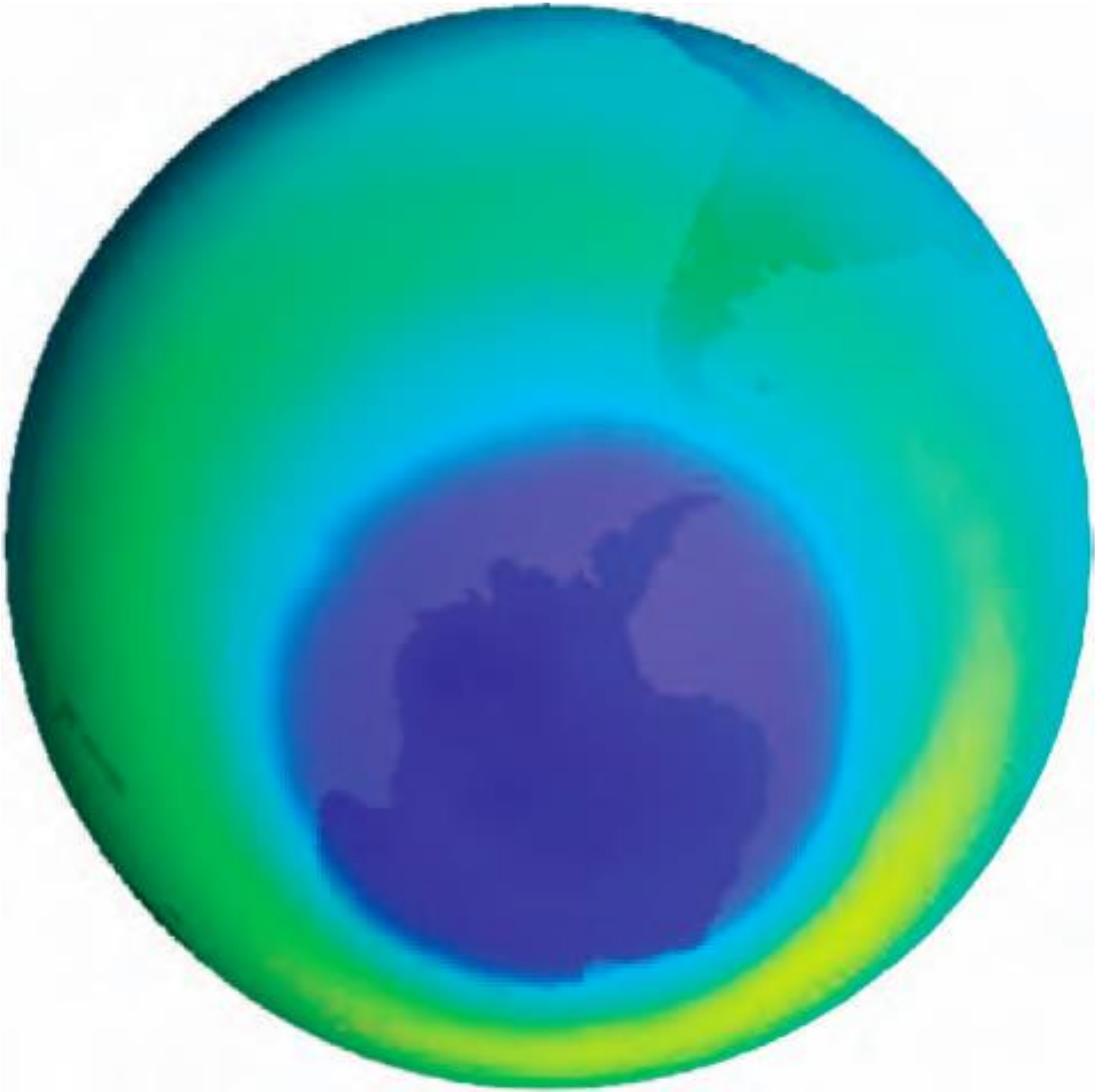
A. 100%
B. 90%
C. 70%
D. 50%

Why is precipitation harder to forecast than temperature?

A. Climate is changing
B. Precipitation formation is very complex
C. Precipitation events can be very local
D. Precipitation is a result of solving the Navier Stokes equations

Monthly CO2 Concentrations show a rising trend; seasonal cycles are caused by photosynthesis activity changes modulated by transport.





NASA TOMS picture of the Antarctic ozone hole in 2000

Atmospheric Basic Facts and Thermodynamics

- Atmospheric Chemical Composition
- Vertical Structure
- Mass of Atmosphere and Hydrostatic Equation
- Moisture in the Atmosphere
- Air Parcels, Stability and Condensation Level
- Introduction to Motions in the Atmosphere

Mixture of Gases in the Atmosphere

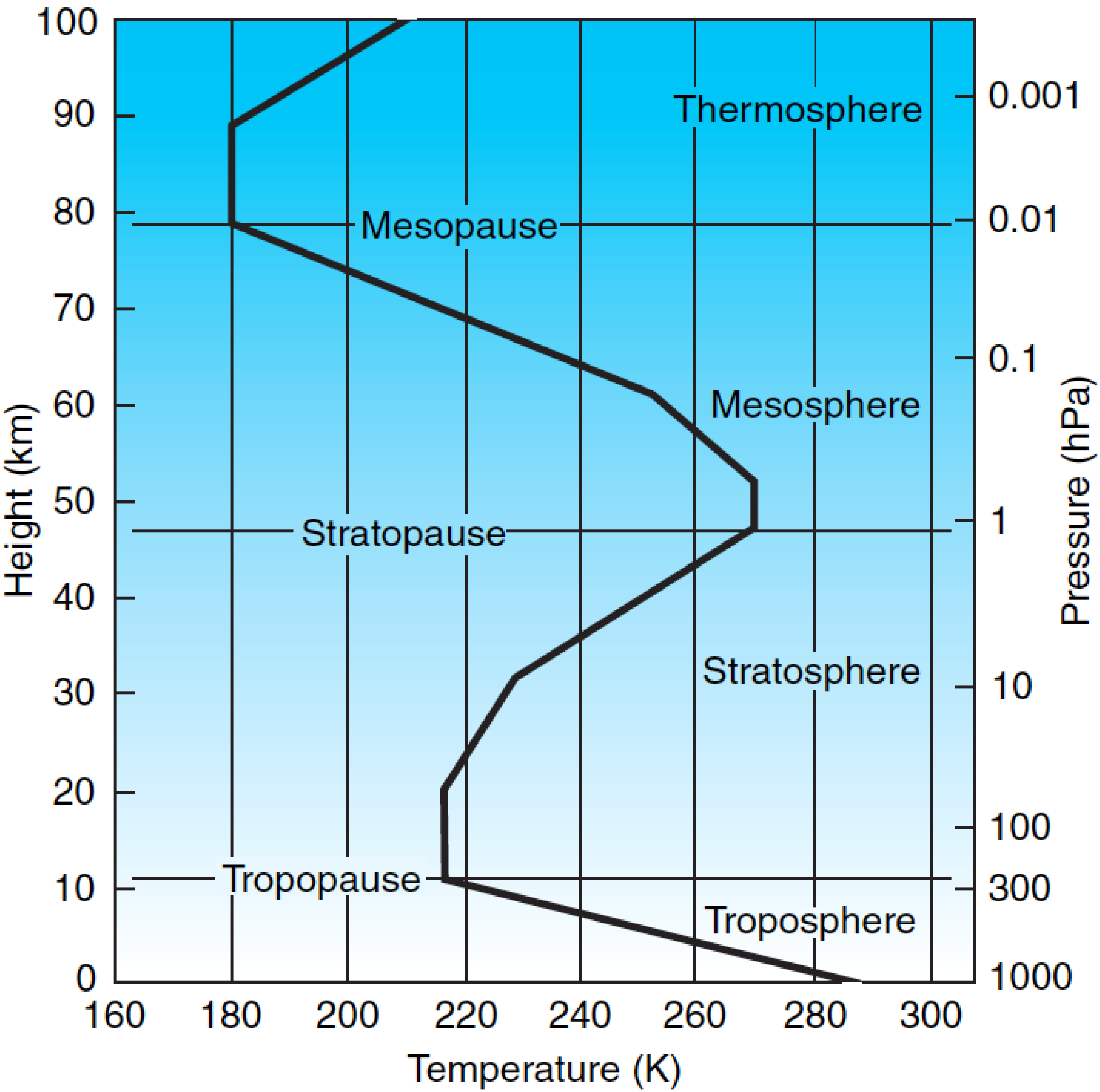
Table 1.1 Fractional concentrations by volume of the major gaseous constituents of the Earth's atmosphere up to an altitude of 105 km, with respect to dry air

Constituent[a]	Molecular weight	Fractional concentration by volume
Nitrogen (N_2)	28.013	78.08%
Oxygen (O_2)	32.000	20.95%
Argon (Ar)	39.95	0.93%
Water vapor (H_2O)	18.02	0–5%
Carbon dioxide (CO_2)	44.01	380 ppm
Neon (Ne)	20.18	18 ppm
Helium (He)	4.00	5 ppm
Methane (CH_4)	16.04	1.75 ppm
Krypton (Kr)	83.80	1 ppm
Hydrogen (H_2)	2.02	0.5 ppm
Nitrous oxide (N_2O)	56.03	0.3 ppm
Ozone (O_3)	48.00	0–0.1 ppm

[a] So called *greenhouse gases* are indicated by bold-faced type. For more detailed information on minor constituents, see Table 5.1.

