Lesson 1:
Planetary Boundary Layer

Course: Laboratory of Atmospheric Remote Sensing
Laurea Magistrale in Atmospheric Science and Technology

CONTENTS (1)

- Planetary Boundary Layer: definition and main characteristics, mean wind, waves and turbulence
- Turbulence: mechanical and thermal turbulence, buoyancy, adiabatic and environmental lapse rate
- Atmospheric stability
- Variables of interest: Reynolds number, Reynolds decomposition, standard deviation of fluctuating velocity, Taylor's hypothesis, convective velocity scale, friction velocity, Monin-Obukhov length
- Monin-Obukhov similarity theory
- PBL depth over oceans, low and high pressure regions
- PBL vertical structure and daily evolution
- Convective PBL: main characteristics, height and turbulent velocities
- Stable PBL: main characteristics, height and turbulent velocities

CONTENTS (2)

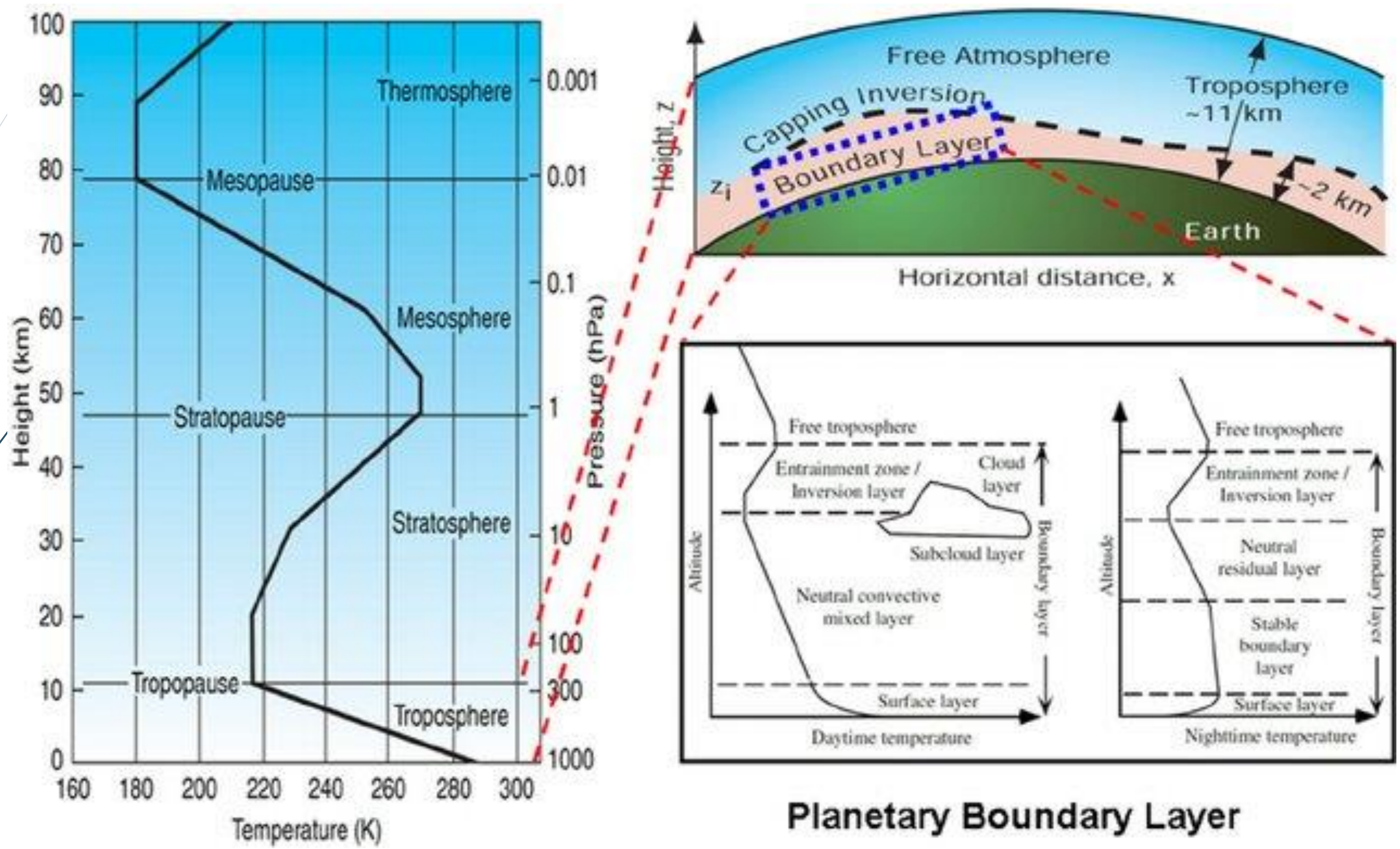
- Micrometeorology
- Urban boundary layer: definition, vertical structure, aerodynamic roughness length, zero displacement height
- Flow around an isolated building
- Flow regime in 2D and 3D street canyons
- Dispersion and stability

Reading material:

R.B. Stull ''An Introduction to Boundary Layer Meteorology''

Chapter 1

PLANETARY BOUNDARY LAYER



Planetary Boundary Layer

PLANETARY BOUNDARY LAYER

The **Atmospheric Boundary Layer** (**ABL**), or **Planetary Boundary Layer** (**PBL**), is the lowest 1-3 km of the atmosphere, where the transport processes modify the structure and the composition of the atmosphere.

The remnant of the atmosphere above is termed “free” atmosphere.

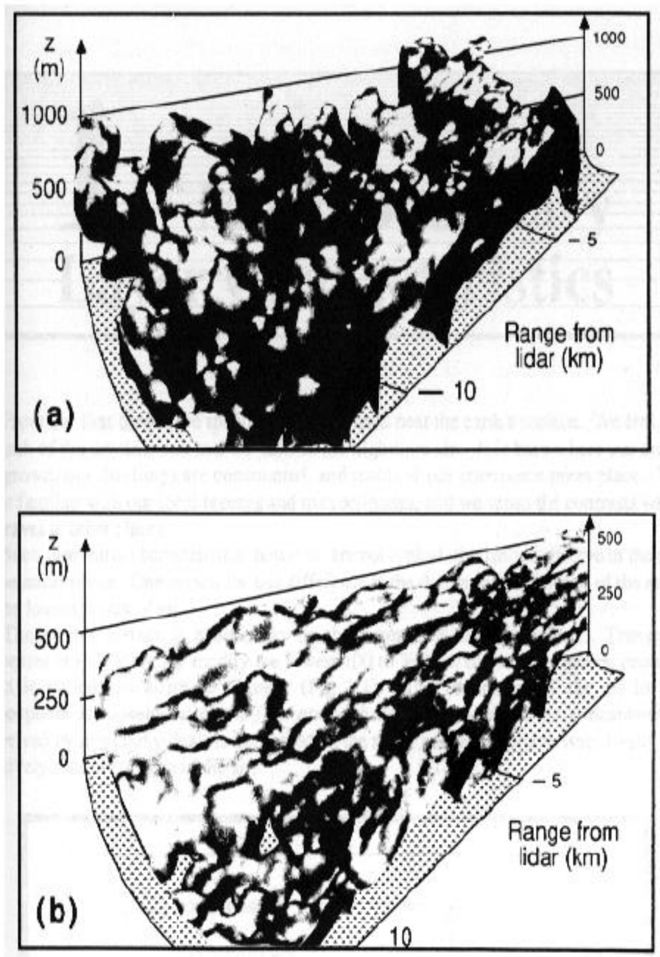
The PBL is defined as the part of the troposphere directly influenced by the presence of the earth’s surface and responds to surface forcings with a timescale of about an <u>hour or less</u> and with typical spatial scales of a <u>few kilometers.</u>

The main forcings are:

- heat transfer from/to the ground
- evapotranspiration
- frictional drag
- pollutant emissions
- flow modification due to topography

PLANETARY BOUNDARY LAYER

Day time boundary layer is usually very turbulent, due to ground-level heating.



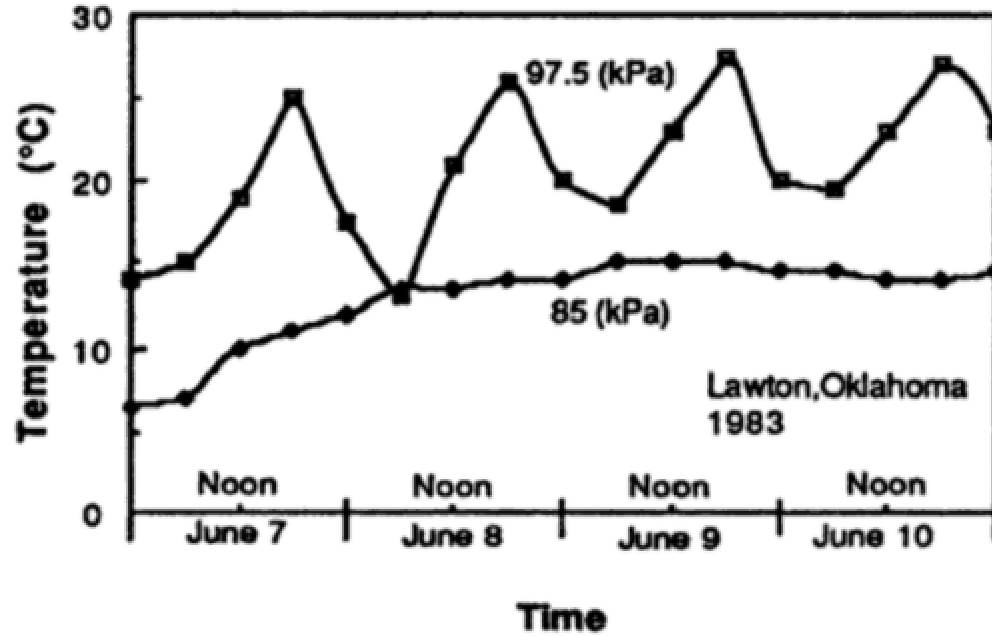
Lidar images of the aerosol-laden boundary layer, obtained during the FIFE field experiment in Kansas. (a) Convective mixed layer observed at 1030 local time on 1 July 1987, when winds were generally less than 2 m/s. (b) Slightly-stable boundary layer with shear-generated turbulence, observed at 530 local time on 7 July 1987. Winds ranged from 5 m/s near the surface to 15 m/s near the top of the boundary layer. Photographs from the Univ. of Wisconsin lidar are courtesy of E. Eloranta, Boundary Layer Research Team.

COMPARISON OF BOUNDARY LAYER AND THE FREE ATMOSPHERE CHARACTERISTICS

Property	Boundary Layer	Free Atmosphere
Turbulence	• Almost continuously turbulent over its whole depth.	• Turbulence in convective clouds, and sporadic CAT in thin layers of large horizontal extent.
Friction	• Strong drag against the earth's surface. Large energy dissipation.	• Small viscous dissipation.
Dispersion	• Rapid turbulent mixing in the vertical and horizontal.	• Small molecular diffusion. Often rapid horizontal transport by mean wind.
Winds	• Near logarithmic wind speed profile in the surface layer. Subgeostrophic, cross-isobaric flow common.	• Winds nearly geostrophic.
Vertical Transport	• Turbulence dominates.	• Mean wind and cumulus-scale dominate
Thickness	• Varies between 100 m to 3 km in time and space. Diurnal oscillations over land.	• Less variable. 8-18 km. Slow time variations.

PLANETARY BOUNDARY LAYER (PBL)

Fig. 1.2
Evolution of temperatures measured near the ground (97.5 kPa) and at a height of roughly 1100 m above ground (85 kPa). Based on rawinsonde lauches from Ft. Sill, OK.

