AOSC400-2015: Lectures # 18 and # 19 -Review

Chapter 9:

- Connection between Chapter 4 and Chapter 9
- Components of the Surface Energy Budget (S**E**B)-in addition to radiation also fluxes of *heat and moisture.*
- Fluxes of heat and moisture are transported by turbulence in the PLB
- What is Planetary Boundary Layer (PBL)? Concept of Capping Inversion
- What is turbulence? How do we represent it? Reynold's Averaging.

Supplementary background material for understanding Chapter 9

- Adiabatic Lapse Rate
- Environmental Lapse Rate
- Concept of stability
- Hydrostatic balance
- **Potential temperature - will be covered in Lecture # 20**

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*Sources of information used for this lecture are listed in updated Syllabus.

Potential temperature

The *potential temperature* ϑ of an air parcel is defined as the temperature which the parcel of air would have if it were expanded or compressed adiabatically from its existing pressure and temperature to a standard pressure p_0 (generally taken as 1000 mb).

Potential temperature is denoted as θ and, for air is given by

$$\theta = T\left(\frac{P_0}{P}\right)^{\frac{R}{c_p}}$$

where *T* is the absolute temperature (in K) of the parcel, *R* is the gas constant of air, and c_p is the specific heat capacity at a constant pressure. This equation is often known as *Poisson's* equation.

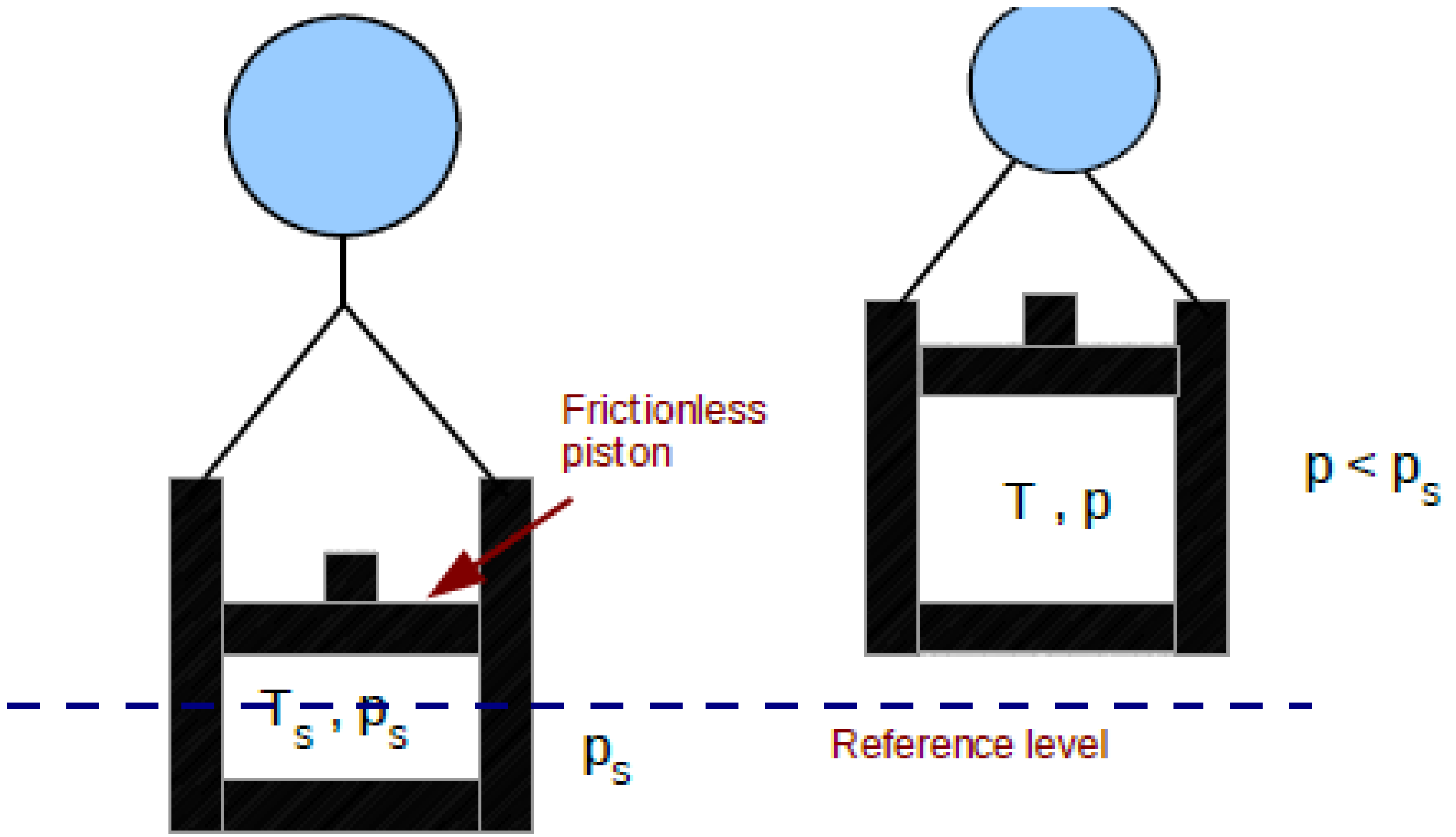
Potential temperature is conserved for all dry adiabatic processes.