- Well mixed up to 100 km
- Water Vapor confined to lower part
- Above 100 km, heavier gases become relatively less abundant

Vertical Structure



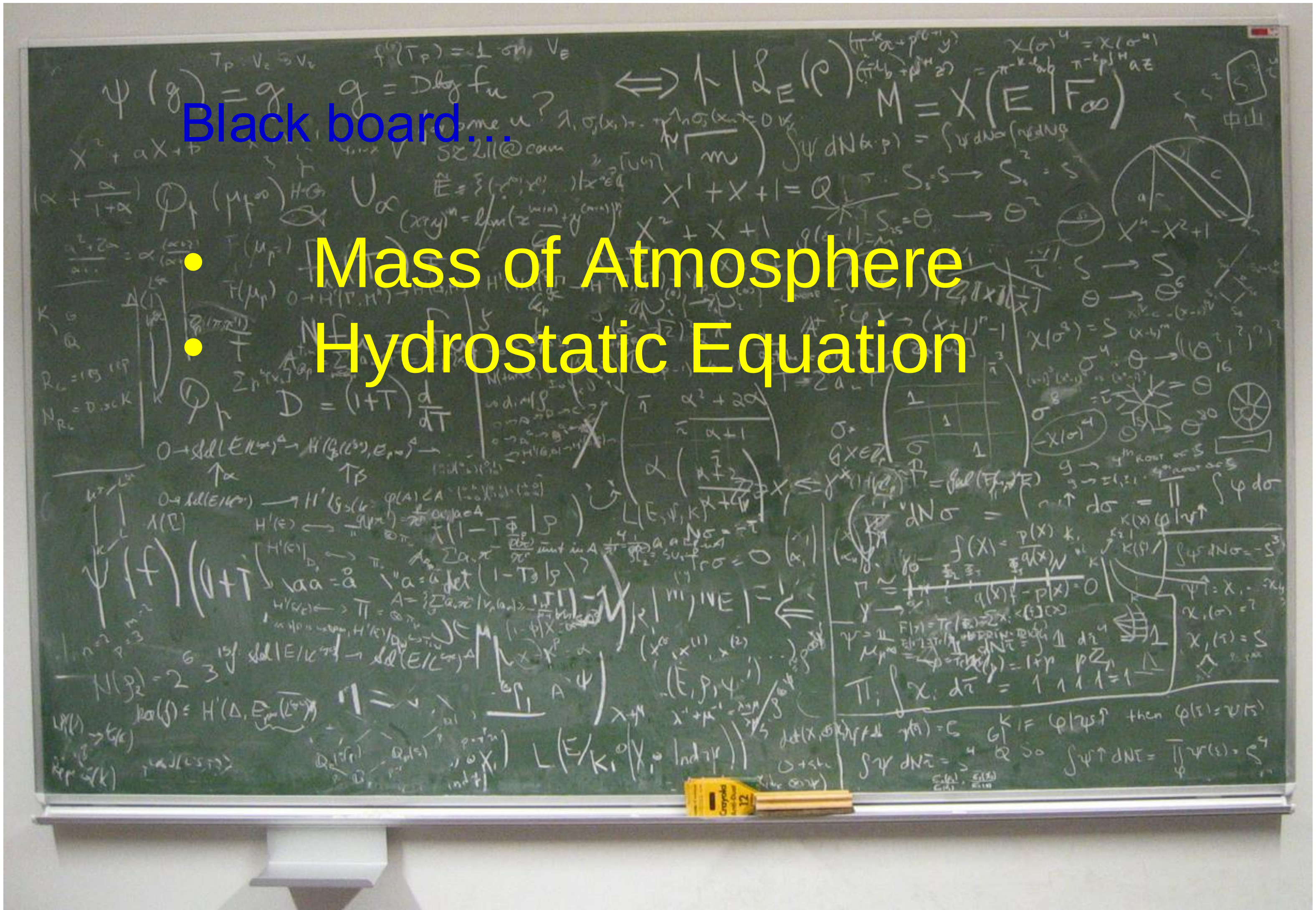
Troposphere:

$$\Gamma = \frac{-\partial T}{\partial z} \approx 6.5 K\ km^{-1}$$

Stratosphere: Absorption of UV radiation by Ozone

Thermosphere: Potodissociation and Photoionization through UV and X-ray

Basic Properties



Black board…

- Mass of Atmosphere
- Hydrostatic Equation

Mass of Atmosphere

Newton's second law:

$$F = m'\,a;$$
$[F] = \mathrm{N}$
$[m'] = \mathrm{kg}$
$[a] = \mathrm{m\ s^{-2}}$

Define (as in the book) :

$$m = \frac{m\prime}{A} \Longrightarrow m = \frac{p}{g}\ ;$$

Pressure on a unit surface:

$$\frac{F}{A} = p = \rho\ g\ h;$$
$[A] = \mathrm{m^2}$
$[p] = \mathrm{Pa}$
$[\rho] = \mathrm{kg\ m^{-3}}$
$[h] = \mathrm{m}$
$[g] \approx 9.81\ \mathrm{m\ s^{-2}}$

With a pressure of approx. 1000 hPa, we have about 10^4 kg m^{-2} mass of atmosphere above each square meter on the surface of the earth.

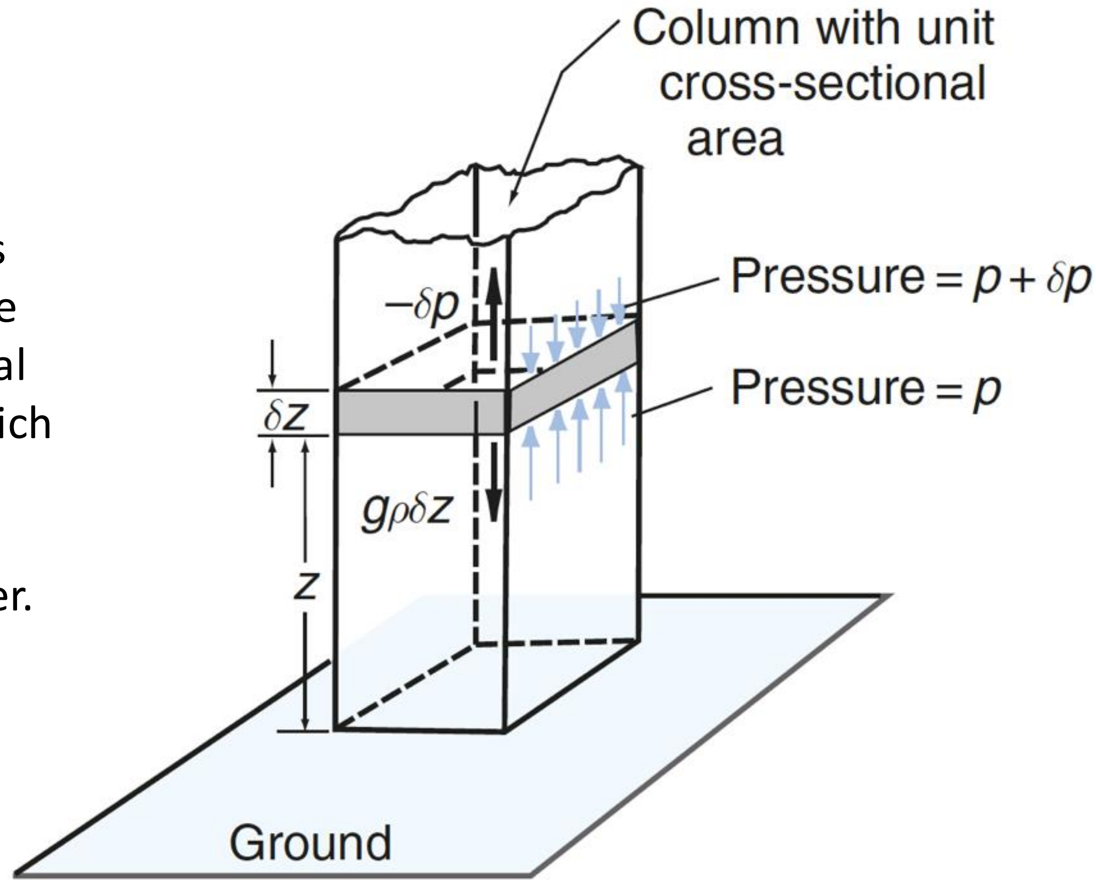
Hydrostatic Equation

Balance of Forces:

$$-\delta p = \rho\, g\delta z;$$
$$\delta z \rightarrow 0;$$

$$\frac{\partial p}{\partial z} = -\rho\, g;$$

The hydrostatic equation describes the force balance in an atmosphere at rest, in particular without vertical motion. It is an approximation, which can be used at large scales but not e.g. in clouds, for flow over mountains, or in the boundary layer.



Geopotential, Height and Pressure

Define:

$$d\Phi \equiv \; gdz = -\frac{1}{\rho}dp;$$

$$\Phi(z) = \int_0^z gdz\,;$$

$[\Phi] = \mathrm{J\ kg^{-1}}$

Define geopotential height:

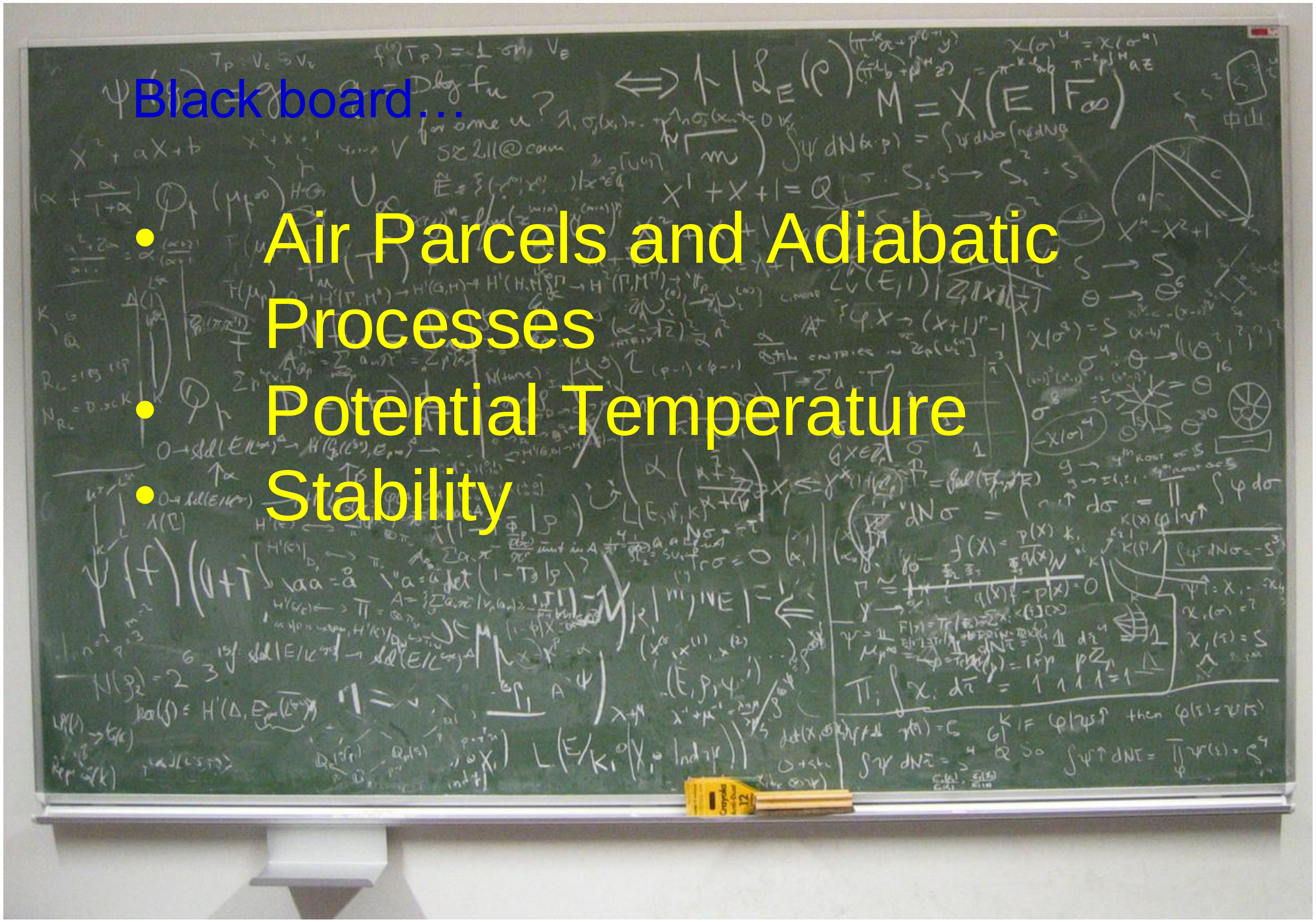
$$Z \equiv \frac{\Phi(z)}{g_0} = \frac{1}{g_0}\int_0^z gdz\,;$$

$[Z] = m$

The geopotential Φ (z) is the work required to lift 1 kg against gravity to a height z. It is often used to replace pressure gradients as a driving force to atmospheric motion. This has been useful because the spatial variation of *g* is then already considered.

The geopotential height is a useful vertical coordinate (normalized with respect to g) in that in can be converted to geopotential by using a globally averaged g_0=9.81 m s^{-2}.

Stability and Condensation Level



Black board…

- Air Parcels and Adiabatic Processes
- Potential Temperature
- Stability

Energy Conservation:

$$\mathrm{d}q = c_p\, dT - \frac{1}{\rho} dp;$$

work

$[q] = \mathrm{J\ kg^{-1}}$
$[T] = \mathrm{K}$
$c_p = 1004 \mathrm{J\ kg^{-1} K^{-1}}$

change in internal energy

heat exchange

For the atmosphere, the exchange of work and heat most often is characterized by a change in temperature. Therefore, temperature is often used to describe the energy state of an air parcel.

An adiabatic process is one in which the heat exchange with the environment is zero:
$\mathrm{d}q = 0.$
For simplification we often regard movement of air parcels in the atmosphere and sometimes approximate these movements to happen adiabatically. Examples are a polar airmass outbreak that moves over Europe or an airmass that crosses the Alps.

Dry Adiabatic Lapse Rate

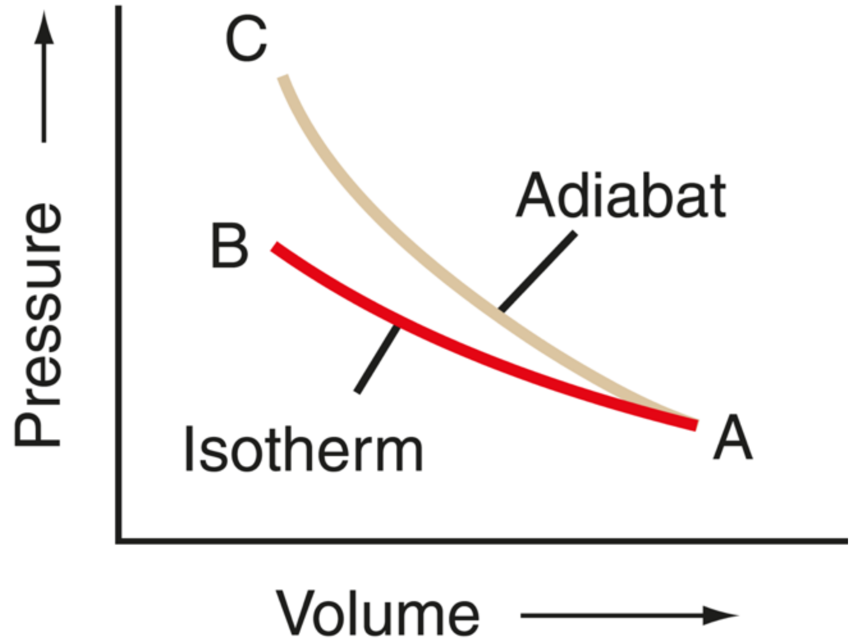
Adiabatic motion of an air parcel in a hydrostatic atmosphere:

$$0 = c_p\, dT - \frac{1}{\rho} dp;$$

Take the derivative with respect to height, z:

$$c_p \frac{dT}{dz} = \frac{1}{\rho}\frac{dp}{dz} = -\frac{1}{\rho}\rho g;$$

$$\left(\frac{dT}{dz}\right)_{ad} = -\frac{g}{c_p} = -\Gamma_d \approx -9.8\ K km^{-1}$$



In a dry atmosphere (with no phase changes), the temperature decreases approximately 1° C per 100 m.

Potential Temperature

Potential Temperature is the temperature an air parcel would have when adiabatically brought to sea level (1000 hPa)

$$0 = c_p\, dT - \frac{1}{\rho} dp;$$

Replace density by ideal gas law:

$$p = \rho R T;$$
$$R \approx 287 \text{ J kg}^{-1}\text{ K}^{-1}$$
$$c_p \approx 1000 \text{ J kg}^{-1}\text{ K}^{-1};$$
$$c_p \frac{dT}{T} = R\frac{dp}{p};$$

Integrate between sea level and some level, z, with pressure, p:

$$\frac{c_p}{R}\int_{\theta}^{T}\frac{dT'}{T'} = \int_{p_0}^{p}\frac{dp'}{p'};$$

$$\frac{c_p}{R} ln\frac{T}{\theta} = ln\frac{p}{p_0};$$

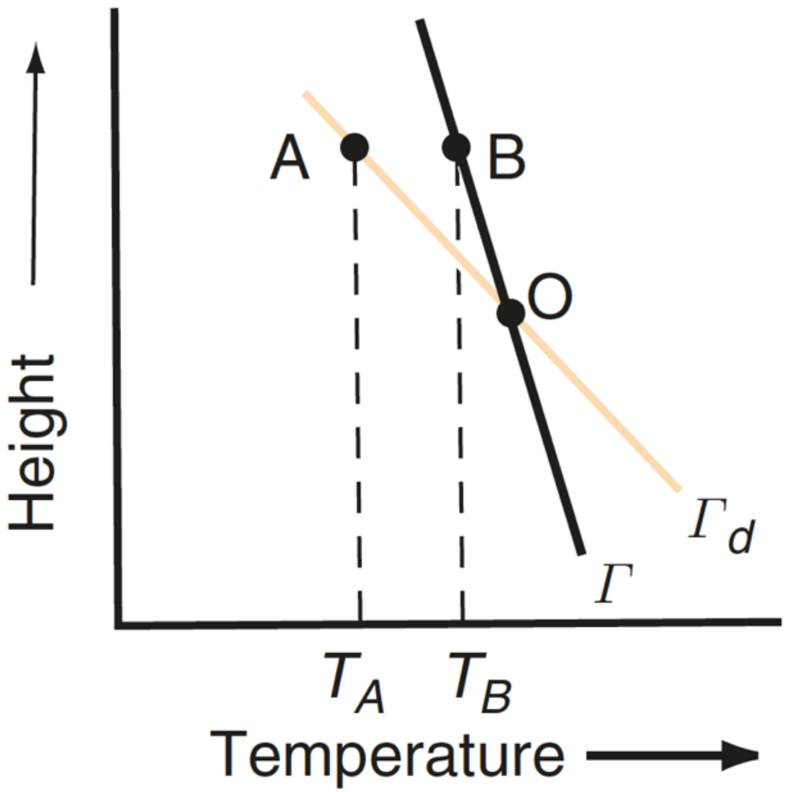
$$\left(\frac{T}{\theta}\right)^{c_p/R} = \left(\frac{p}{p_0}\right);$$

$$\theta = T\left(\frac{p_0}{p}\right)^{R/c_p};$$

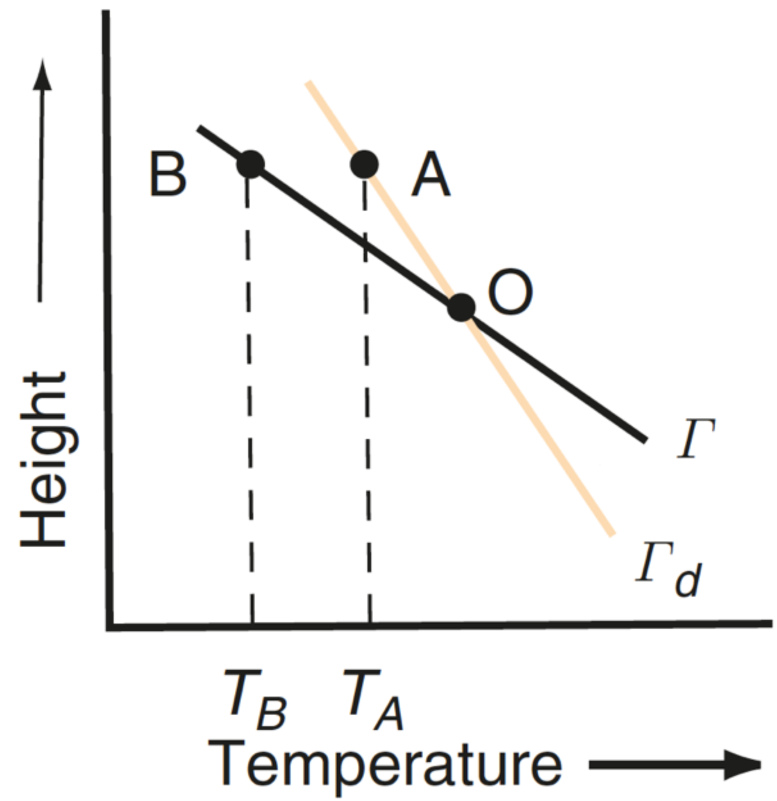
Potential temperature can be calculated from the temperature and the two pressure levels. If you move an air parcel from some height to sea level and there are no diabatic processes then its temperature at sea level will be equal to its potential temperature. On the way down (or up) it will follow the dry adiabatic lapse rate