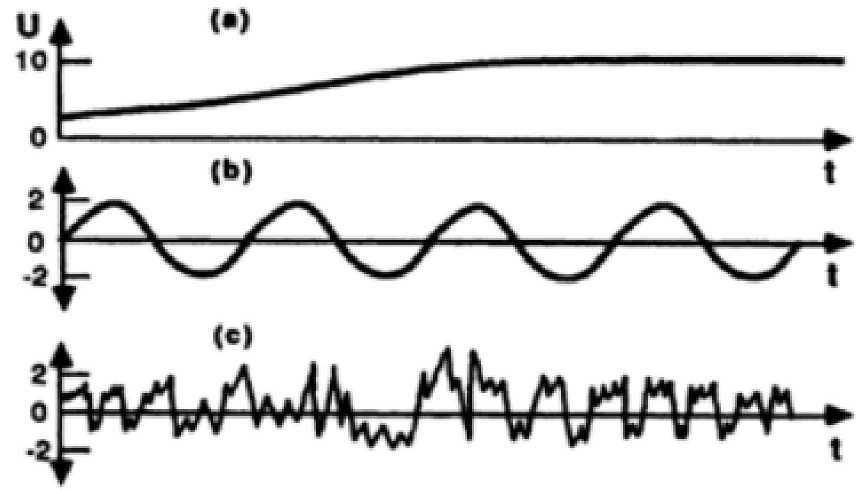


Temporal trends in the lower troposphere obtained by rawinsonde (wind speed and direction) soundings made every several hours in Oklahoma.

Near the ground the net diurnal variation is due to the absorption of solar radiation by the ground, that warms and cools in response of the radiation, which in turn forces changes in the UBL via transport processes.
In the free atmosphere the diurnal cycle of air temperature is not evident.

WIND AND FLOW (1)

Fig. 1.3 Idealization of (a) Mean wind alone, (b) waves alone, and (c) turbulence alone. In reality waves or turbulence are often super-imposed on a mean wind. U is the component of wind in the x-direction.



Three categories of air flow can exist in the UBL:

- **Mean wind** (horizontal: 2-10 m/s, vertical: 0.01-0.1 m/s) is mainly responsible for the horizontal transport, or *advection*, of pollutants, heat and momentum. It tends to zero near the ground because of the friction with surface and obstacles.
- **Waves** can be generated by mean-wind shears, flow over obstacles and can propagate also for long distances in the case of thunderstorms or explosions. Typically transport little heat, humidity and scalars during nighttime.

WIND AND FLOW (2)

- **Turbulence**, consists of several irregular superimposed swirls of motion with different size (1 mm - 3000 m) called *eddies.* The relative strengths of these scale eddies define the *turbulence spectrum*.

Turbulence can be generated by the friction with the ground:

- <u>Solar heating</u>: during sunny days causes thermals and fover the ascending motion of warm/cold air
- <u>Frictional drag</u>: at the contact with the surface
- <u>Obstacles</u>: like trees or buildings can perturb the flow and generate turbulence near the obstacles themselves

TURBULENCE

- Mechanical turbulence (wind-related)
- Thermal turbulence (temperature-related)

- Eddy – gusts, swirls of wind in the vertical plane caused by turbulence
- Carry heat, momentum, water vapor, CO_2 etc
- As large as the boundary layer, as small as a few molecules.

MECHANICAL AND THERMAL TURBULENCE

MECHANICAL TURBULENCE

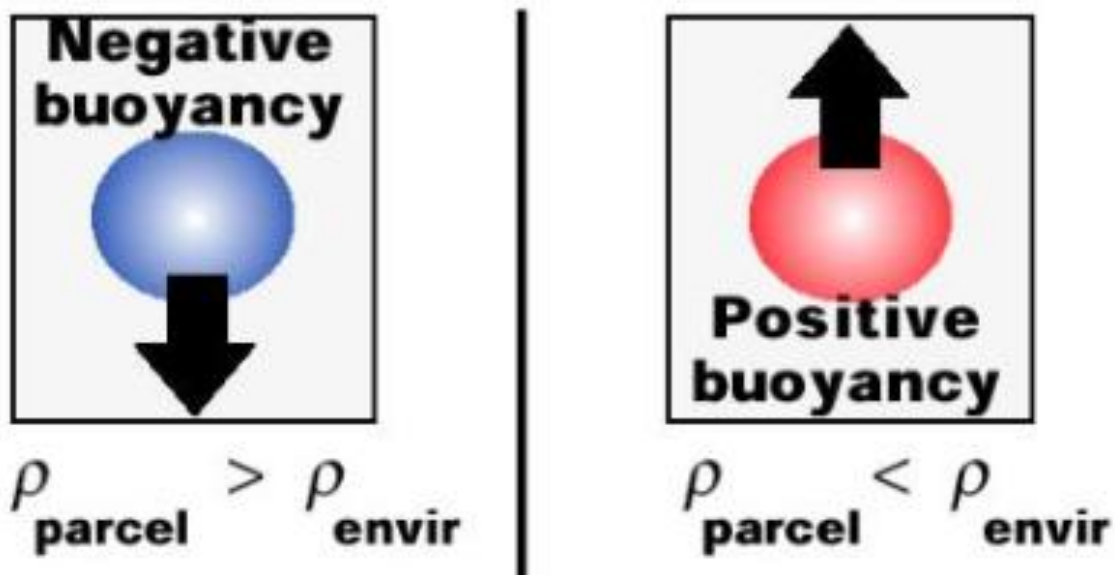
- Shear stress
- Directional shear
- Surface frictional effects

THERMAL TURBULENCE

- Density depends on temperature (ideal gas law)

$PV=nRT \Rightarrow P=\rho RT \Rightarrow \rho=P/RT$

BUOYANCY IN FLUIDS



Parcel colder than environment

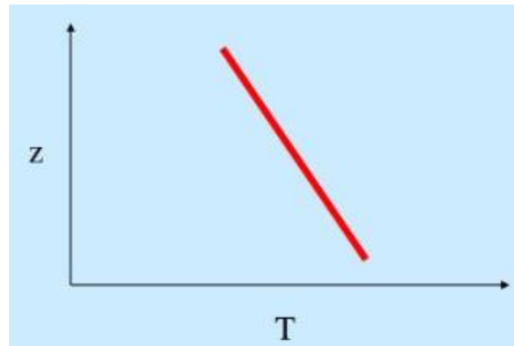
Parcel warmer than environment

ADIABATIC AND ENVIRONMENTAL LAPSE RATE

Adiabatic lapse rate:

Rate of temperature change that an air parcel experiences as it changes elevation without any heat exchange.

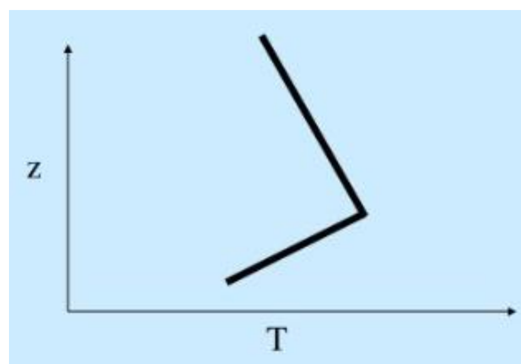
$$\left(\frac{dT}{dz}\right)_{adiab} = \Gamma = \frac{-g}{c_P}$$



Environmental lapse rate:

Actual rate of temperature change with height of the current atmosphere.

$$\left(\frac{dT}{dz}\right)_{env}$$



ATMOSPHERIC STABILITY (1)

1. **<u>Unstable conditions</u>**: usually daytime, caused by swirling eddied rising off the heated surface because they are more buoyant than surrounding air.

$$\left(\frac{dT}{dz}\right)_{env} < \Gamma$$

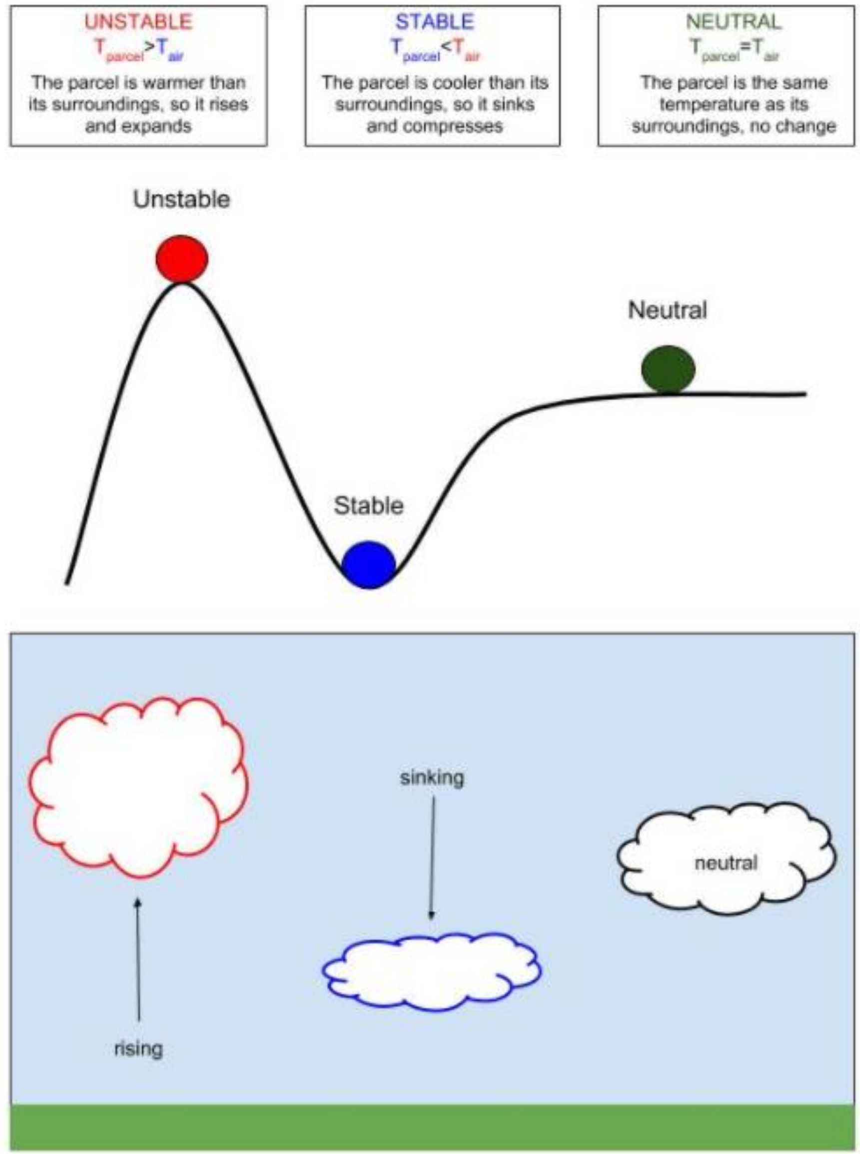
2. **<u>Stable conditions</u>**: usually at night, only mean wind and waves, little turbulence causes only horizontal transport. Stable air provides excellent conditions for high pollution levels.

$$\left(\frac{dT}{dz}\right)_{env} > \Gamma$$

3. **<u>Neutral conditions</u>**: upper PBL at night, turbulence at equal intensity in all directions.

$$\left(\frac{dT}{dz}\right)_{env} = \Gamma$$

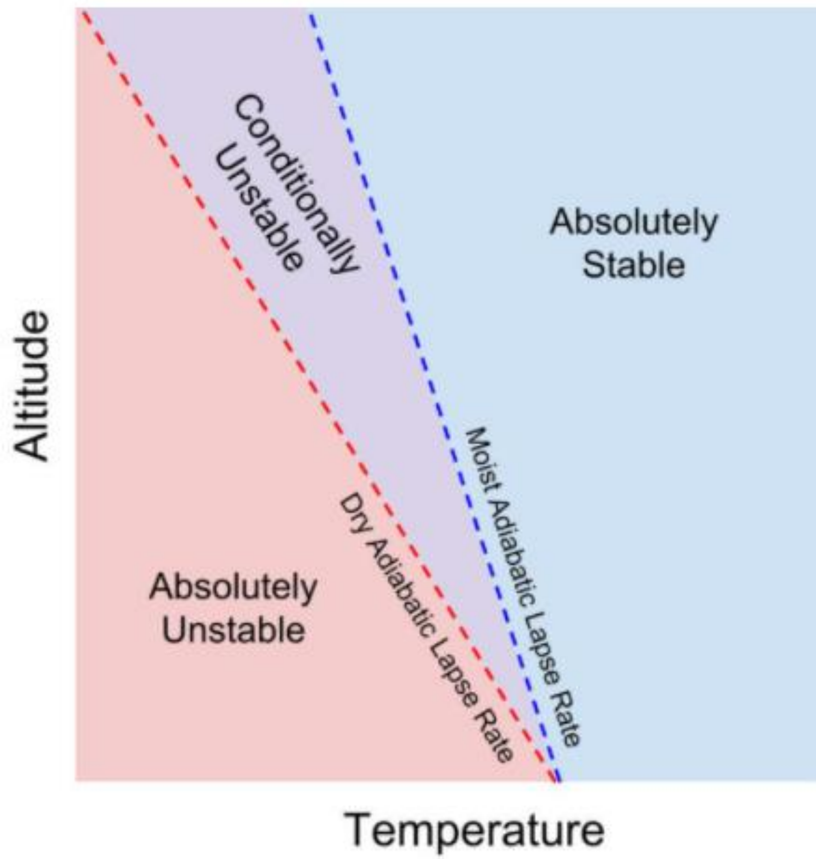
ATMOSPHERIC STABILITY (2)



Examples of stability and instability in relation to air and parcel temperatures (created by Britt Seifert).