Above equation assumes an adiabatic process and :

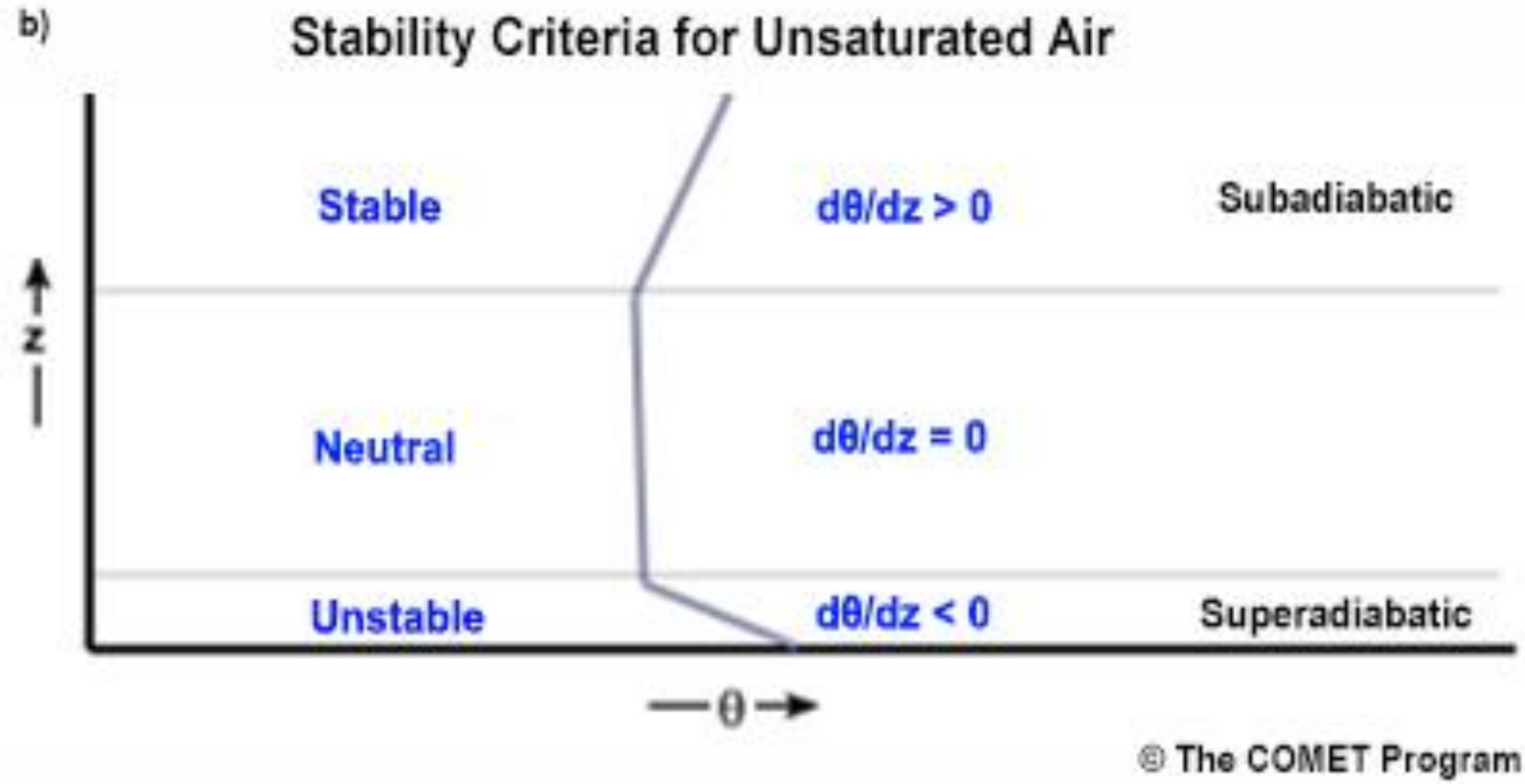
- Hydrostatic balance
- Ideal gas: $\rho = p/RT$,

Can explain with the help of figure in slide # 4:

If we extract an air parcel out of the environment at level z and push it down adiabatically to the reference level where pressure is p_S, then the temperature Ts of the parcel when it reaches the reference level (denoted as ϑ) is the potential temperature. Potential temperature is a useful measure of the static stability of the unsaturated atmosphere.