Static Stability – Is the stratification stable or unstable?

Stability analysis is based on whether “warmer – lighther – less dense” air is below “heavier” air or above “heavier” air. You can judge this from the temperature profile alone by comparing to the dry adiabatic lapse rate. The real (measured) lapse rate is shown by Γ in the graphs below.



(a) stable



(b) unstable

Moisture in the Atmosphere

Mixing Ratio: $w = \frac{m_v}{m_d}$;

m_v mass of vapor (kg); m_d mass of dry air (kg)

Specific Humidity: $q = \frac{m_v}{m_d + m_v} = \frac{w}{1+w} \approx w$

Relative Humidity RH: $RH = 100\frac{w}{w_s} = 100\frac{e}{e_s}$;

$e_{(s)}$ (saturation) vapor pressure (Pa)

Dewpoint Temperature: T_d (K)

The hypothetical temperature to which an air mass needs to be cooled to become saturated

$\left[\text{Saturated MR: } w_s = \frac{m_{vs}}{m_d} = 0.622\frac{e_s}{p - e_s} \approx 0.622\frac{e_s}{p}\right]$

Why is temperature decreasing with elevation in the atmosphere

A. The sun heats the Earth's surface
B. Pressure is decreasing
C. Space is cold
D. Clouds are producing shadow

P

Sinking of air parcels can happen because of radiative cooling at cloud tops

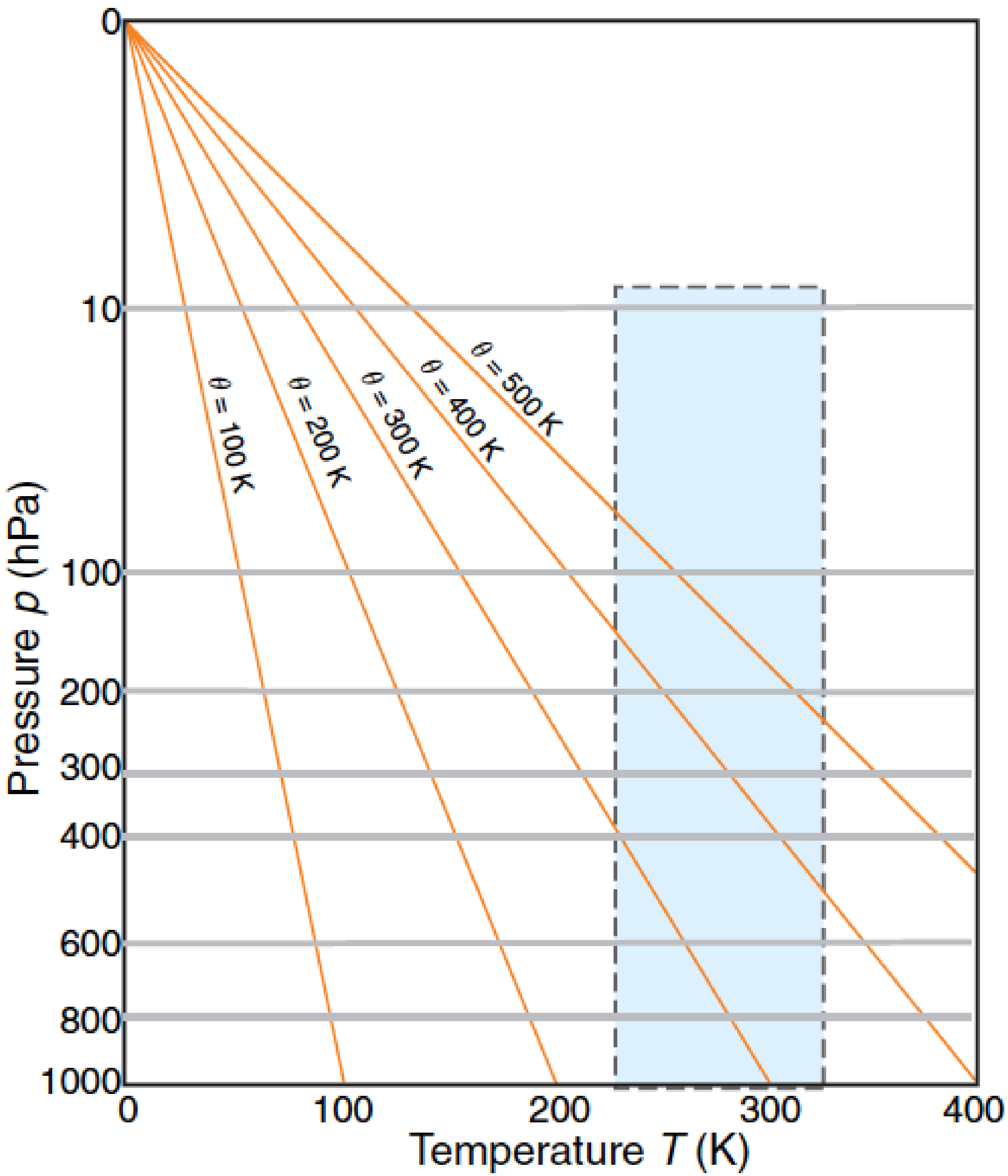
A. True
B. False

Sinking of air parcels can happen because of convective rising of air elsewhere

P

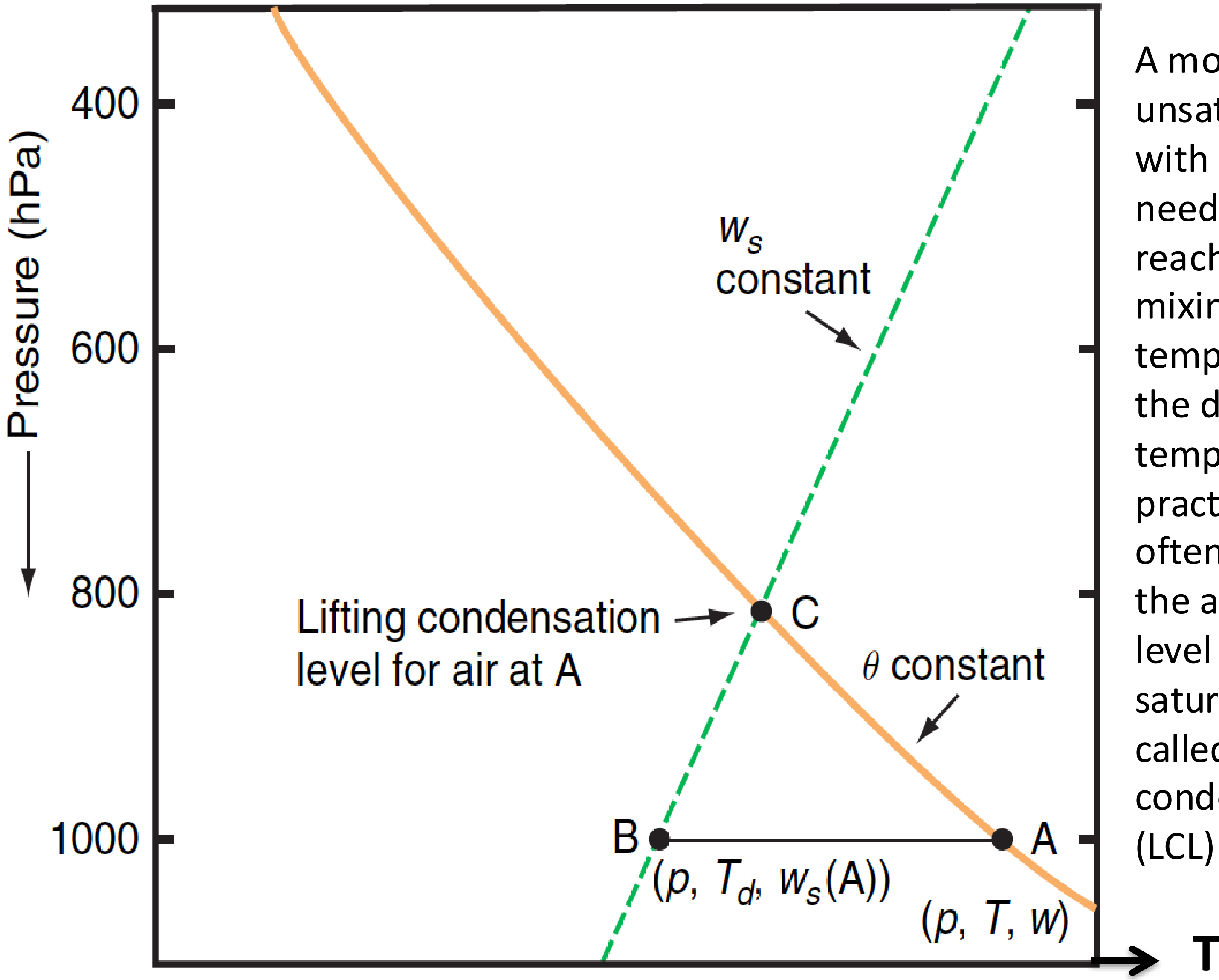
A. True
B. False

Thermodynamic Diagram: Pseudoadiabatic Chart



The chart plots $p^{0.286}$ vs. T. The dry adiabats are straight lines, isotherms are vertical (straight) lines and pressure levels horizontal. Only the blue area is typically used in atmospheric applications. Since isotherms and adiabats have only a small (acute) angle between them, a different chart will be introduced and used in class.