ATMOSPHERIC STABILITY (3)

The effects of moisture change the lapse rate of the air parcel and, therefore, affects stability.

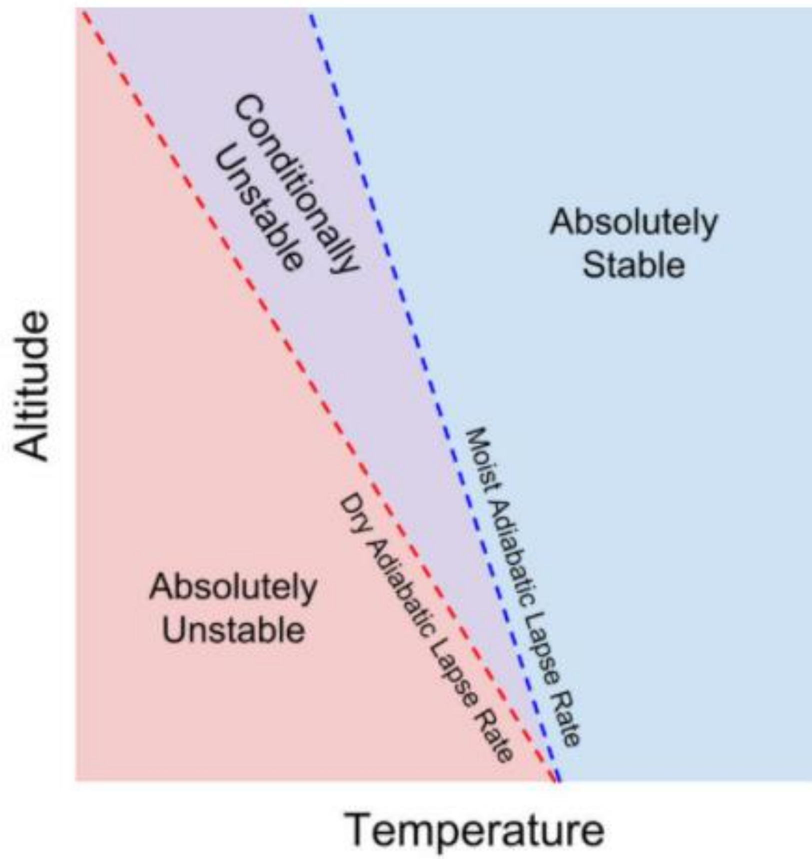


A thermodynamic diagram showing the stability of the atmosphere based on the dry ($\Gamma_d = 9.8$ K km^{-1}) and moist ($\Gamma_m = 4.5$ K km^{-1}) adiabatic lapse rates (Created by Britt Seifert).

Absolutely stable: the environmental lapse rate is less than the moist adiabatic lapse rate.

This means that a rising air parcel will always cool at a faster rate than the environment, even after it reaches saturation. If an air parcel is cooler at all levels, then it will not be able to rise, even after it becomes saturated (when latent heating will counteract some cooling).

ATMOSPHERIC STABILITY (4)

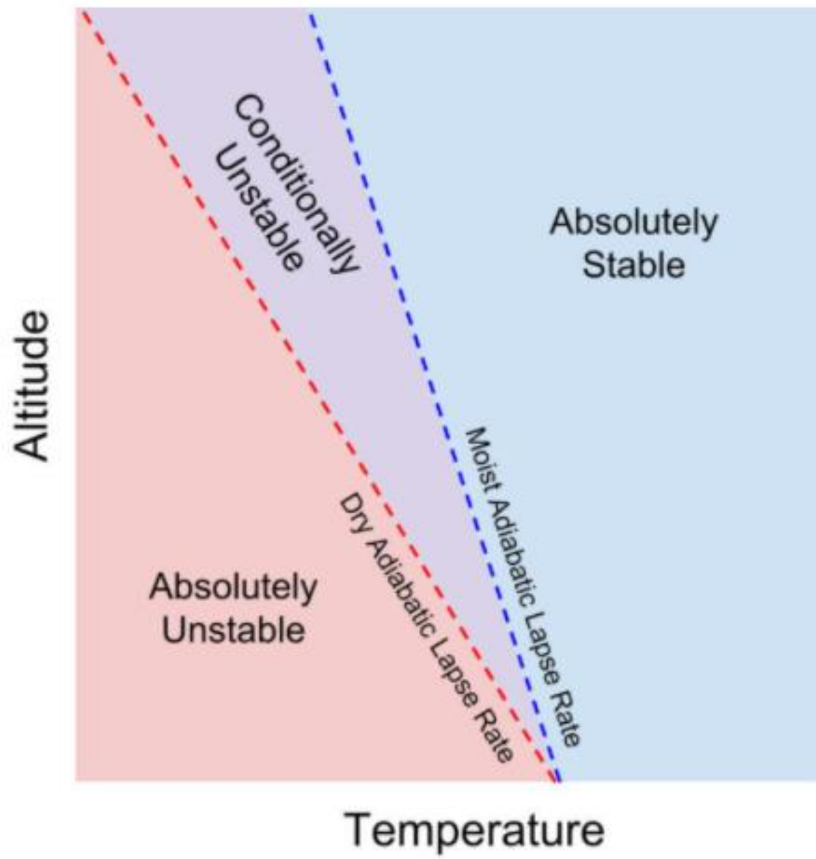


A thermodynamic diagram showing the stability of the atmosphere based on the dry ($\Gamma_d = 9.8 \text{ K km}^{-1}$) and moist ($\Gamma_m = 4.5 \text{ K km}^{-1}$) adiabatic lapse rates (Created by Britt Seifert).

Absolutely unstable: the environmental lapse rate is greater than the dry adiabatic lapse rate.

This means that a rising air parcel will always cool at a slower rate than the environment, even when it is unsaturated. This means that it will be warmer (and less dense) than the environment, and allowed to rise.

ATMOSPHERIC STABILITY (5)



A thermodynamic diagram showing the stability of the atmosphere based on the dry ($\Gamma_d = 9.8$ K km^{-1}) and moist ($\Gamma_m = 4.5$ K km^{-1}) adiabatic lapse rates (Created by Britt Seifert).

Conditionally unstable: the environmental lapse rate is between the moist and dry adiabatic lapse rates.

This means that the buoyancy of an air parcel depends on whether or not it is saturated.

In a conditionally unstable atmosphere, an air parcel will resist vertical motion when it is unsaturated, because it will cool faster than the environment at the dry adiabatic lapse rate. If it is forced to rise and is able to become saturated, however, it will cool at the moist adiabatic lapse rate.

REYNOLDS NUMBER

Turbulent flows occur when the inertial forces acting on the fluid are much greater than the stabilizing viscous forces.

The ***Reynolds number*** is a measure of the ratio of these two forces.

It is defined as:

$$\boxed{Re = \frac{Ud}{\nu}}$$

where

U is the mean velocity of the flow

d is the length scale of the flow (e.g., the diameter of the pipe through which the fluid is flowing)

ν is the kinematic viscosity of the fluid (for air ν=5x10^{-6} m^2/s)

Turbulent flows are characterized by Reynolds numbers much greater than 1000.

REYNOLDS DECOMPOSITION

Assumption: the flow is *steady*, in the sense that time averages converge to specific values when the averaging time is long enough.

We can write the instantaneous velocity, u(t), using the so-called *Reynolds decomposition* as follows:

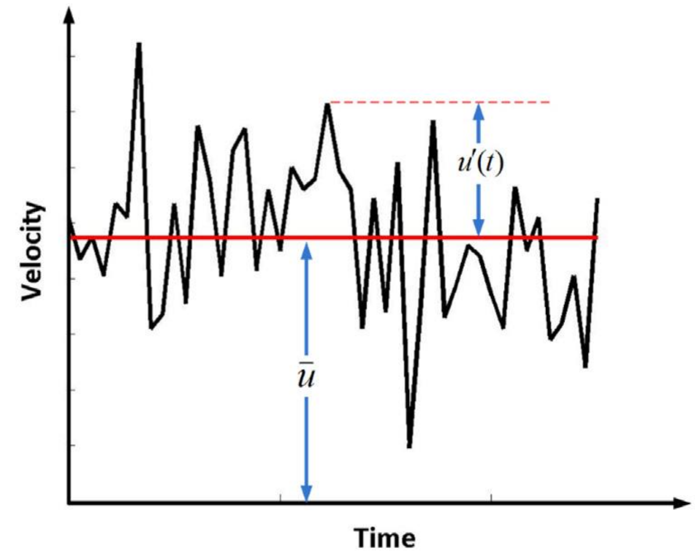
$$\boxed{u(t) = U + u'(t)}$$

where U is the time average or mean defined by

$$U = \lim_{T\to\infty} \frac{1}{T}\int_0^T u(t)dt$$

and $u'(t)$ is the fluctuating component.

Moreover, we can define u and v as the horizontal components of the velocity are denoted by and w as the vertical wind component.



STANDARD DEVIATION OF FLUCTUATING VELOCITY

The statistic of greatest relevance to dispersion of pollution is the **standard deviation of the fluctuating velocity** defined by

$$\boxed{\sigma_U = [\lim_{T\to\infty} \frac{1}{T} \int_0^T u^2 dt]^{0.5}}$$

Notice that, by definition, the time averages of the fluctuating quantities are zero.
However, the average of the product of two fluctuating quantities is not zero.

TAYLOR'S HYPOTHESIS

We would like to be able to take snapshots of the eddies in three dimensions and measure all their sizes each instant.
We can only measure the fluctuations of a variable such as wind speed, specific humidity, or temperature with a sensor at one location for a period of time.
In this way, we watch the eddies drift by the sensor. But the eddies could be changing size and shape as they drift by the sensor.

In 1938, Taylor suggested that in some cases the turbulent eddies (that we can think of as large air parcels) can be considered *frozen:* the measurements of wind speed can be used to translate the turbulence as a function of time to the correspondent function of space.

The turbulence is not really frozen, it is only an approximation of the reality.

TAYLOR'S HYPOTHESIS

EDDIES PERIOD

$$M^2 = U^2 + V^2$$

$$\boxed{\mathbf{P} = \lambda / M}$$

M= total wind speed
U = eastward Cartesian wind component
V = northward Cartesian wind component
λ = eddy diameter
P = period to pass by a stationary sensor

TAYLOR'S HYPOTHESIS

TEMPERATURE APPROXIMATION

The temperature variation with time at any location (such as where a sensor might be placed) is the sum of the total derivative and the temperature advection:

$$\frac{\partial T}{\partial t} = \frac{DT}{dT} - \vec{U} \cdot \vec{\nabla} T$$

The total derivative is the change in temperature of an air parcel moving past the sensor.
The temperature advection is the change in temperature at the sensor due to the advection of warmer or colder air past the sensor.