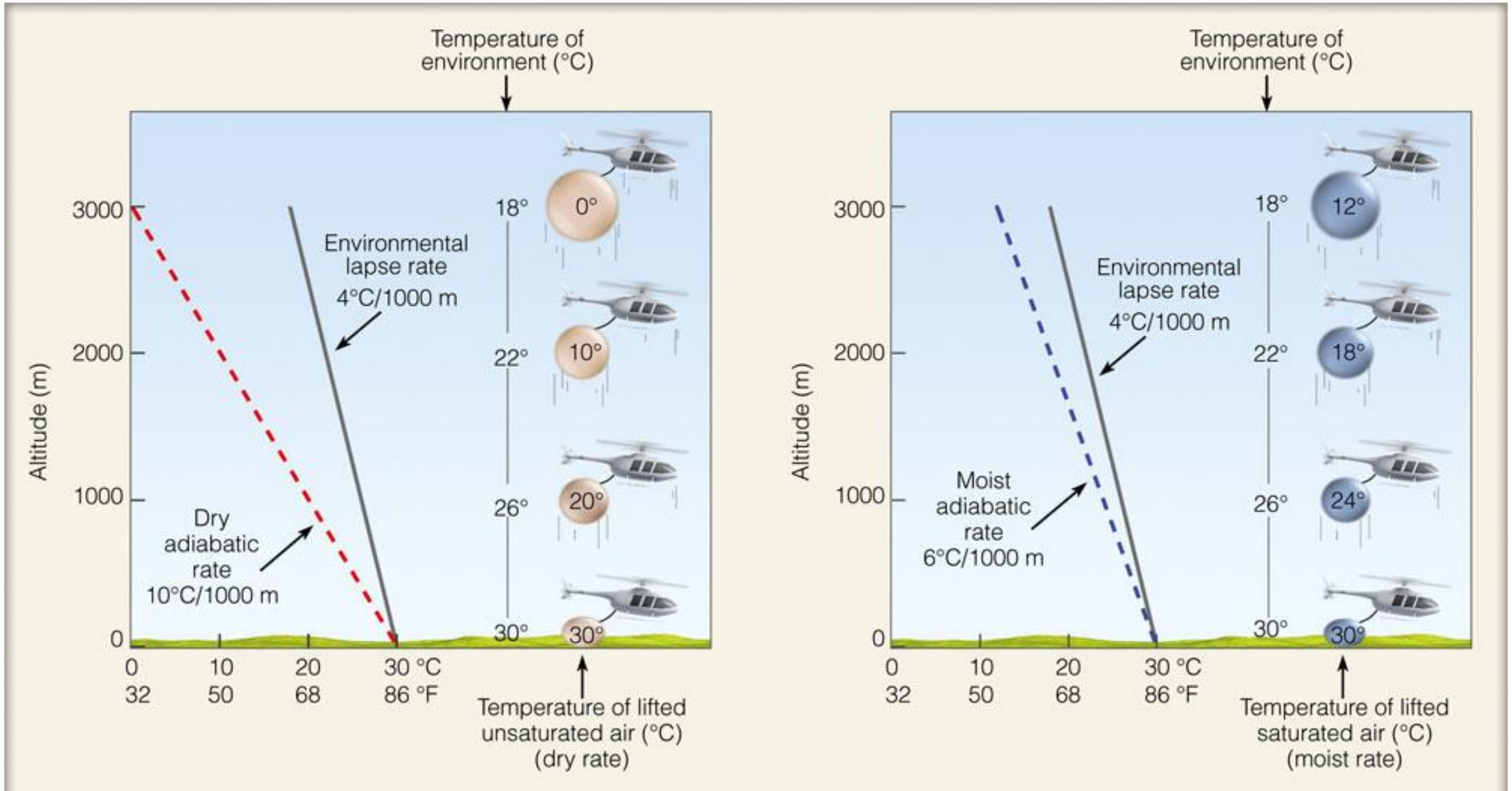


If we let the apparatus fall back (adiabatically) to the level of $p = p_S$, the temperature in the chamber will be restored to $T_S = \vartheta(s)$



Common to use stability criteria for unsaturated air based on the vertical gradient of the potential temperature, θ.

Discussed stability before as shown here: Ability of the parcel to rise depends on: Relationship between environmental lapse rate and *dry adiabatic lapse rate or* moist adiabatic lapse rate.



(a) Lifted, unsaturated air at each level is colder and heavier than the air around it. If released, the parcel would return to its original position, the surface.

(b) Lifted, saturated air at each level is colder and heavier than the air surrounding it. If released, the parcel would return to its original position, the surface.

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AOSC400-2015
November 24, Lecture # 20

- Planetary Boundary Layer (PBL) or Atmospheric Boundary Layer (ABL)
- Turbulent and Laminar Flow
- Concept of stability and Richardson Number
- Eddy Covariance Flux Measurements
- Turbulent Kinetic Energy