Lifting Condensation Level and T_d



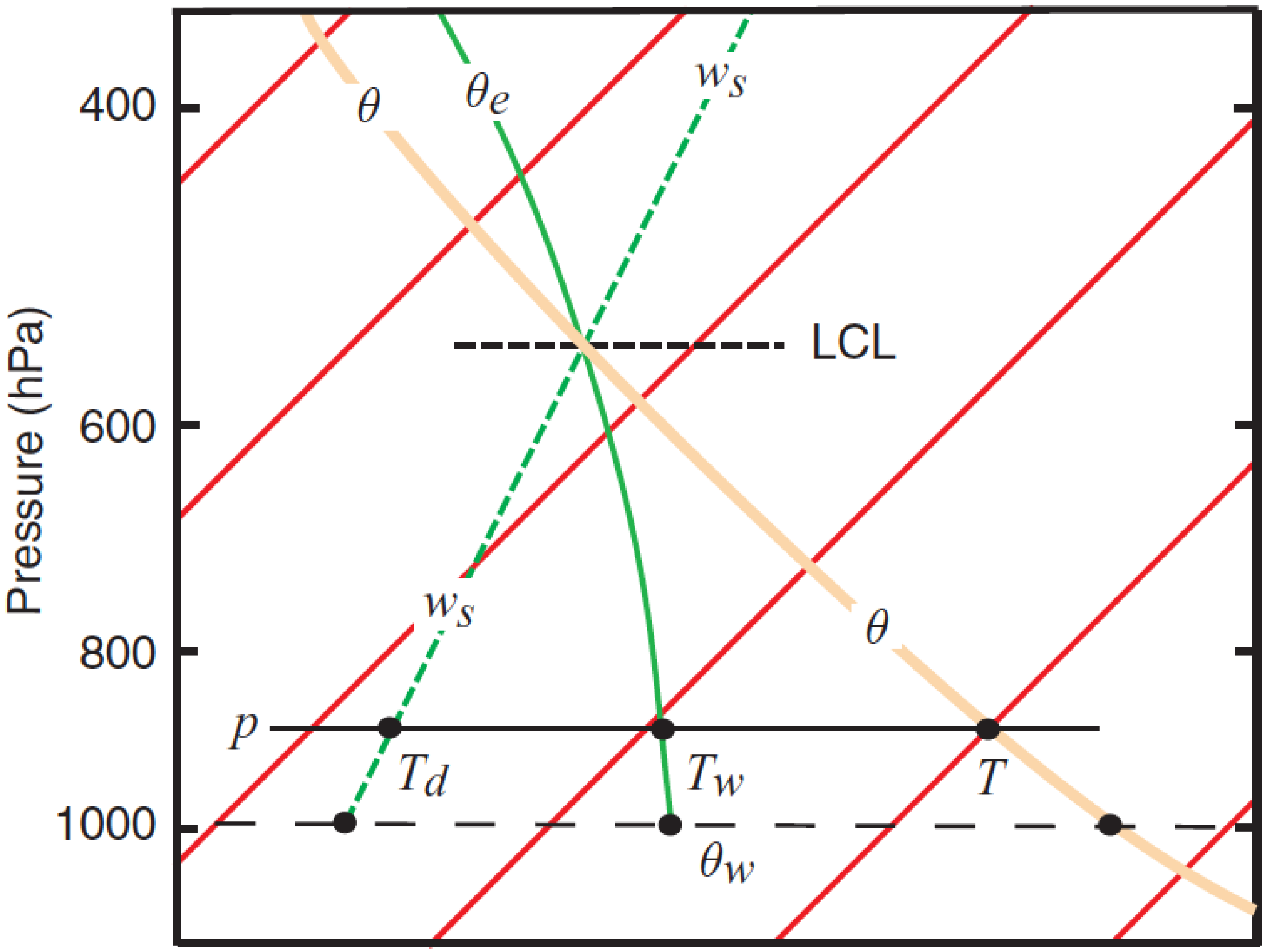
A moist but unsaturated air parcel with mixing ratio w (A) needs to be cooled to reach the saturation mixing ratio w_s. This temperature is called the dewpoint temperature T_d. In practice, the cooling is often achieved by lifting the air parcel and the level at which saturation occurs is called the lifting condensation level (LCL).

Moist Adiabatic Lapse Rate

The moist adiabatic lapse rate is the rate of cooling an air parcel experiences when it is lifted but has already reached saturation. That means that the cooling is no longer given by the dry adiabatic lapse rate. The cooling is reduced because condensation of water vapour occurs as the parcel is further lifted in a continuous way. This condensation adds heat to the parcel and therefore, less cooling occurs.

The “fate” of an air parcel can be graphically described by the thermodynamic diagrams. One example is in the next slide.

Thermodynamic Diagram: Skew *T* – *ln p* Chart



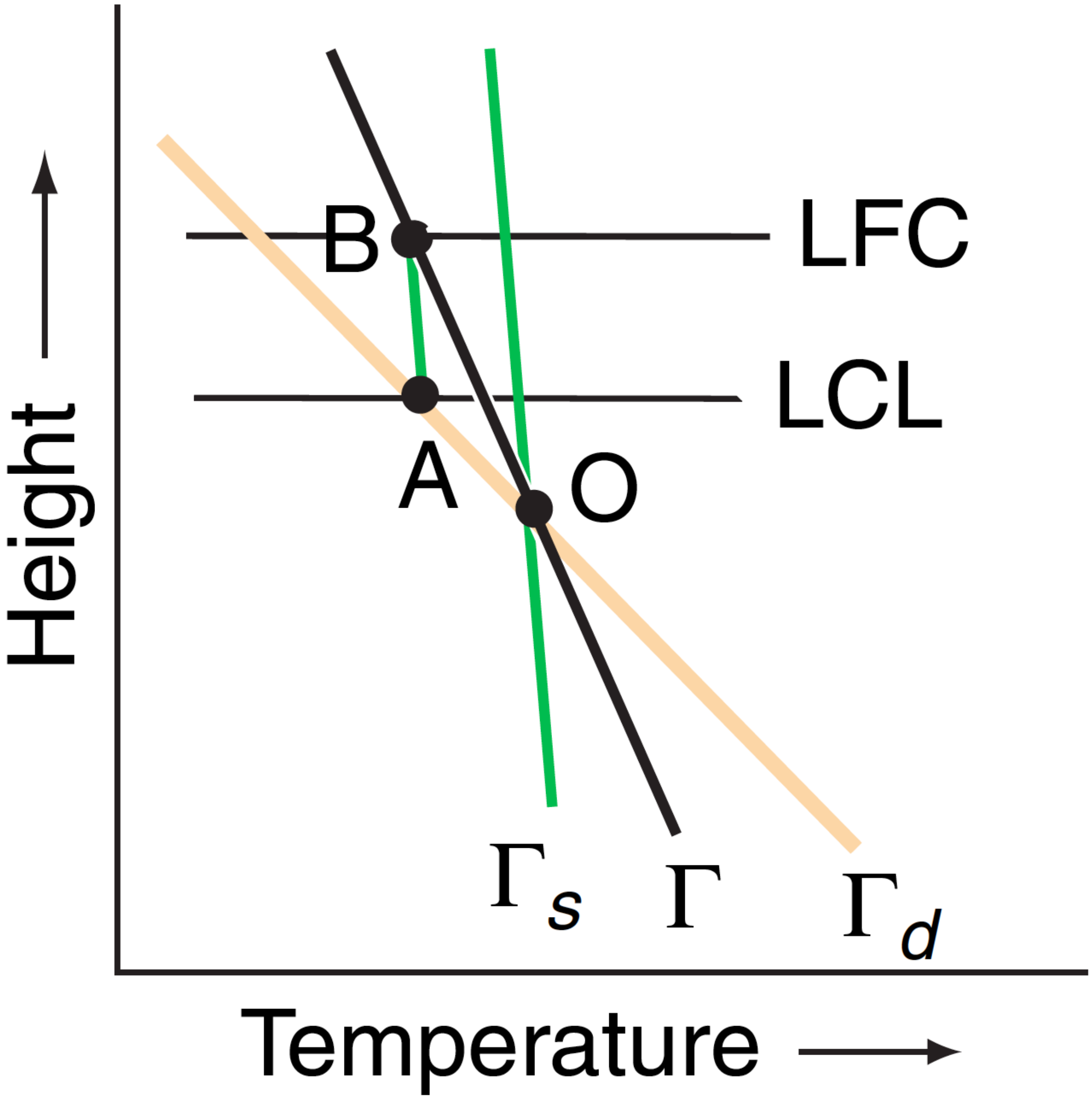
The chart has isotherms (red-orange), a dry-adiabat (θ - yellow), a moist adiabat (θ_e - green), a line of constant saturation mixing ratio (w_s) and the (horizontal) constant pressure (p - black). An air parcel with T, p, ($w = w_s$) has the lifting condensation level (LCL) as shown. Also shown as black points are dew point temperature (T_d), wet bulb temperature (T_w) and their respective potential counterparts at the 1000 hPa level.