If we consider the Taylor's hypothesis, the turbulence is frozen and, so, the change in temperature within each eddy is negligible.

$$\frac{DT}{dT} = 0$$

and so

$$\frac{\partial T}{\partial t} = -\vec{U} \cdot \vec{\nabla} T$$

TAYLOR'S HYPOTHESIS

TEMPERATURE APPROXIMATION

$$\frac{\partial T}{\partial t} = -\vec{U} \cdot \vec{\nabla} T$$

Local temperature gradients, which might be present from one side of an eddy to another, are advected across the sensor by the mean wind without the eddy changing.

- ➢ When is this condition valid?
- ✓ Experiments suggest that this hypothesis is valid when the variation of the wind speed due to turbulence is less than ½ of the mean wind speed.

CONVECTIVE VELOCITY SCALE (1)

Heat flux from the ground to the atmosphere creates buoyancy forces, which in turn generate turbulent velocities.
We can estimate these *turbulent velocities* by considering an air parcel of unit mass with a temperature excess of ΔT over its surroundings that it acquires at the heated ground.

Considering the specific volume (volume per unit mass) of the parcel as v, the two forces on the parcel (neglecting drag forces for the moment) are:

- Downward gravitational force = g
- Upward buoyancy force = $vg\rho_s$;

where ρ_s is the density of the surroundings.

CONVECTIVE VELOCITY SCALE (2)

The net upward force is:

$$F_u = vg\rho_s - g = g(v\rho_s - 1)$$

$$v = \frac{1}{\rho}$$

$$F_u = g\left(\frac{\rho_s}{\rho} - 1\right) = g\left(\frac{T}{T_s} - 1\right) \quad \text{(from the gas law)}$$

$$F_u = g\frac{\Delta T}{T_s} \cong g\frac{\Delta\theta}{\theta_s}$$

This force, acting over a distance z, generates a kinetic energy ≈w^2 so that

$$g\frac{\Delta\theta z}{\theta_s} \approx \mathrm{w}^2$$

CONVECTIVE VELOCITY SCALE (3)

Now let us multiply both sides of the equation by w

$$g\frac{(\Delta\theta w)z}{\theta_s} \approx \mathrm{w}^3$$

The term inside the parentheses is the velocity of the parcel multiplied by the temperature excess carried by the parcel.

This quantity is proportional to the surface heat flux:

$$\Delta\theta w \sim \frac{H}{\rho c_p}$$

Then

$$\mathrm{w} \approx \left(\frac{g}{\theta_s}\frac{H}{\rho c_p}\mathrm{z}\right)^{1/3}$$

$\frac{H}{\rho c_p}$ is referred to as the <u>kinematic heat flux</u>, and is denoted by

$$\frac{H}{\rho c_p} = Q_0$$

CONVECTIVE VELOCITY SCALE (4)

Now, define a free convection scale, u_f :

$$u_f = (\frac{g}{T_0} Q_0 z)^{1/3}$$

where T_0 is the near surface temperature which is approximately equal to θ_s in a well-mixed convective boundary layer.

Another velocity scale that is used to characterize a boundary layer dominated by surface heating is the *convective velocity scale* given by:

$$\boxed{w = (\frac{g}{T_0} Q_0 z_i)^{1/3}}$$

where z_i is the PBL height, which is also called the *mixed layer* because vertical motion induced by buoyancy leads to vigorous vertical mixing of the properties of the boundary layer.

FRICTION VELOCITY (1)

Except very close to the ground, the horizontal shear stress is supported by macroscopic turbulent motion.
When parcels of air travel vertically, they exchange momentum between layers of air with different velocities.
Vertical gradients in horizontal mean velocity lead to changes in instantaneous horizontal velocities during this transfer of momentum.

If we denote the horizontal velocity fluctuation, u', created by a parcel of air with vertical velocity, w, the horizontal momentum transferred across a horizontal layer by the parcel is $\rho u'w'$, where ρ is the air density.
If the horizontal shear stress is roughly constant with height and is equal to the surface stress, τ_0, then

$$\boxed{\tau_0 = -\rho\overline{u'w'}}$$

where the overbar denotes a time average.
The negative sign ensures that τ_0 is positive because a positive w is associated with a negative u when the mean horizontal velocity increases with height.

FRICTION VELOCITY (2)

These arguments suggest that the turbulent velocities associated with shear production of turbulence scale with the surface friction velocity, u, defined by

$$\boxed{u_* = \sqrt{\frac{\tau_0}{\rho}}}$$

<u>Buoyant and shear production of turbulence operate together to determine the structure of the boundary layer</u>.

A length scale, referred to as the <u>*Monin-Obukhov length*</u>, allows us to combine the effects of these mechanisms into a single framework that describes the vertical structure of the near surface atmospheric boundary layer.

MONIN-OBUKHOV LENGTH

The absolute value of the *Monin-Obukhov length*, **L**, is roughly the height at which the turbulent velocity generated by shear is equal to that produced by buoyancy:

$$u_* \sim u_f(z = L) = \left(\frac{g}{T_0} Q_0 L\right)^{1/3}$$

which yields the definition

$$\boxed{L = -\frac{T_0 u_*^3}{g k Q_0}}$$

where k is the von Karman constant 0.4.

The negative sign indicates that when Q_0 is positive during the day, L is negative and positive when the heat flux is toward the ground.

So L is positive when the boundary layer is stable, and negative when it is unstable.

Shear production of turbulence dominates that by buoyancy at heights below the Monin-Obukhov length, while buoyant production becomes dominant above it.

MONIN-OBUKHOV SIMILARITY THEORY (1)

The **Monin-Obukhov similarity theory** (**MOST**) permits to investigate the mean and the turbulent structure of the boundary layer at heights below the order of magnitude of the Monin-Obukhov length.

The mean temperature and velocity gradients can be represented by universal functions if the velocity, temperature, and height are scaled appropriately.
The velocity scale is u_*, the height scale is L.

The temperature scale, θ_* is given by

$$\boxed{\theta_* = -\frac{Q_0}{u_*}}$$

MONIN-OBUKHOV SIMILARITY THEORY (2)

Let us consider a neutral boundary layer, that is dominated by shear. In such a boundary layer, the mean velocity gradient is of the same order as the velocity gradient across the dominant turbulent eddy at that height.
We assume that the dominant eddy at a height z has a length scale of order z and a velocity scale of order u_*.

Then, measurements indicate that we can write

$$\frac{dU}{dz} = \frac{u_*}{kz}$$

where k=0.4 is the von Karman constant.
Integration yields the logarithmic expression for the mean wind speed at height z:

$$\boxed{U(z) = \frac{u_*}{k} \ln\left(\frac{z}{z_0}\right)}$$

where z_0 is the roughness length, related to the physical dimensions of the objects at the surface.

MONIN-OBUKHOV SIMILARITY THEORY (3)

Monin-Obukhov Similarity theory states that we can account for the effects of heat flux by modifying the neutral-conditions log-law as follows:

$$\boxed{\frac{dU}{dz} = \frac{u_*}{kz}\phi_m\left(\frac{z}{L}\right)}$$

and the potential temperature gradient can be expressed as

$$\boxed{\frac{d\theta}{dz} = \frac{\theta_*}{kz}\phi_h\left(\frac{z}{L}\right)}$$

The surface heat flux goes to zero, $L \rightarrow \infty$ and $z/L \rightarrow 0$.
This means that $\phi_m(0) = 1$ and $\phi_h(0) = 1$ to be consistent with the gradient in the neutral boundary layer.
θ_* goes to zero when the surface heat flux goes to zero.

MONIN-OBUKHOV SIMILARITY THEORY (4)

The last formulations are supported by observations which indicate that

$$\phi_m = (1 - 15\frac{z}{L})^{-1/4} \qquad for\ L < 0$$

$$\phi_m = 1 + 4.7\frac{z}{L} \qquad for\ L > 0$$

and

$$\phi_h = 0.74(1 - 9\frac{z}{L})^{-1/2} \qquad for\ L < 0$$

$$\phi_h = 0.74 + 4.7\frac{z}{L} \qquad for\ L > 0$$

PBL DEPTH OVER OCEANS

PBL height varies slowly in space and time.

- Little diurnal cylce in sea surface temperature because of mixing at the top of the ocean
- Large heat capacity of water: water can absorb large amounts of heat with small temperature change

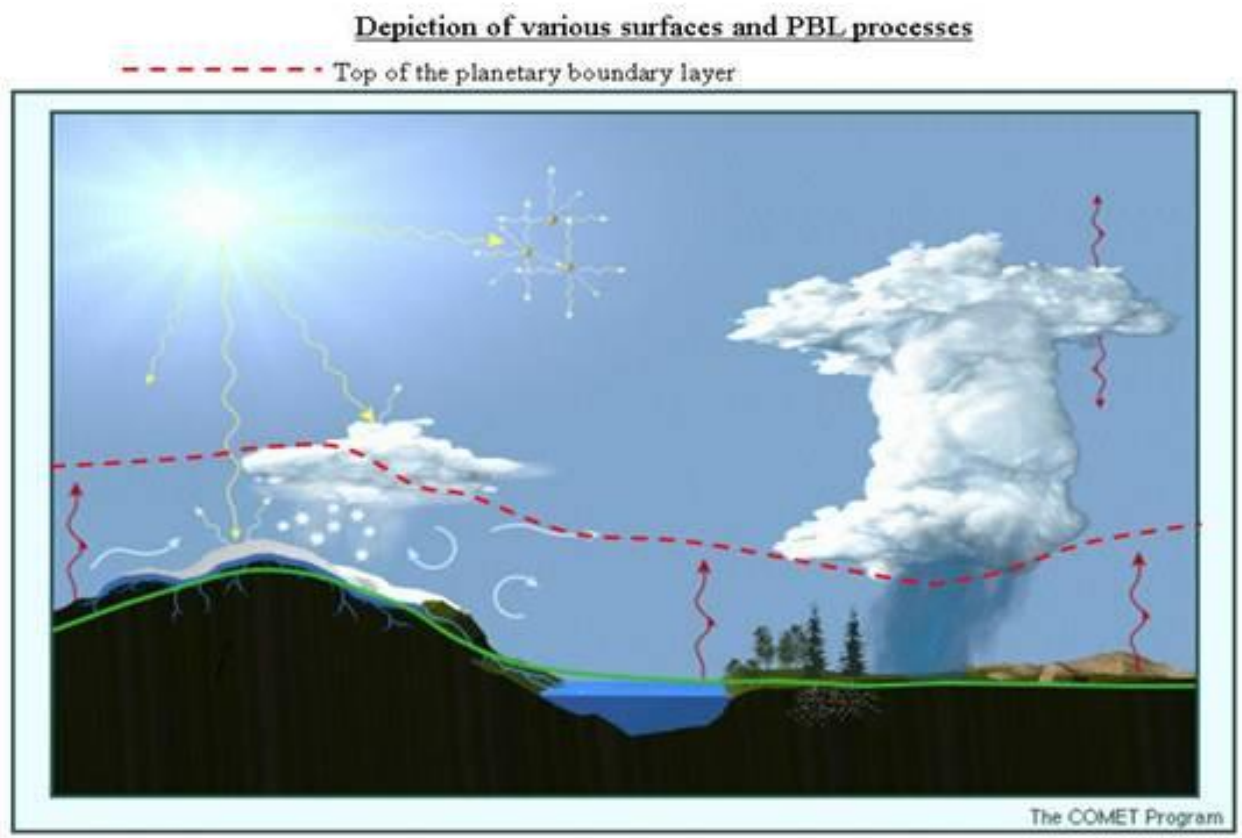
PBL height changes mainly because of <u>synoptic and mesoscale processes</u> of vertical motion and <u>advection</u> of different air masses over the sea surface.