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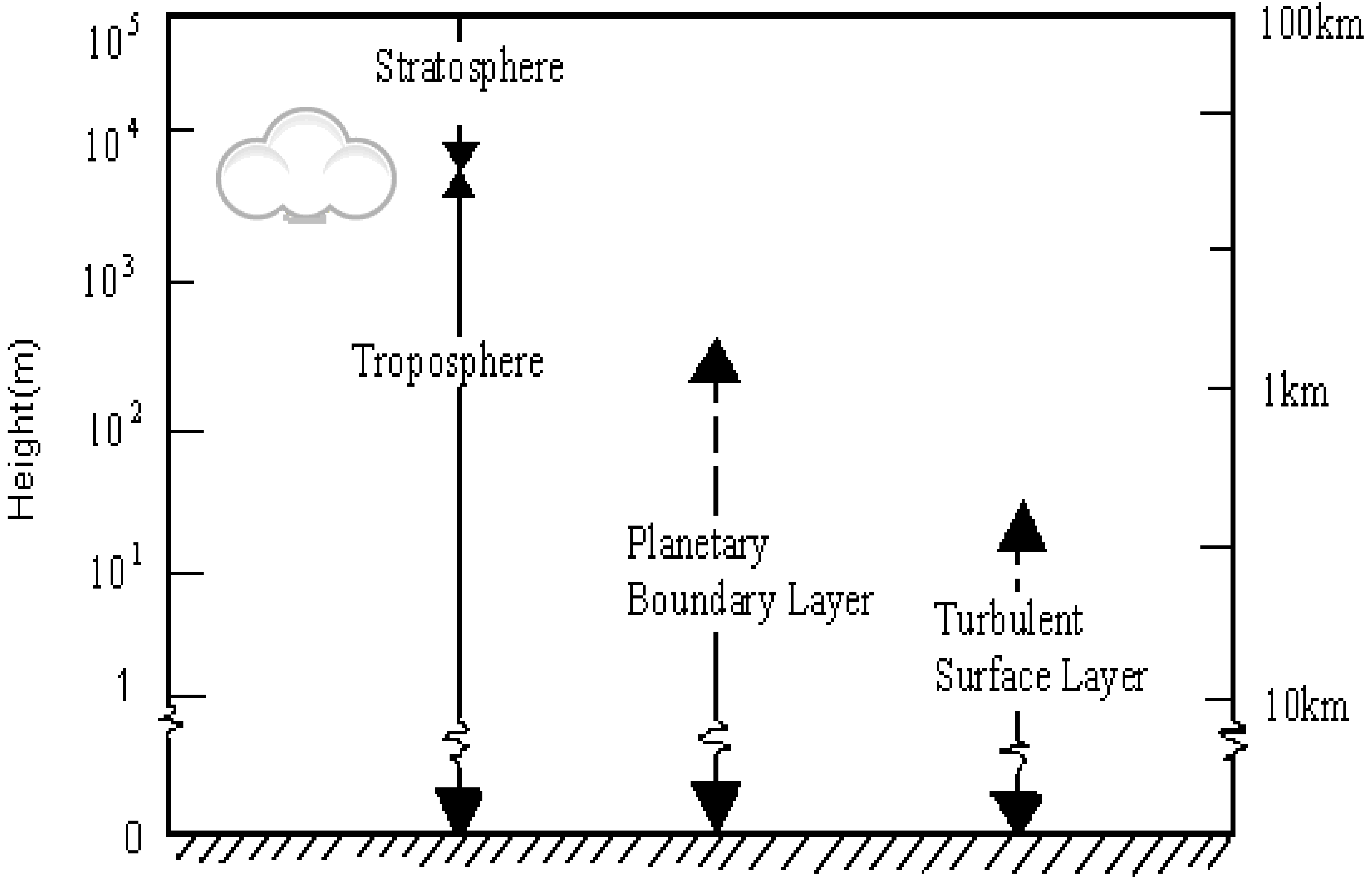
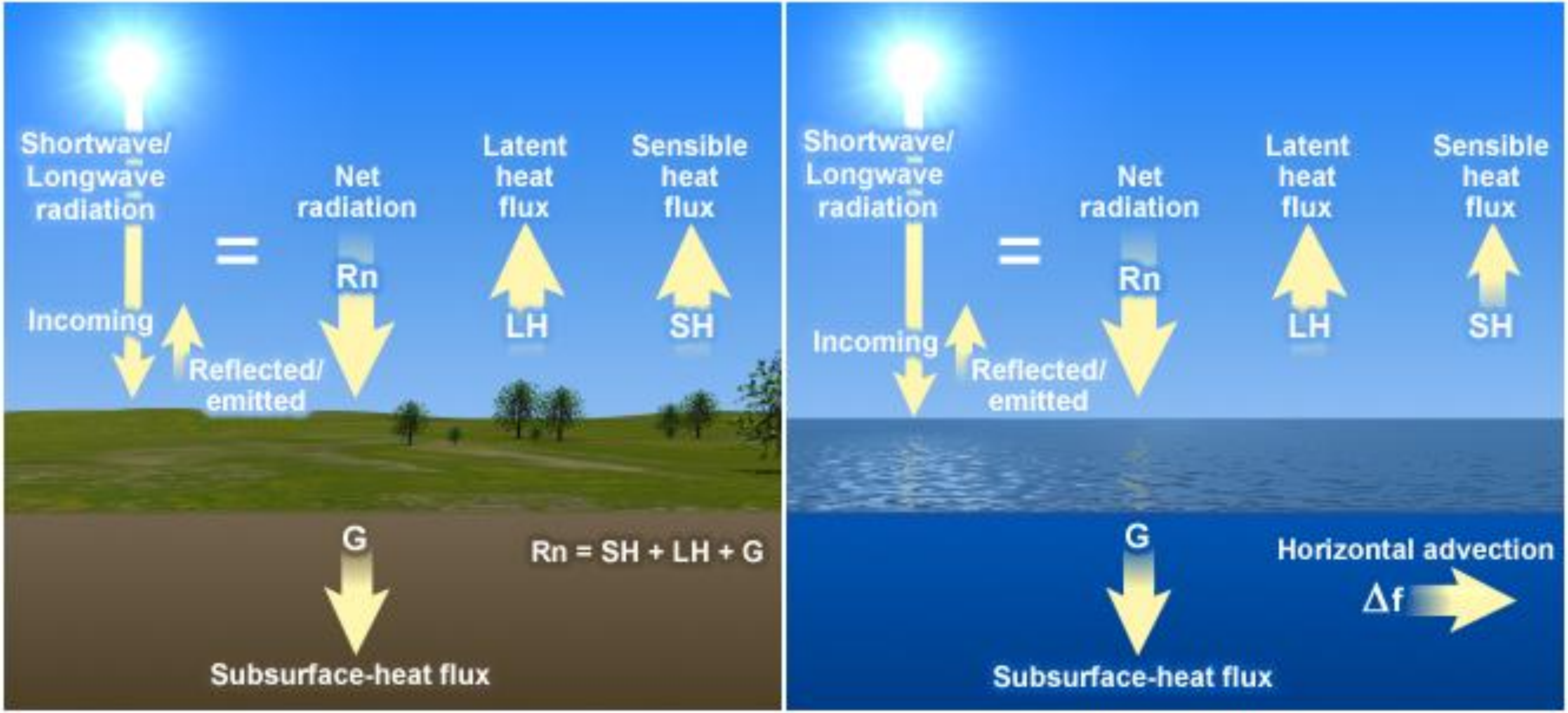
The troposphere can be subdivided as follows:



- The portion of the atmosphere most affected by Earth's surface is called the Atmospheric Boundary Layer (ABL)
- ABL, also known as the Planetary Boundary Layer (PBL), is the lowest layer of the troposphere.
- The ABL is in contact with the surface and experiences frictional effects.
- Heat, moisture, momentum, aerosols, and gases are exchanged between the free atmosphere and the surface through the ABL. The boundary layer depth varies from 10s of meters over tropical oceans to several km over hot, dry continents but the typical height is about 1 km.