Wet bulb temperature measures

P

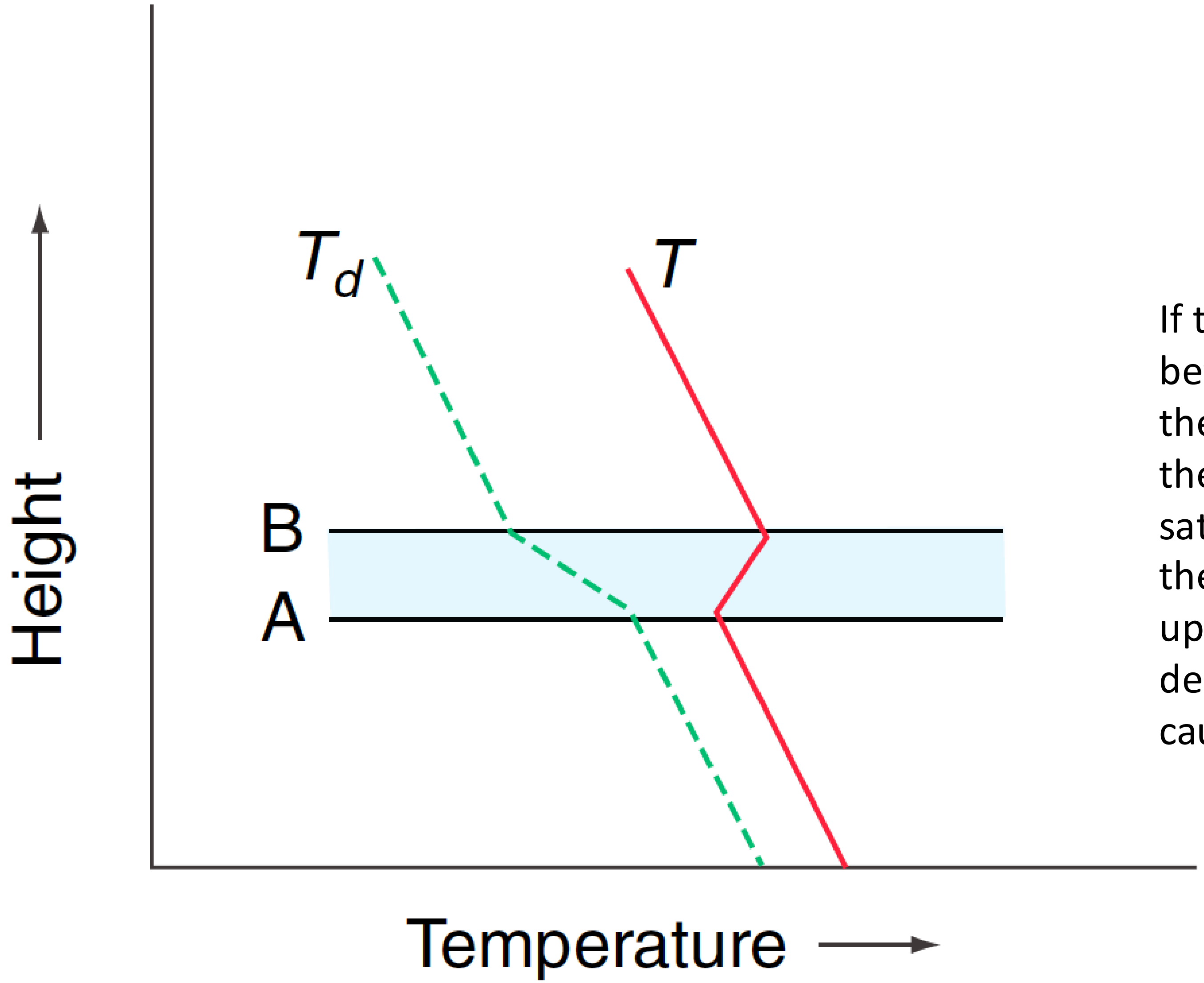
A. The temperature of vegetation close to the surface (flower bulbs)
B. The moisture of the atmosphere
C. The temperature of a moist air parcel

Conditional and Convective Instability



For a given environmental lapse rate Γ, which is between the moist adiabatic lapse rate Γ_s and the dry adiabatic lapse rate Γ_d, lifting may lead to free convection. If an air parcel is lifted from **O**, it will first cool according to Γ_d until the lifting condensation level (LCL) is reached **(A)** then it will follow Γ_s until it crosses the environmental lapse rate **(B)**. From this point on it will find itself warmer and therefore less dense than the environment and lifting will continue without external forcing. This second level is therefore called the level of free convection (LFC).

Convective or Potential Instability

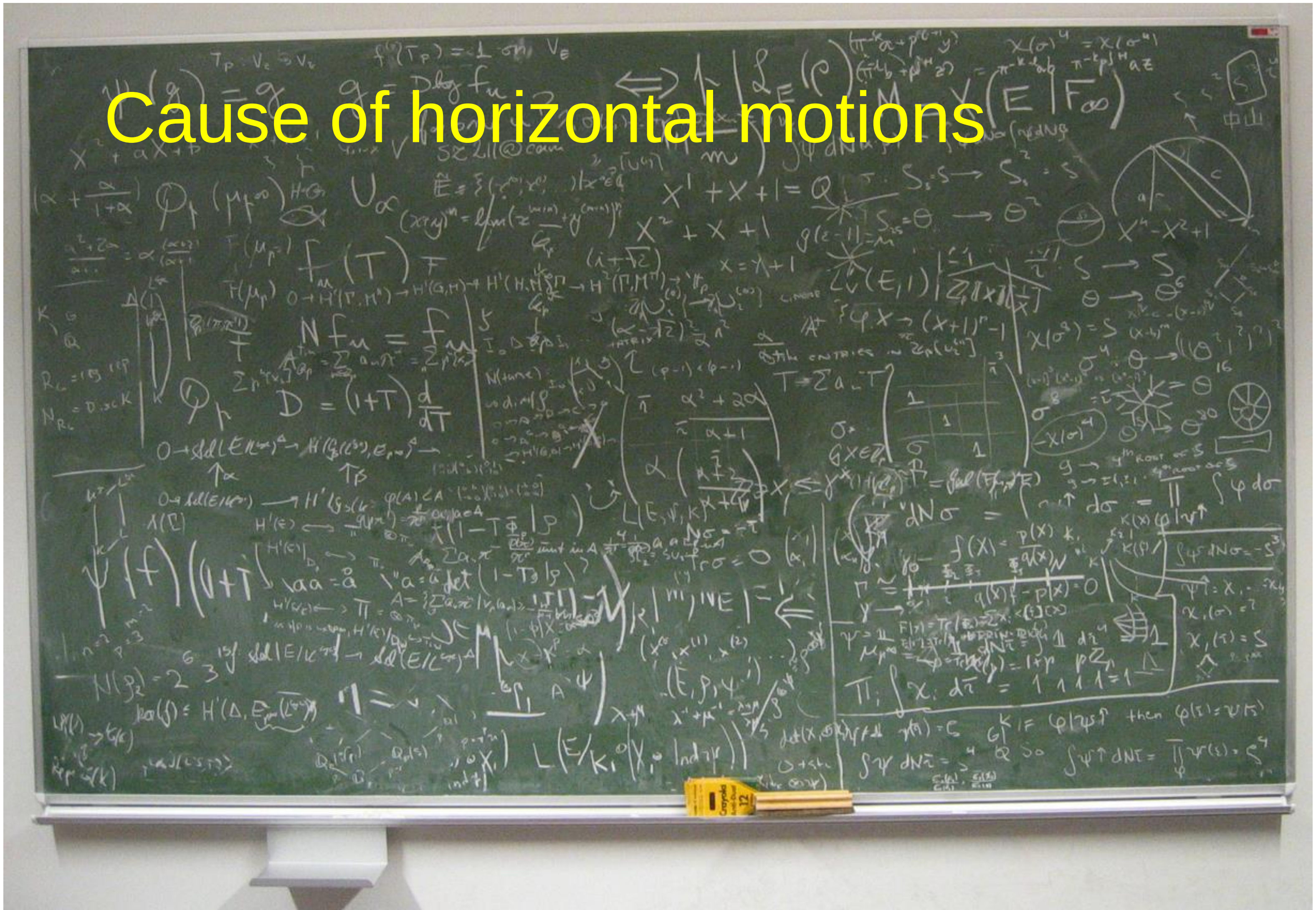


If the inversion layer between A and B is lifted, the lower (moister) part of the layer reaches saturation first and from then on cools less than the upper part. This leads to de-stabilization and may cause deep convection.

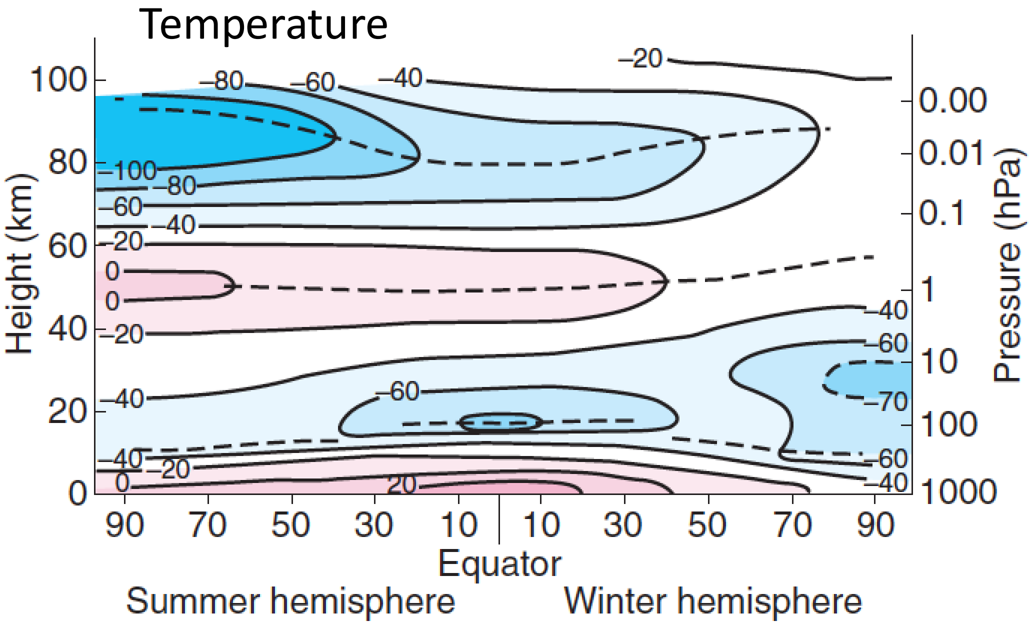
At the same pressure and temperature is an air parcel