Very gentle variation: PBH height varies of about 10% over a horizontal distance of 1000 km.
Expections: borders between ocean currents with different temperatures

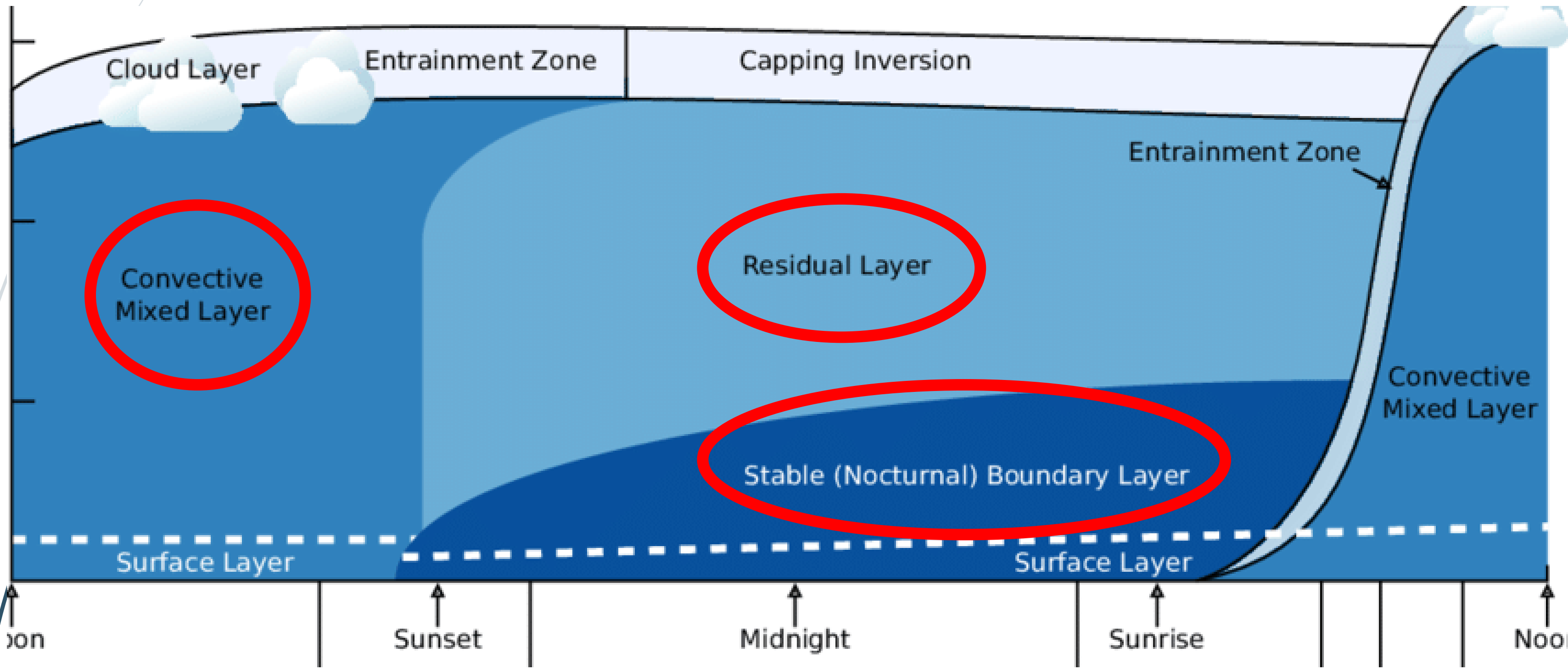
PBL DEPTH OVER LOW PRESSURE REGIONS

The upward motions carry boundary-layer air away from the ground to large altitudes throughout the troposphere.

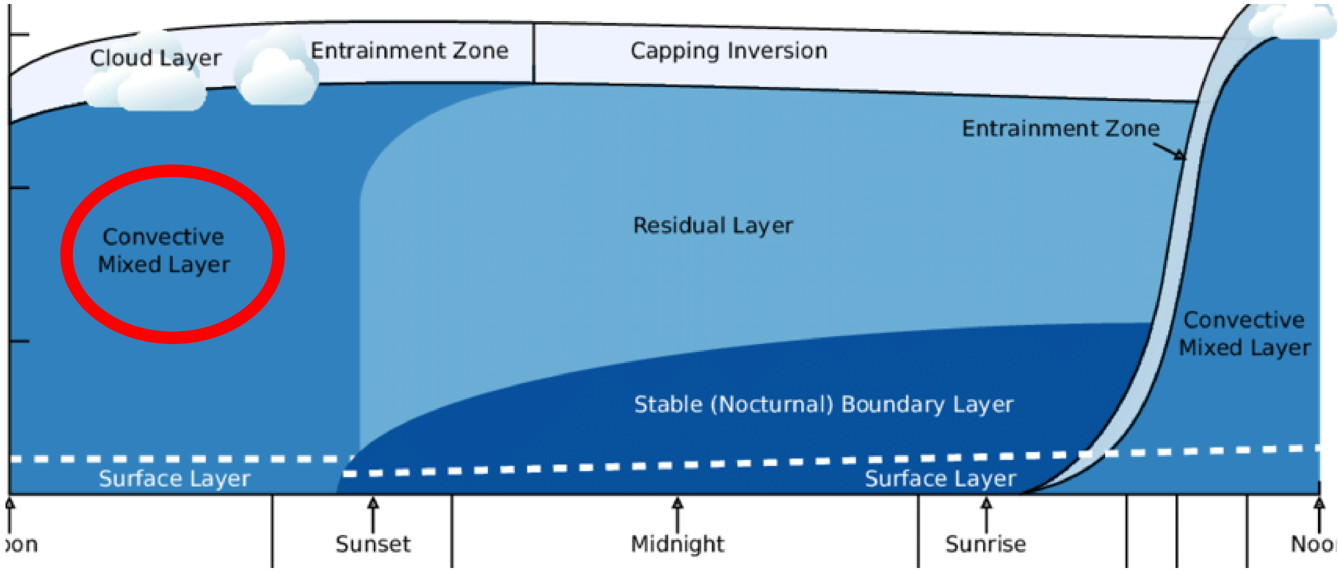
It is complex to identify a PBL top so, cloud base is typically considered as an arbitrary cut-off of the PBL.



PBL STRUCTURE OVER HIGH PRESSURE REGIONS



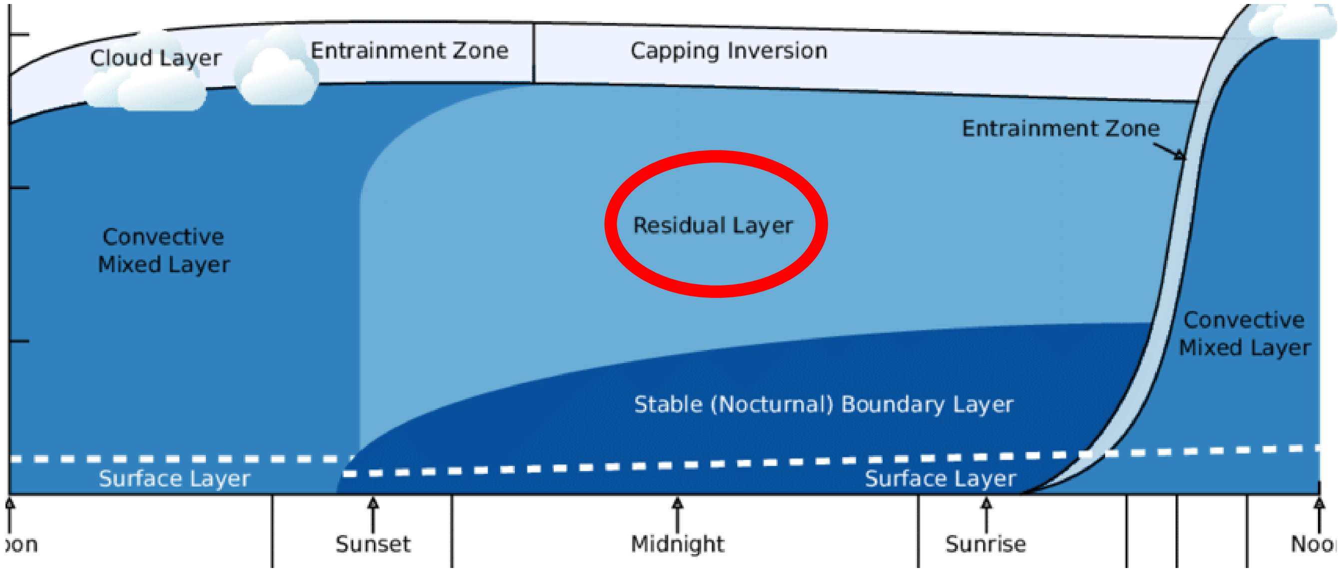
PBL STRUCTURE OVER HIGH PRESSURE REGIONS



Mixed Layer (Convective Boundary Layer):

- turbulence driven by convection (large eddies or thermals)
- heat transfer from solar heating of the ground to the atmosphere
- mixed layer grows by entrainment of air from above it
- virtual temperature nearly adiabatic in middle; superadiabatic (i.e., potential temperature decreases with height) near surface; subadiabatic (i.e., potential temperature increases with height) at top, where exchange of air between the ABL and the free troposphere occurs
- wind speeds are sub-geostrophic in mixed layer, crossing isobars because of turbulent drag

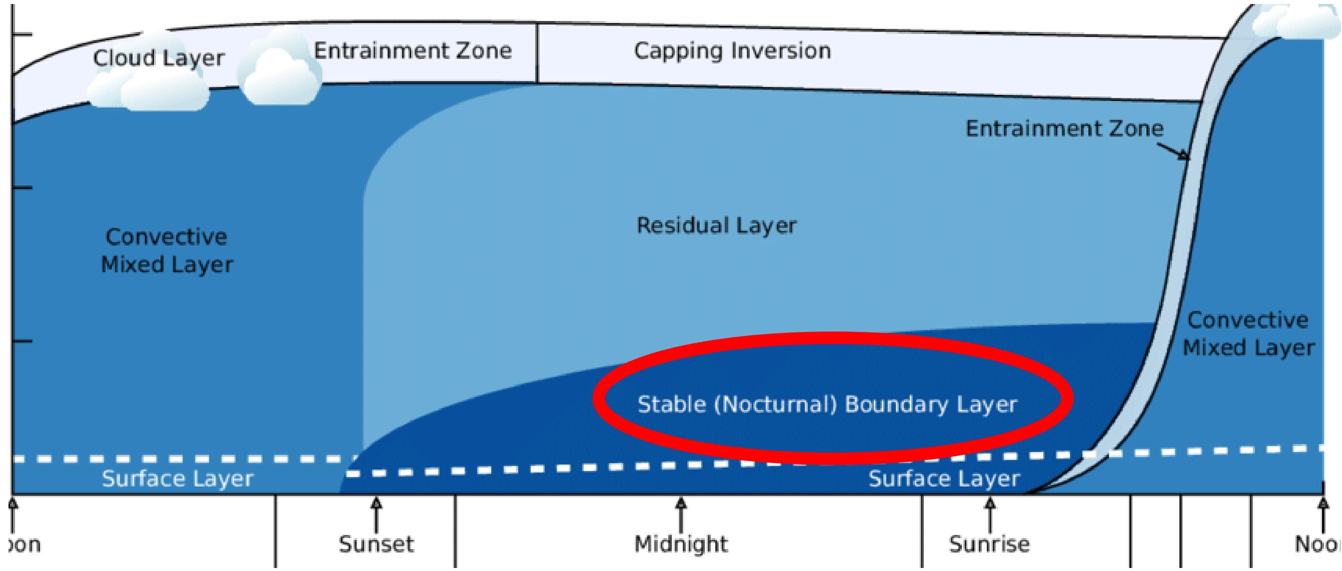
PBL STRUCTURE OVER HIGH PRESSURE REGIONS



Residual Layer

- disconnected from boundary layer and Earth's surface
- neutrally stratified, i.e. it has small but near-equal turbulence in all directions
- contains moisture and trace atmospheric constituents from the day before