Vertical Transport

- Heat energy drives the physical cycles of the earth system.
- Solar radiation is absorbed by the surface then energy is transferred from the surface to the troposphere by latent heat, longwave radiation, and sensible heat.
- Latent heat energy stored in water vapor and released into the atmosphere with condensation, is the primary means of surface-to-atmosphere energy transport.
- Sensible heat flux from the surface, which directly warms the atmosphere, while small in the annual global heat budget, is a significant energy source during the daytime over land.



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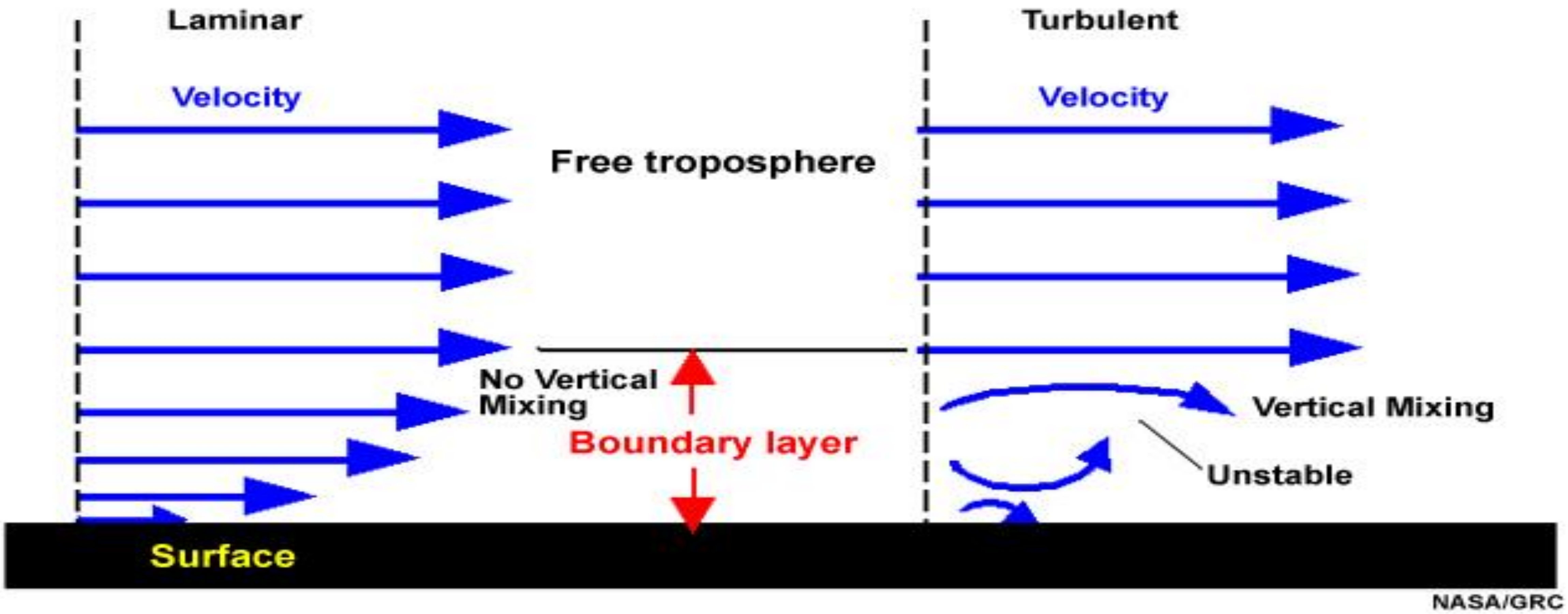
Surface *energy* budget for land and ocean surfaces during the daytime.

Turbulent and Laminar Flow



Schematic of laminar and turbulent flow in the boundary layer; velocity is zero at the surface.

[Introduction to Tropical Meteorology *2nd Edition*](), *Produced by* The COMET Program

With turbulent flow, there is vertical mixing, while with laminar flow mixing between layers is non-existent or negligible.
For the atmosphere, the Richardson number, Ri, used to evaluate whether an atmospheric layer is predominantly turbulent or laminar; it is the ratio between the *static stability* and *shear stability*.

$$Ri = \frac{buoyancy}{mechanical} = \frac{N^2}{(\partial u/\partial z)^2} \quad where\ N^2 = -(g/\rho)d\rho/dz \ or\ (g/\theta_v)d\theta_v/dz \qquad (1)$$

Here u is horizontal velocity, z is the height, g is the acceleration due to gravity, ρ is the density as a function of height. In general:

Laminar flow becomes turbulent when Ri < 0.25
Turbulent flow becomes laminar when Ri > 1.0

Sounding data and a finite difference estimate of equation (1) is used to calculate the Richardson number in practice. This form, known as the bulk Richardson number, is:

Eq. (2)

$$Ri = \frac{g \Delta \theta_v \Delta z}{\theta_v [(\Delta u)^2 + (\Delta v)^2]}$$

where Δz is the depth of the layer of interest.

Now we will explain how heat is transferred by turbulence in the atmosphere.