A. Heavier when moist
B. Lighter when moist

Black board…



Seasonal Distribution of Temperature and Wind



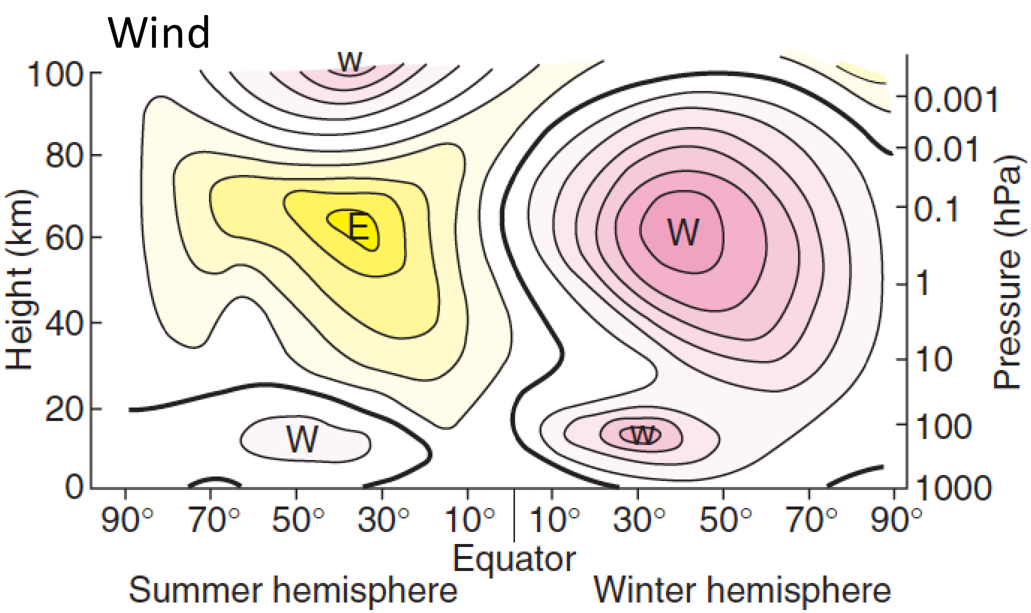
Mesopause

Stratopause

Tropopause

North – South Gradients

drive

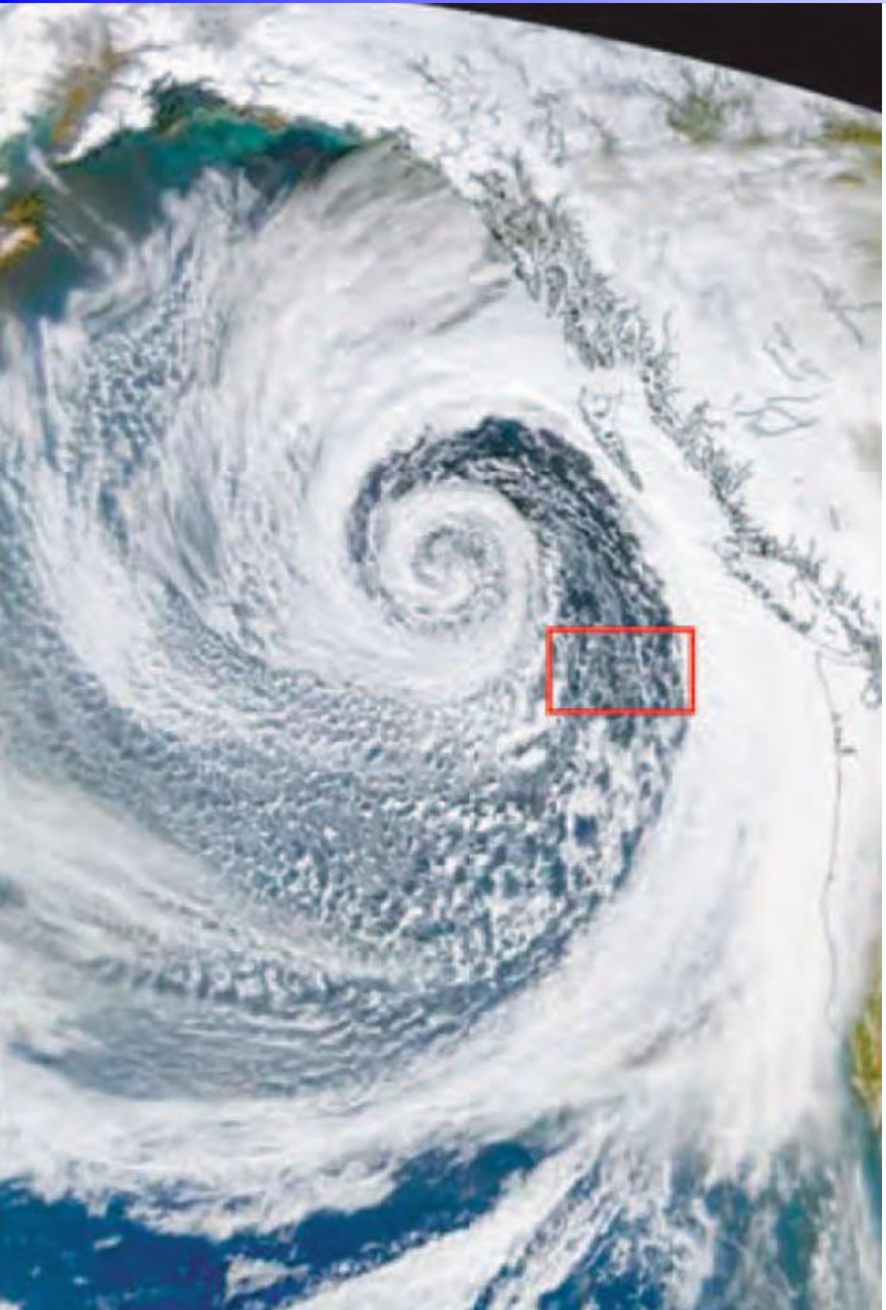


W: Westerlies

E: Easterlies

Global Wind Patterns with tropospheric and mesospheric jet streams

Extratropical Cyclone



North – South differences in radiation and therefore temperatures lead to westerly winds and “baroclinic instabilities”, i.e. waves that develop fronts and ultimately bring warm air to the North and cold air to the South. Size of the system is approximately 2000 km.