PBL STRUCTURE OVER HIGH PRESSURE REGIONS



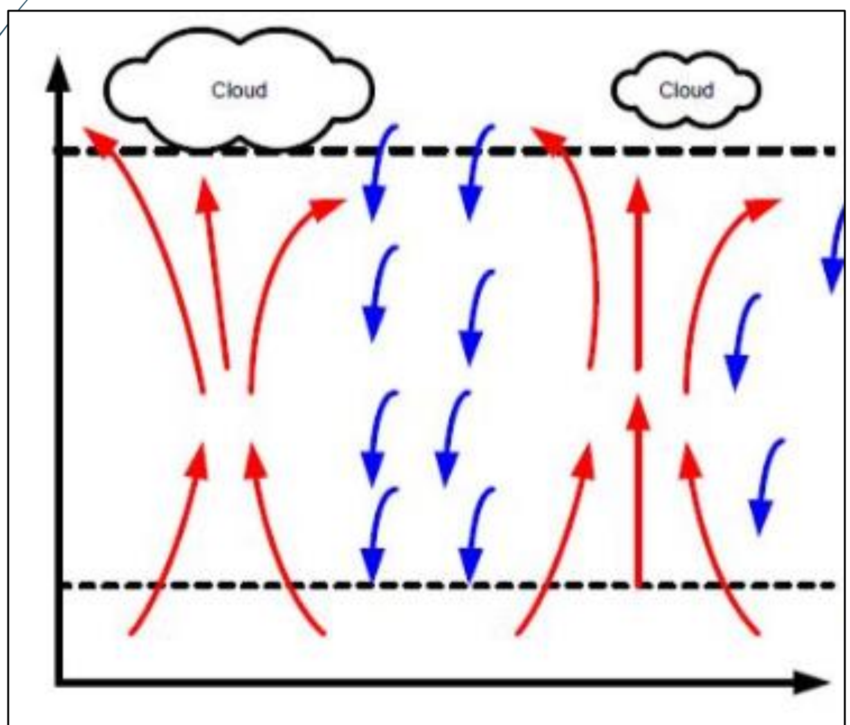
Stable Boundary Layer

- statically stable with weaker turbulence that occurs sporadically
- winds aloft may increase to supergeostrophic speeds (low-level jet or nocturnal jet)
- stability tends to suppress turbulence, except for occasional shear-generated turbulence caused by the low-level jet

PBL STRUCTURE OVER HIGH PRESSURE REGIONS

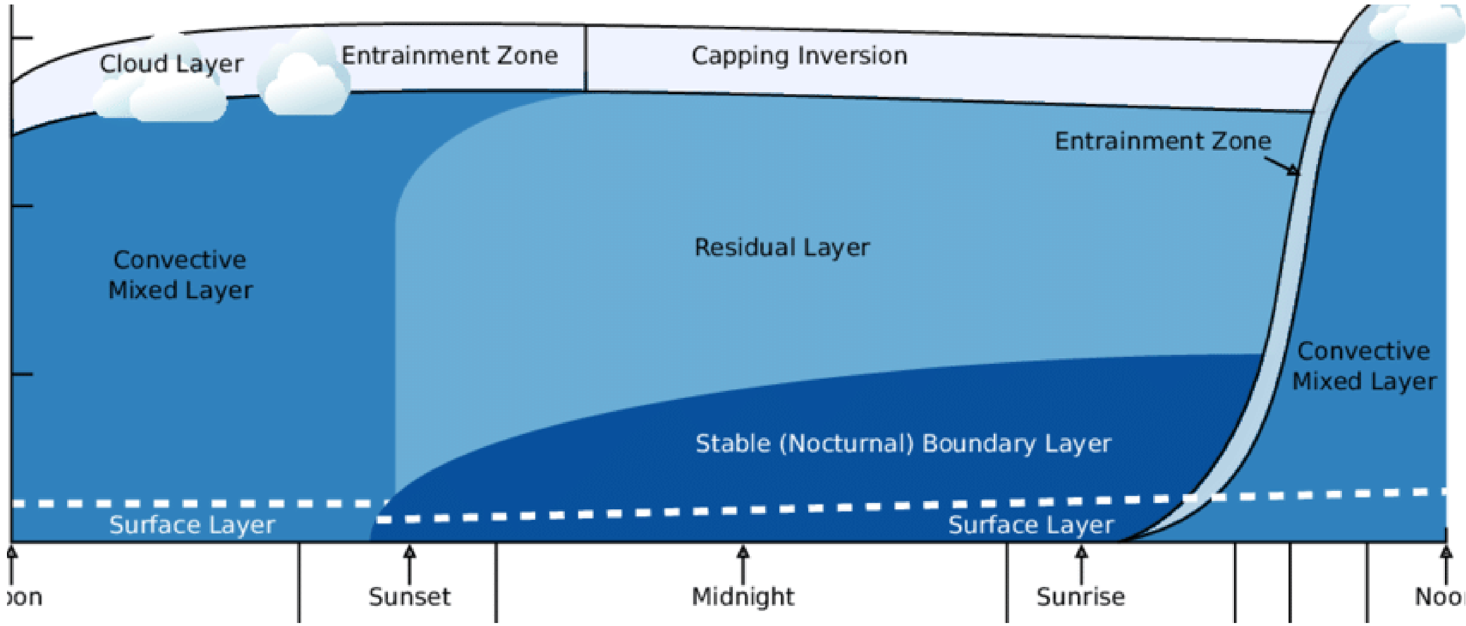
The **Surface layer** is a very shallow region close to the ground (up to 10% of PBL). It is characterized by a superadiabatic lapse rate, moisture decrease with height and strong wind shear.[2] Almost all wind shear and all the potential temperature gradient in the CBL are confined in the surface layer.

Microlayer (or **Interfacial layer)** is identifies as the few centimeters of air closest to the ground, where molecular transport dominated over turbulent transport.



The **Entrainment zone** (or **Capping inversion**) is about 40% of the depth of the PBL. It is the region of statically stable air at the top of the mixed layer, where there is entrainment of free atmosphere air downward and overshooting thermals upward. It inhibits mixing and confines pollutants.

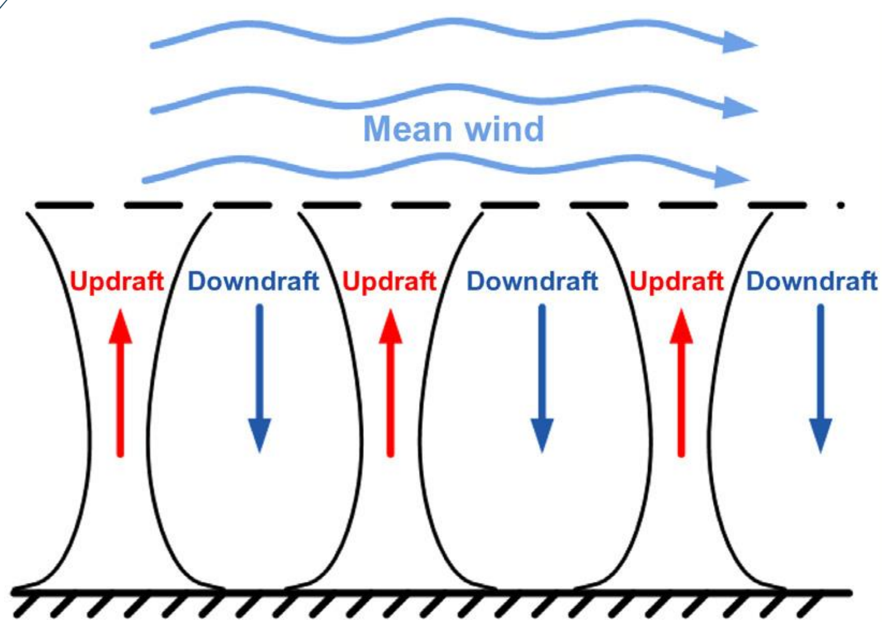
Daily evolution of PBL



1. <u>During daytime</u>: radiation coming from the sun passes through the atmosphere. The soil warms up and a mixed layer forms and grows during the morning.
2. <u>One or two hours before sunset</u>: the surface emits more energy than it receives, and convective turbulence is no longer fed by the ground.
3. <u>During nighttime</u>: the ground continues to radiate more infrared energy upward than it receives from the overlying atmosphere. The air above it gets colder and colder, generating a layer of cold (denser) air below a layer of warmer (lighter) air, giving rise to the stable stratified layer.

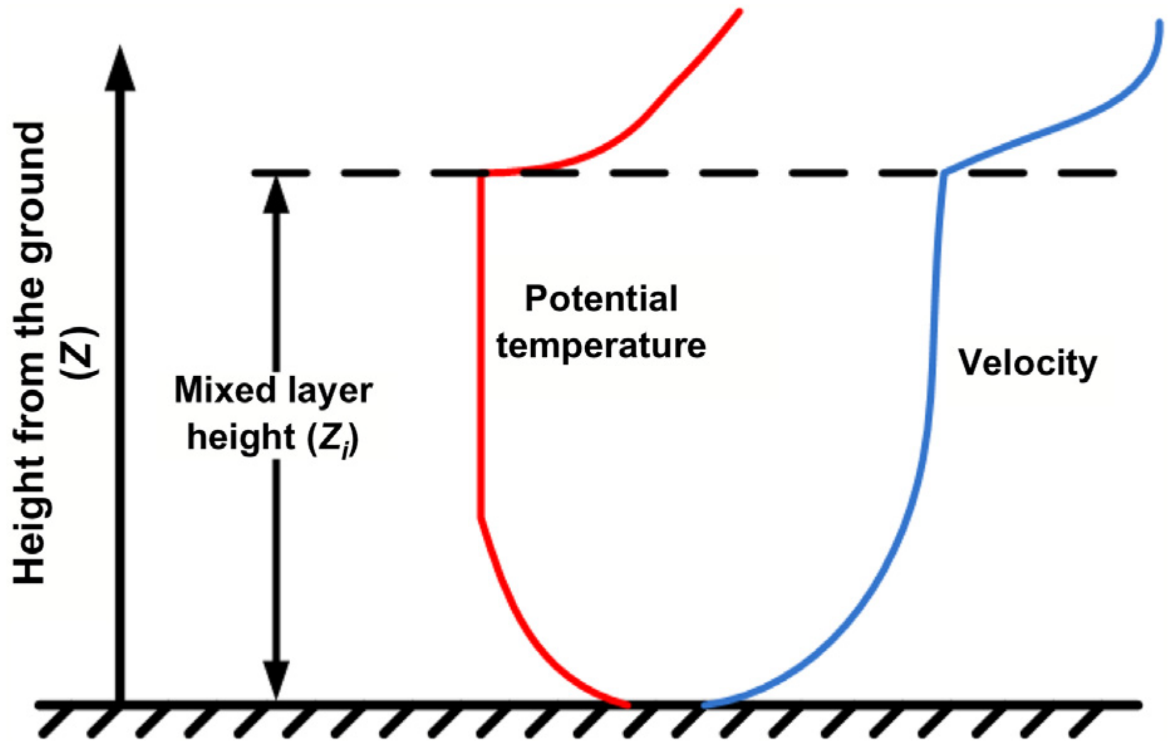
CONVECTIVE PBL (1)

Turbulence in the daytime boundary layer is maintained primarily by sensible heating at the surface, which results in parcels of air that are warmer than their surroundings.
The turbulent motion in the convective boundary layer is organized into long-lived updrafts and downdrafts that extend through the depth of the boundary layer and are carried by the mean wind.