As seen before (when we discussed Reynolds averaging) an average over a time period T (say half an hour) is :

$$\overline{u} = \frac{1}{N}\sum_{i=1}^{N} u_i \tag{9.2}$$

In the atmosphere, this mean value can change from one half-hour period to the next, resulting in a slow variation of the mean-wind components with time.

Subtracting the mean from the instantaneous component u gives just the fluctuating (gust) portion of the flow (indicated with a prime)

$$u_i' = u_i - \overline{u} \tag{9.3}$$

The intensity of turbulence in the u direction is then defined by the variance:

$$\sigma_u^2 = \frac{1}{N}\sum_{i=1}^{N}[u_i - \overline{u}]^2 = \frac{1}{N}\sum_{i=1}^{N}[u_i']^2 = \overline{[u']^2} \quad (9.4)$$

In the atmosphere, fluctuations in velocity are often accompanied by fluctuations in scalar values such as temperature, humidity, or pollutant concentration.

For example, in a field of thermals there are regions where warm air is rising (positive potential temperature ϑ' accompanies positive vertical velocity w'), surrounded by regions where cold air is sinking (negative ϑ' accompanies negative w'). One measure of the amount that ϑ and w vary together is the covariance (cov).

If warm air parcels are rising and cold parcels are sinking, as in a thermally direct circulation, then $\overline{w'\theta'} > 0$. Covariances can also be negative or zero

$$\mathrm{cov}\,(w,\theta) = \frac{1}{N}\sum_{i=1}^{N}[(w_i - \overline{w})\cdot(\theta_i - \overline{\theta})]$$

$$= \frac{1}{N}\sum_{i=1}^{N}[(w'_i)\cdot(\theta'_i)] = \overline{w'\theta'} \qquad (9.5)$$

What is Flux?

Flux – how much of something moves through a unit area per unit time

Flux is dependent on: (1) number of things crossing the area; (2) size of the area being crossed, and (3) the time it takes to cross this area

Flux Measurements

- Flux measurements are widely used to estimate heat, water, and CO_2 exchange, as well as methane and other trace gases
- Eddy Covariance is one of the most direct and defensible ways to measure such fluxes
- The method is mathematically complex, and requires a lot of care setting up and processing data

.

"A Brief Practical Guide to Eddy Covariance Flux Measurements: Principles and Workflow Examples for Scientific and Industrial Applications"

G. Burba and D. Anderson of LI-COR Biosciences

Objective:

- To help a non-expert gain a basic understanding of the Eddy Covariance method and to point out valuable references

- To provide explanations in a simplified manner first, and then elaborate with specific details

Mathematical foundation

In mathematical terms, "eddy flux" is computed as a covariance between instantaneous deviation in vertical wind speed (w') from the mean value (w-overbar) and instantaneous deviation in gas concentration, mixing ratio (s'), from its mean value (s-overbar), multiplied by mean air density (ρa). Several mathematical operations and assumptions, including Reynolds decomposition, are involved in getting from physically complete equations of the turbulent flow to practical equations for computing "eddy flux", as shown below.

Eddy covariance system consisting of an ultrasonic anemometer and infrared gas analyser (IRGA).

- The Eddy Covariance method is one of the most accurate, direct and defensible approaches available to date for measurements of gas fluxes and monitoring of gas emissions from areas with sizes ranging from a few hundred to millions of square meters
- The method relies on direct and very fast measurements of actual gas transport by a 3-D wind speed in real time *in situ*, resulting in calculations of turbulent fluxes within the atmospheric boundary layer

Several networks have been established over the globe to measure turbulent fluxes.

Existing Flux Networks:

Fluxnet, Fluxnet-Canada, AsiaFlux, CarboEurope and AmeriFlux networks

They collect Eddy Covariance information.

http://gcmd.nasa.gov/records/GCMD_AMERIFLUX_SHIDLER-CDIAC.html

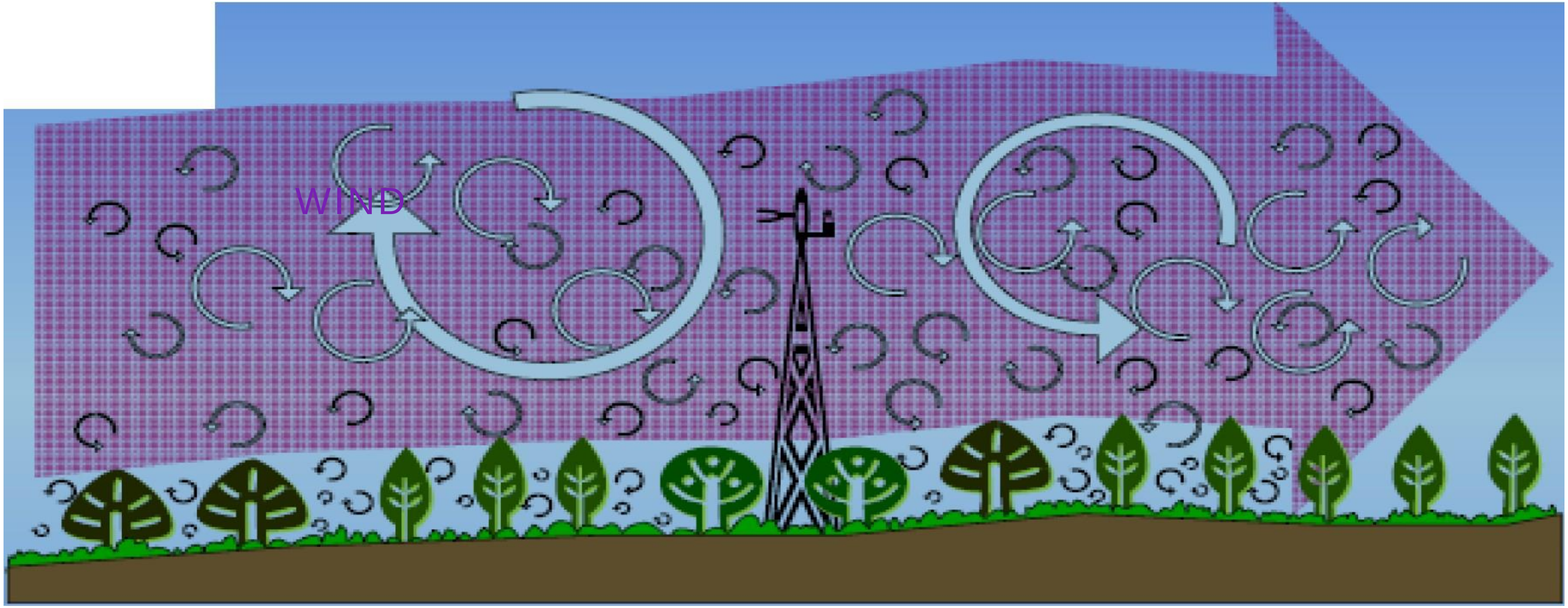
http://terraweb.forestry.oregonstate.edu/chair2. htm