Horizontal Movements

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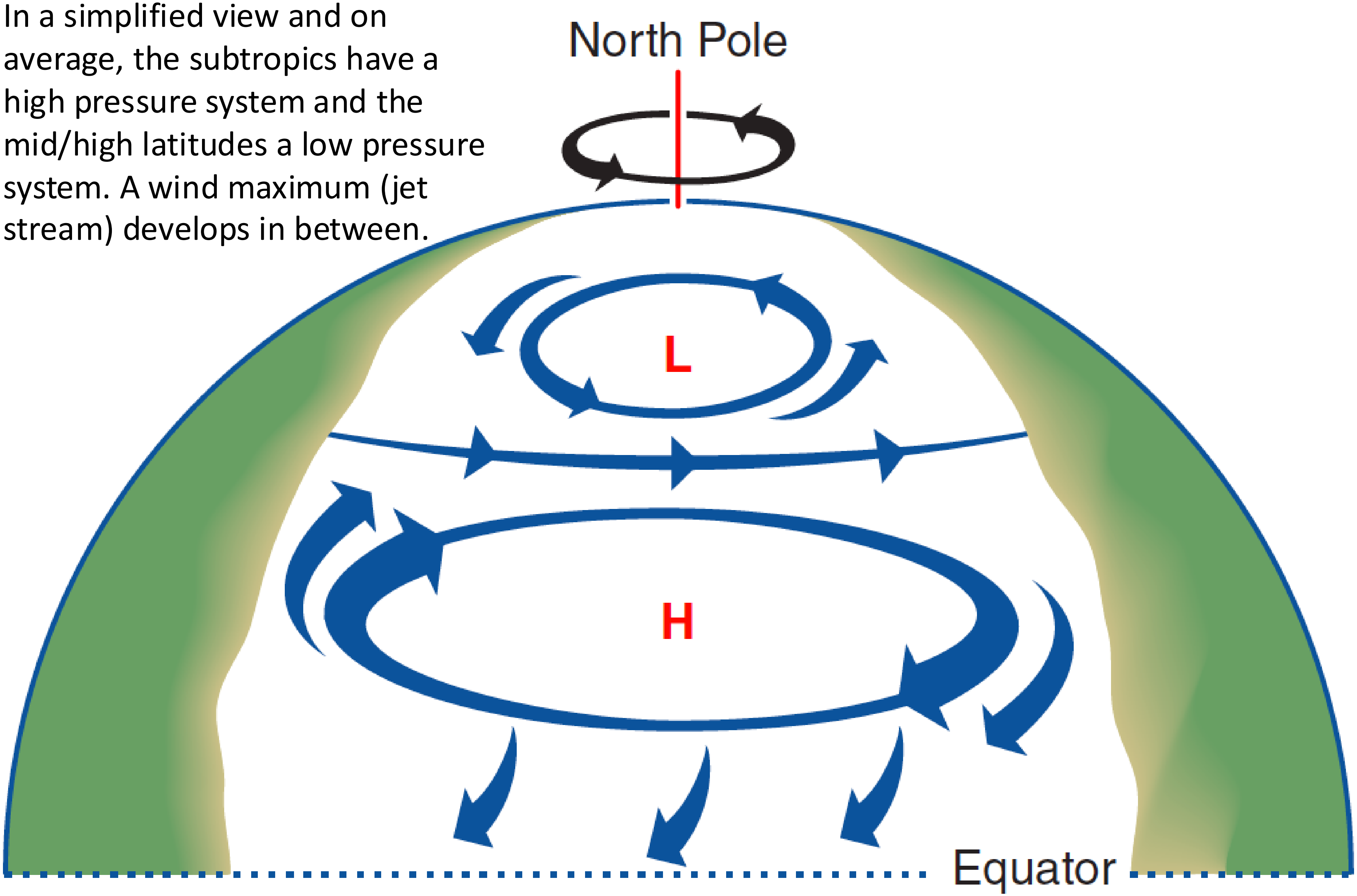
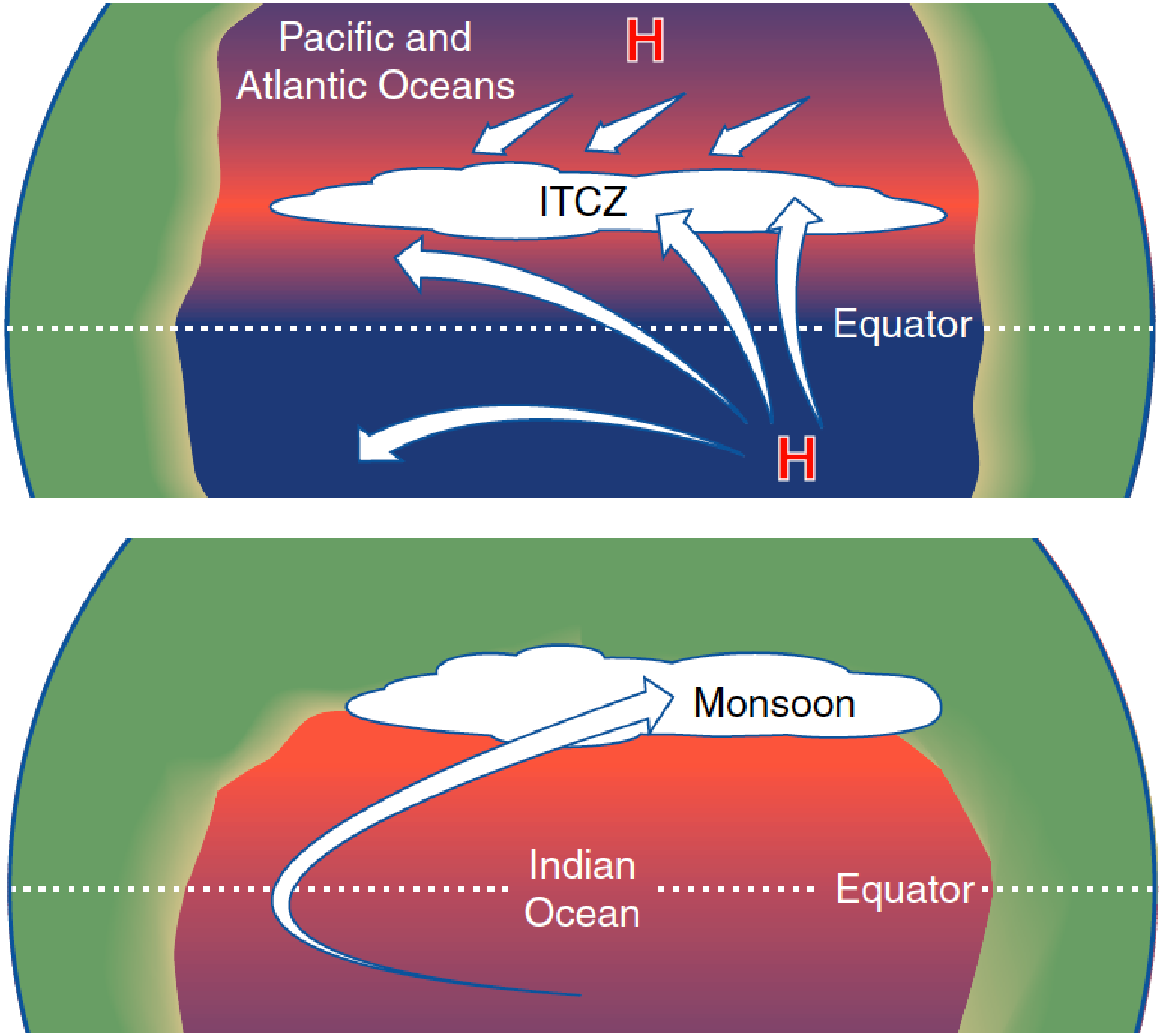
In a simplified view and on average, the subtropics have a high pressure system and the mid/high latitudes a low pressure system. A wind maximum (jet stream) develops in between.



Horizontal Movements with Variations

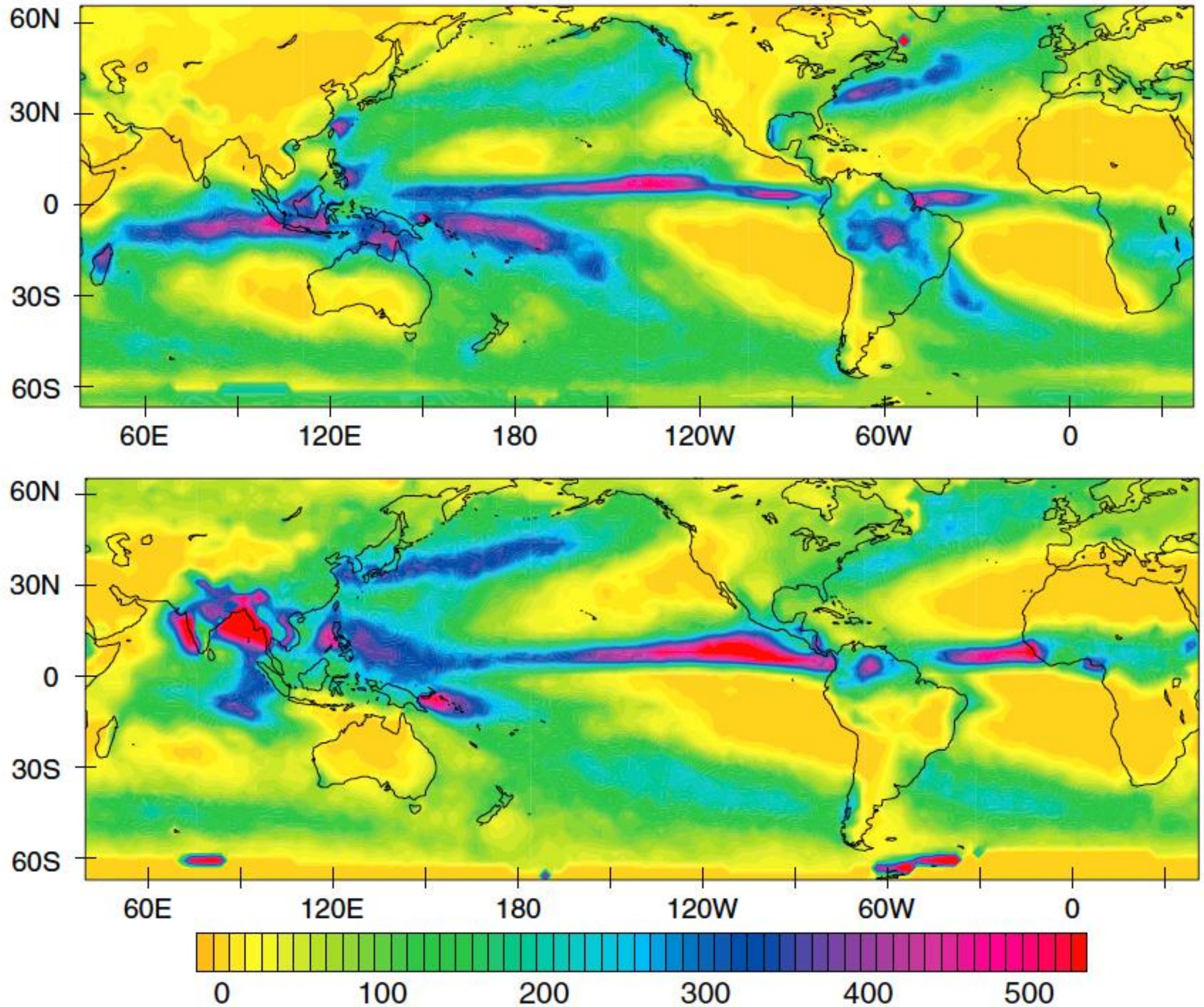


The simple system explained in the previous slide is modified by seasonal variations as well as land – sea distributions and mountains.

Precipitation

January

July



Precipitation maxima are achieved either along the Inner Tropical Convergence Zone (ITCZ) or along the dominant storm tracks (Westerlies). Persistent high-pressure zones (e.g. Sahara) are persistently dry.

Take Home Messages

- Atmosphere is thin and air is thinning quickly with height as described by the hydrostatic equation

- Vertical temperature gradient (= lapse rate) is determined by a) chemistry (structure of the atmosphere), b) dry adiabatic (Γ) and moist (Γ_s) vertical movements and c) surface energy balance

- Know what relative humidity, mixing ratio and potential temperature are

- Instability is characterized by a potential temperature gradient smaller than zero, stable conditions occur for potential temperature gradients larger than zero

- Instability can be caused by lifting, surface heating or moisture dynamics

- A moist adiabatic lapse rate indicates less cooling with height than a dry adiabatic lapse rate

- Atmospheric Motions are caused by energy redistribution, which first cause vertical and then (secondary) horizontal motions

What you need to know / be able to do

- Understand adiabatic motions and potential temperature

- Make simple calculations involving lapse rates, potential temperature and pressure levels

- Work with thermodynamic diagrams and use them to assess vertical air parcel motions

- Know causes and effects of vertical motions in the atmosphere

NEXT: Large-Scale Horizontal Motions and Atmospheric Dynamics