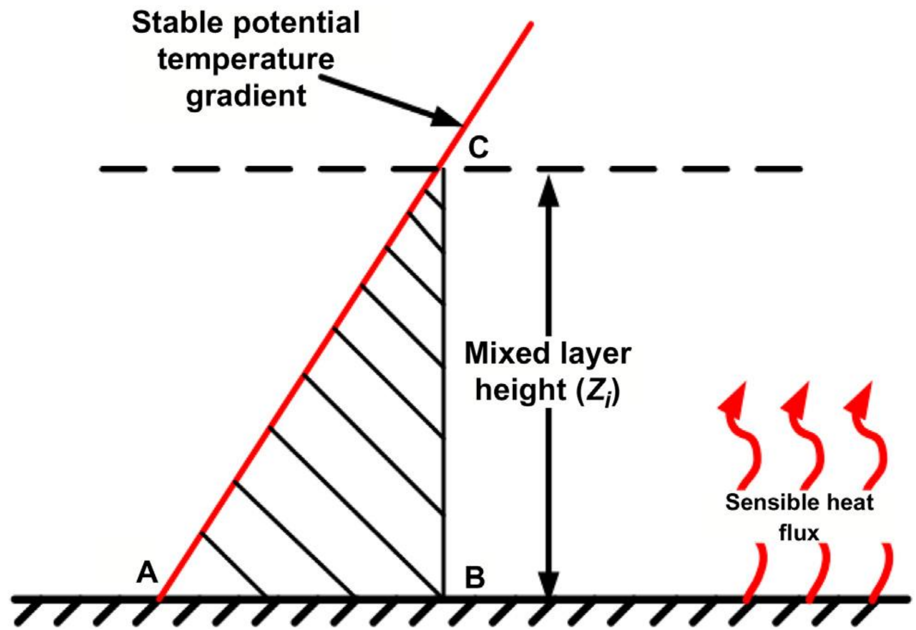
The updrafts consist of accelerating parcels, while the downdrafts are caused by compensating downward motion. Thus, the velocities in updrafts are higher than those in downdrafts; mass balance requires that the horizontal area occupied by downdrafts is higher than that of updrafts.

CONVECTIVE PBL (2)



The potential temperature is super-adiabatic close to the surface: the potential temperature decreases with height.
Above a tenth of the mixed layer height, the potential temperature is relatively uniform because of vigorous vertical mixing.
The layer above the mixed layer can be stably stratified.
The velocity profile in the daytime boundary layer is relatively flat in the mixed layer. The rapid change in velocity at the top of the boundary layer reflects the fact that the velocity is vertically mixed below the top.

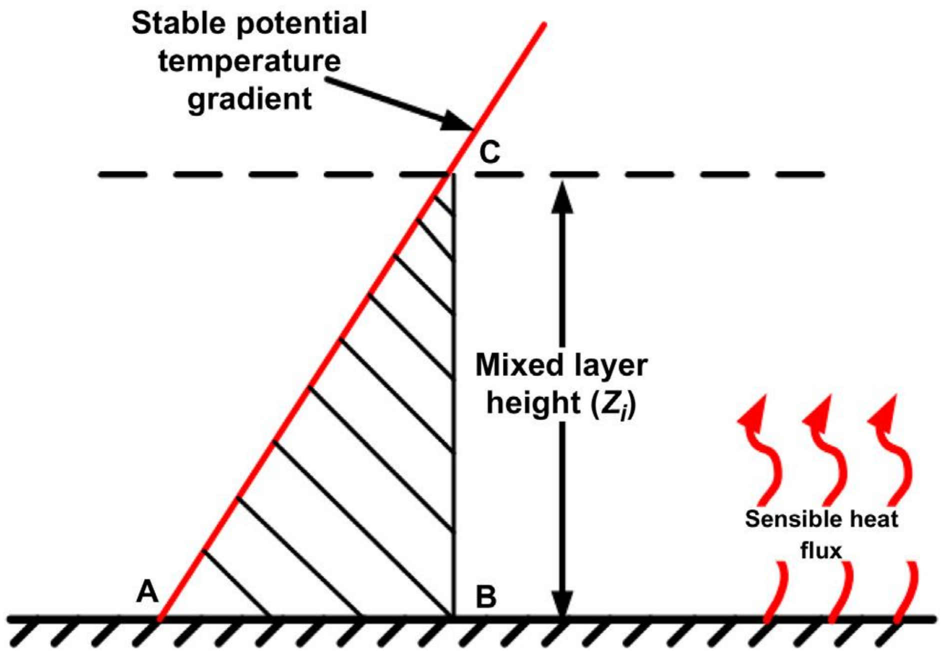
HEIGHT OF THE CONVECTIVE PBL (1)



Assumption: the sensible heat input into the atmosphere modifies the potential temperature in the mixed layer.

Consider a mixed layer that grows by eroding a layer with a stable potential temperature gradient.
AC → initial temperature profile
BC → potential temperature after sensible heating has occurred over a time, T, since sunrise
AB → temperature change at the surface
triangle ABC → modification of the energy of the PBL

HEIGHT OF THE CONVECTIVE PBL (2)



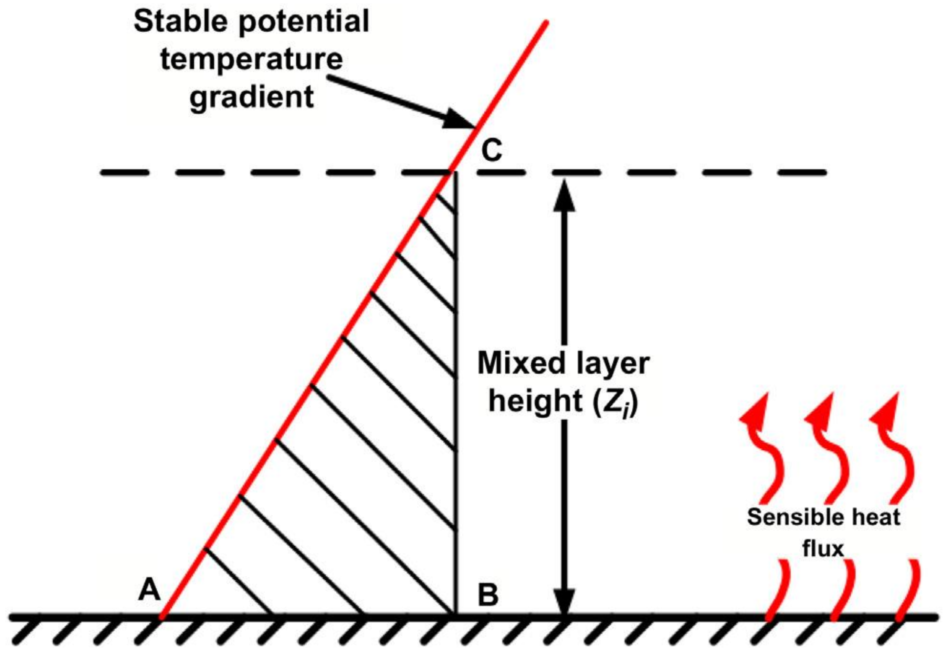
$\gamma \rightarrow$ potential temperature gradient of AC
$\Delta\theta \rightarrow$ temperature change in AB

The energy equivalent of the triangle ABC can be written as:

$$Energy\ in\ ABC = \rho c_p \frac{1}{2} \Delta\theta z_i$$

Noticing that $\Delta\theta = \gamma z_i$ we can equate this energy to the sensible heat flux integrated over T to obtain

HEIGHT OF THE CONVECTIVE PBL (3)



Noticing that $\Delta\theta = \gamma z_i$ we can equate this energy to the sensible heat flux integrated over T to obtain

$$\rho c_p \frac{1}{2} \gamma z_i^2 = \int_0^T H(t)dt$$

where $H(t)$ is the time-varying sensible heat flux.

For simplicity, if we assume that the sensible heat flux increases linearly with time, we obtain the following expression for the mixed layer height:

$$\boxed{z_i^2 = \frac{H_{max} T}{\gamma \rho c_p}}$$

where H_{max} is the maximum heat flux.

HEIGHT OF THE CONVECTIVE PBL (4)

$$\boxed{z_i^2 = \frac{H_{max}T}{\gamma \rho c_p}}$$

Example:

$$\frac{H_{max}}{\rho c_p} = 0.3\ \frac{m}{sK}$$
$$T = 6h$$
$$\gamma = 5K/1000\ m$$

We obtain $z_i \approx 1000\ m$

Note that the boundary layer height increases with time as long as the heat flux is positive.

TURBULENT VELOCITIES IN THE CONVECTIVE PBL

Different formulations for the turbulent velocities have been proposed, basing on the *free convection velocity scale*

$$\boxed{u_f = \left(\frac{g}{T_s} Q_0 z\right)^{1/3}}$$

$$\boxed{\sigma_w = 1.3 u_* \qquad for\ z \leq 0.1 z_i}$$

$$\boxed{\sigma_w = 1.3\left(u_*^3 + u_f^3\right)^{1/3} \quad for\ \ z(1.3u_*) \leq z \leq\ z(u_f)}$$

Between $0.1z_i$ and close to the top of the mixed layer, σ_w associated with buoyancy production of turbulence is proportional to the *convective velocity scale* given by

$$\boxed{w_* = \left(\frac{g}{T_s} Q_0 z_i\right)^{1/3}}$$

where z_i is the mixed layer height. Then, we find that

$$\boxed{\sigma_w = \sigma_v = \sigma_u \cong 0.6 w_*}$$

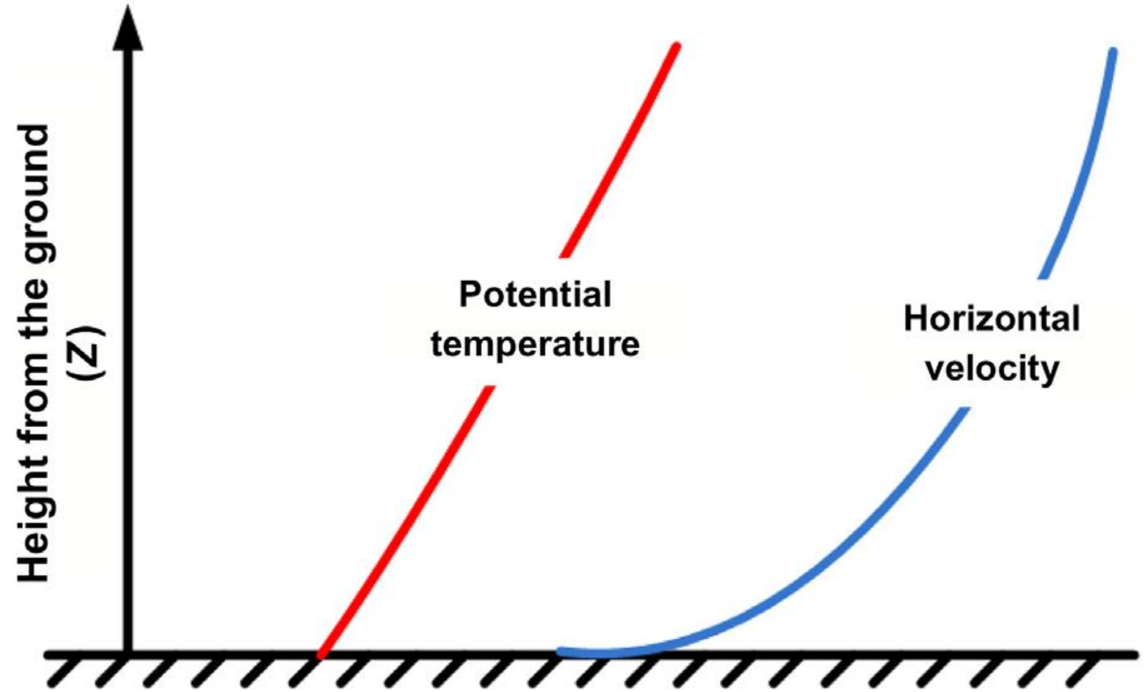
STABLE PBL (1)

When the sun sets, turbulence energy production by buoyancy comes to a stop.
Over a period of an hour, the turbulence in the mixed layer collapses, and shear becomes the primary mechanism for the production of turbulence. Because the ground is initially warmer than the atmosphere, the thermal radiation leaving the ground exceeds that being supplied by the atmosphere. This deficit leads to a cooling of the ground.