About the AmeriFlux Network

AmeriFlux is a network of PI-managed sites measuring ecosystem CO_2, water, and energy fluxes in North and South America. It was established to connect research on field sites representing major climate and ecological biomes, including tundra, grasslands, savanna, crops, and conifer, deciduous, and tropical forests. As a grassroots, investigator-driven network, the AmeriFlux community has tailored instrumentation to suit each unique ecosystem. This “coalition of the willing” is diverse in its interests, use of technologies and collaborative approaches.

The **eddy covariance** (also known as **eddy correlation** and **eddy flux**) technique is a key atmospheric measurement technique to measure and calculate vertical turbulent fluxes within atmospheric boundary layers.
The method analyzes high-frequency wind and scalar atmospheric data series, and yields values of fluxes of these properties.
It is a statistical method used in meteorology and other applications (micrometeorology, oceanography, hydrology, agricultural sciences, industrial and regulatory applications, etc.) to determine exchange rates of trace gases over natural ecosystems, agricultural fields, and to quantify gas emissions rates from other land and water areas. It is particularly frequently used to estimate momentum, heat, water vapour, carbon dioxide and methane fluxes

The technique is also used extensively for verification and tuning of global climate models, mesoscale and weather models, complex biogeochemical and ecological models, and remote sensing estimates from satellites and aircraft. The technique is mathematically complex, and requires significant care in setting up and processing data. To date, there is no uniform terminology or a single methodology for the Eddy Covariance technique, but much effort is being made by flux measurement networks (e.g., Fluxnet, Ameriflux, ICOS, CarboEurope, Fluxnet Canada, OzFlux, NEON, and iLEAPS) to unify the various approaches.



- Airflow can be imagined as a horizontal flow of numerous rotating eddies
- Each eddy has 3-D components, including a vertical wind component
- The diagram looks chaotic but components can be measured from tower

EDDY COVARIANCE INSTRUMENTATION

In turbulent flow, vertical flux can be presented as:
($s=\rho_c/\rho_a$ is a mixing ratio of substance 'c' in the air)

$$F = \overline{\rho_a w s}$$

Reynolds decomposition is used then to break into means and deviations:

$$F = \overline{(\bar{\rho}_a + \rho'_a)(\bar{w} + w')(\bar{s} + s')}$$

Open parenthesis:

$$F = \overline{(\bar{\rho}_a \bar{w} \bar{s} + \cancel{\bar{\rho}_a \bar{w} s'} + \cancel{\bar{\rho}_a w' \bar{s}} + \bar{\rho}_a w' s' + \cancel{\rho'_a \bar{w} \bar{s}} + \rho'_a \bar{w} s' + \rho'_a w' \bar{s} + \rho'_a w' s')}$$

Averaged deviation from the average is zero

Equation is simplified: $F = (\bar{\rho}_a \bar{w} \bar{s} + \bar{\rho}_a \overline{w' s'} + \bar{w}\, \overline{\rho'_a s'} + \bar{s}\, \overline{\rho'_a w'} + \overline{\rho_a' w' s'})$

Then important assumption is made (for conventional Eddy Covariance) – density fluctuations are assumed negligible:

$$\boxed{F = (\overline{\rho_a}\,\overline{w}\,\overline{s} + \overline{\rho_a}\,\overline{w's'} + \overline{w}\,\cancel{\overline{\rho'_a s'}} + \overline{s}\,\cancel{\overline{\rho'_a w'}} + \cancel{\overline{\rho_a' w' s'}}) = \overline{\rho_a}\,\cancel{\overline{w}}\,\overline{s} + \overline{\rho_a}\,\overline{w's'}}$$

Then another important assumption is made – mean vertical flow is assumed negligible for horizontal homogeneous terrain (no divergence/convergence):

$$\boxed{F \approx \overline{\rho_a}\,\overline{w's'}}$$

'Eddy flux'

Summary of Eddy Covariance Theory

- Measures fluxes transported by eddies
- Requires turbulent flow
- Requires state-of-the-art instruments
- Calculated as covariance of w' and c'
- Many assumptions to satisfy
- Complex calculations
- Most direct way to measure flux
- Continuous new developments

TERRESTRIAL

98% of applications

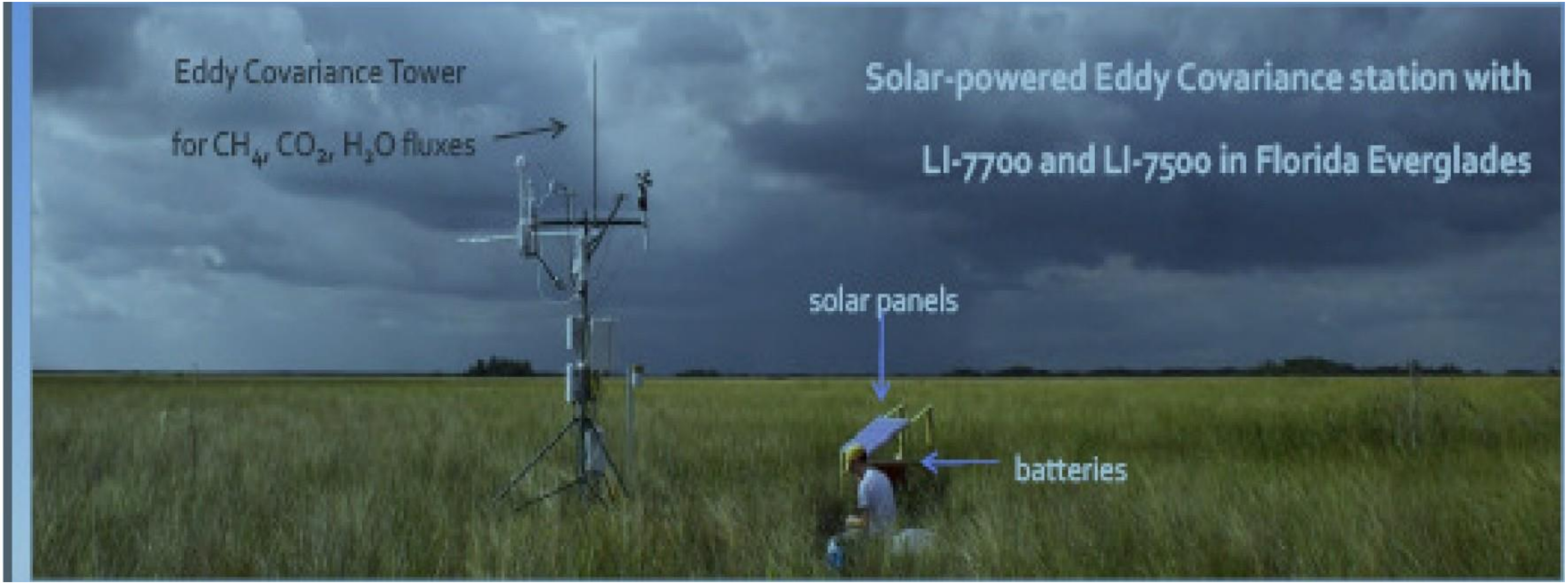
AIRBORNE

<1% of applications

OCEANOGRAPHIC

<1% of applications

OPEN-PATH LI-7700 CALIBRATION



Turbulence kinetic Energy and Turbulence Intensity

Kinetic energy is defined as:

$$KE = \tfrac{1}{2} m V^2,$$

where m is mass and V is velocity. In meteorology we often use specific kinetic energy, namely KE/m, or the kinetic energy per unit mass. Similarly, we can define specific kinetic energy associated with turbulent fluctuations :

$$\frac{TKE}{m} = \frac{1}{2}\left[\overline{u'^2} + \overline{v'^2} + \overline{w'^2}\right]$$

where TKE is turbulence kinetic energy. For laminar flow, which contains no micro-scale motions, TKE = 0, even though u, v, w are not necessarily zero. Larger values of TKE indicate a greater intensity of the microscale turbulence. We see now that the three components of velocity variance represent three contributions to the scalar TKE.

Intensity of Turbulence

The intensity of turbulence is measured in terms of the turbulent kinetic energy (TKE), ē,

Eqn (3)

$$\bar{e} = \frac{1}{2}\left(\overline{u'^2} + \overline{v'^2} + \overline{w'^2}\right) \text{ or in terms of the variance}, \sigma\text{:} \quad \bar{e} = \frac{1}{2}(\sigma_u^2 + \sigma_v^2 + \sigma_w^2)$$

where u, v, and w are the zonal, meridional, and vertical wind components, respectively. The overbar represents the time average and the prime is the perturbation or eddy contribution. TKE is zero for laminar flow, even though the mean wind components are not necessarily zero.

Using what we already know about mechanical and thermal generation of turbulence, and of viscous dissipation, we can write in descriptive form an Eulerian (i.e., fixed relative to the ground) forecast equation for turbulence kinetic energy:

$$\frac{\partial(TKE/m)}{\partial t} = Ad + M + B + Tr - \varepsilon \quad (9.7)$$

where

$$Ad = -\overline{u}\,\frac{\partial(TKE/m)}{\partial x} - \overline{v}\,\frac{\partial(TKE/m)}{\partial y} - \overline{w}\,\frac{\partial(TKE/m)}{\partial z}$$

the advection of TKE by the mean wind, M is mechanical generation of turbulence, B is buoyant generation or consumption of turbulence, Tr is transport of turbulence energy by turbulence itself, and ε is the viscous dissipation rate.