Initially, both the sensible heat flux and the ground heat flux are directed away from the earth's surface. The surface cools rapidly, and a point is reached at which the ground becomes colder than the layers above in the atmosphere. At this stage, the heat flux from the atmosphere is directed toward the earth's surface.

This process is referred to as the formation of a radiation-induced surface inversion — the temperature (and the potential temperature) increases with height.

STABLE PBL (2)



The mean wind, as obtained experimentally, can be written as

$$\boxed{U(z) = \frac{u_*}{k}\left[\ln\left(\frac{z}{z_0}\right) - \psi_m\left(\frac{z}{L}\right) + \psi_m\left(\frac{z_0}{L}\right)\right]}$$

where

$$\psi_m\left(\frac{Z}{L}\right) = -17\left[1 - \exp\left(-0.29\frac{Z}{L}\right)\right]$$

No similar equations are available for the variation of temperature through the depth of the boundary layer.

HEIGHT OF THE STABLE PBL (1)

Most expressions for the height of the stable boundary layer, which we denote by h, are based on dimensional analysis backed by relatively weak physical arguments.

One of the most famous formulations, educed by Nieuwstadt's (1981) is the following:

$$\boxed{h = u_* t \left[\frac{L/h}{a + bL/h}\right]}$$

where a and b are empirical constants.

The problem with diagnostic equations is that the height of the boundary layer reacts instantaneously to u_* and L.
This means that h will drop suddenly (and unrealistically) if the wind speed, and thus u_*decreases quickly.

HEIGHT OF THE STABLE PBL (2)

One way of getting around this problem is to allow the boundary layer to have some inertia.
This is done by using the following equation to estimate the time evolution of h:

$$\frac{dh}{dt} = \frac{h_d - h}{\tau}$$

where h_d is the estimate given by the diagnostic equation, and τ is the timescale, given by

$$\tau = \frac{\beta h}{u_*}$$

where β is an empirical constant.

When $h_d = h$, h does not change.

If h_d increases suddenly in response to an increase in wind speed, dh/dt becomes positive, so that h will grow toward h_d; the time of reaction is proportional to h/u_* . This means that if either h is large, or u_* is small, h reacts slowly to changes in hd.

TURBULENT VELOCITIES IN THE STABLE PBL

The stable potential temperature gradient suppresses the production of turbulence because it opposes vertical motion.
Under these circumstances, shear production of turbulence is matched by the destruction associated with the stable temperature gradient and viscous dissipation.
While we do know that the levels of turbulence in the stable boundary layer are low, we are not in a good position to characterize the variation of these levels as a function of height.

The parameterization that is sometimes used to estimate σ_w is that of Nieuwstadt (1984):

$$\boxed{\sigma_w^2 = 1.7u_*^2\left(1 - \frac{z}{h}\right)}$$

The horizontal turbulent velocities, σ_u and σ_v, in the stable boundary layer do not appear to be related to micrometeorological variables.
They are affected by mesoscale flows and local topography, which are difficult to characterize using models.
In the absence of measurements, a value of σ_v of 1 m/s can be used.

VERTICAL STRUCTURE OF PBL

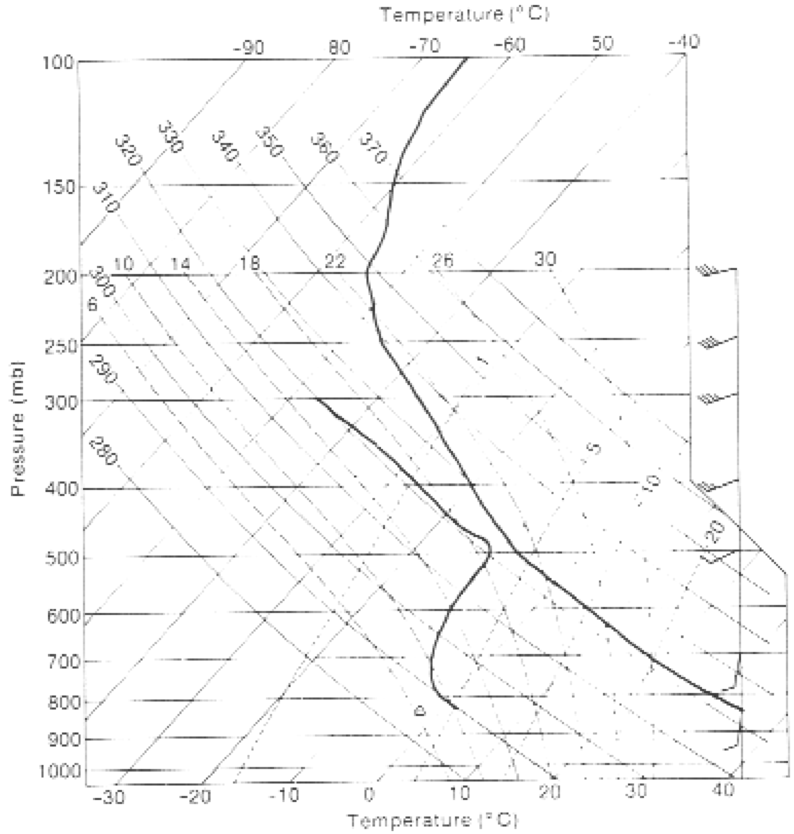


Figure 9a. A composite of five afternoon (0000 UTC) soundings by Brown et al. (1982) for convective events that produced damaging surface winds associated with high-based cumulonimbi in the Front Range area of Colorado. The temperature is represented by the curve on the right, and the dew point temperature by the curve on the left. The sounding is also typical of the type of environment found, during JAWS, to be associated with large numbers of microbursts (Caracena and Flueck, 1988). The sounding shows the characteristic deep, dry mixed layer (with dry adiabatic lapse rate, ~9.8°C km^{-1}) topped by a moist, cloud-bearing layer (low dew point depression).

An example of a deep well mixing boundary layer in the Front range area of the Rockies, shown in Skew-T diagram.

VERTICAL STRUCTURE OF PBL

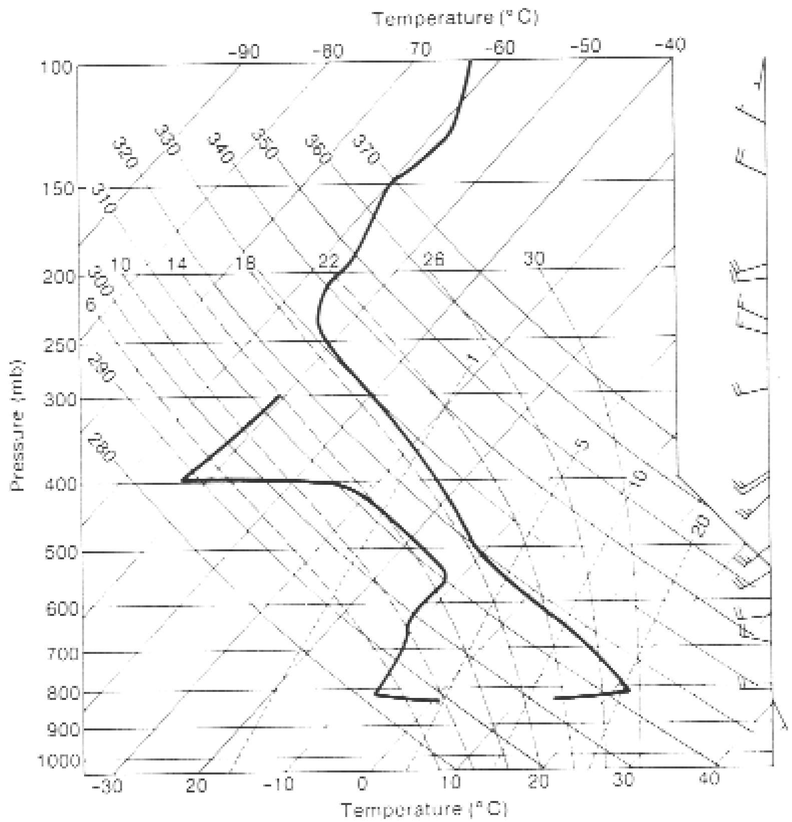


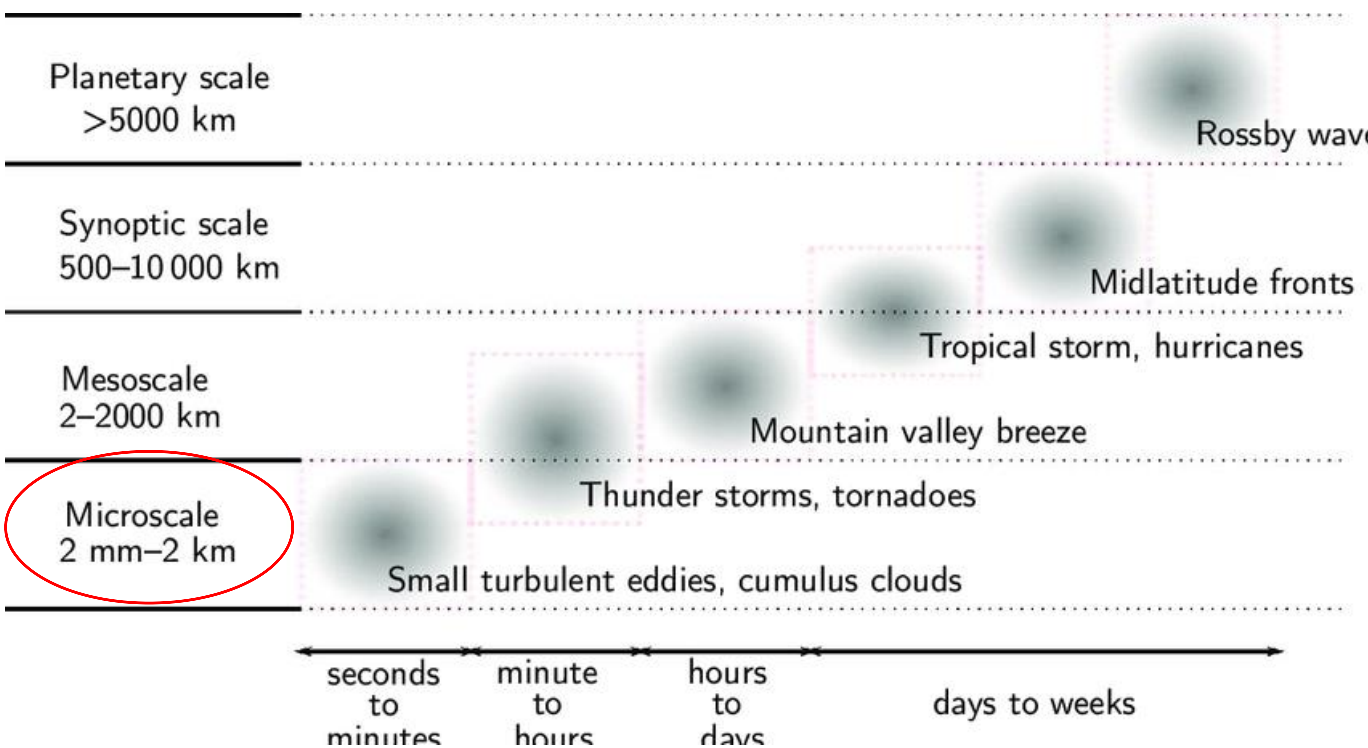
Figure 9b. A dry microburst sounding, as in Fig. 9a, but taken in the morning (1200 UTC) of 31 May 1984, showing the kind of shallow inversion near the surface that usually disappears later in the day to produce a sounding like Fig. 9a, thereby implying a high potential for dry microbursts later in the day. This sounding was taken about 7 hours before a microburst-related near-accident at Stapleton International Airport.

An example morning sounding showing the surface inversion (stable) layer that developed due to night-time surface cooling. Such a shallow stable layer can usually be quickly removed after sunrise.

MICROMETEOROLOGY (1)

Phenomena, such as turbulence, with space scales smaller than about 2 km and with time scales shorter than 1h are classified as **microscale**.

The study of small-scale phenomena is named **micrometeorology.**



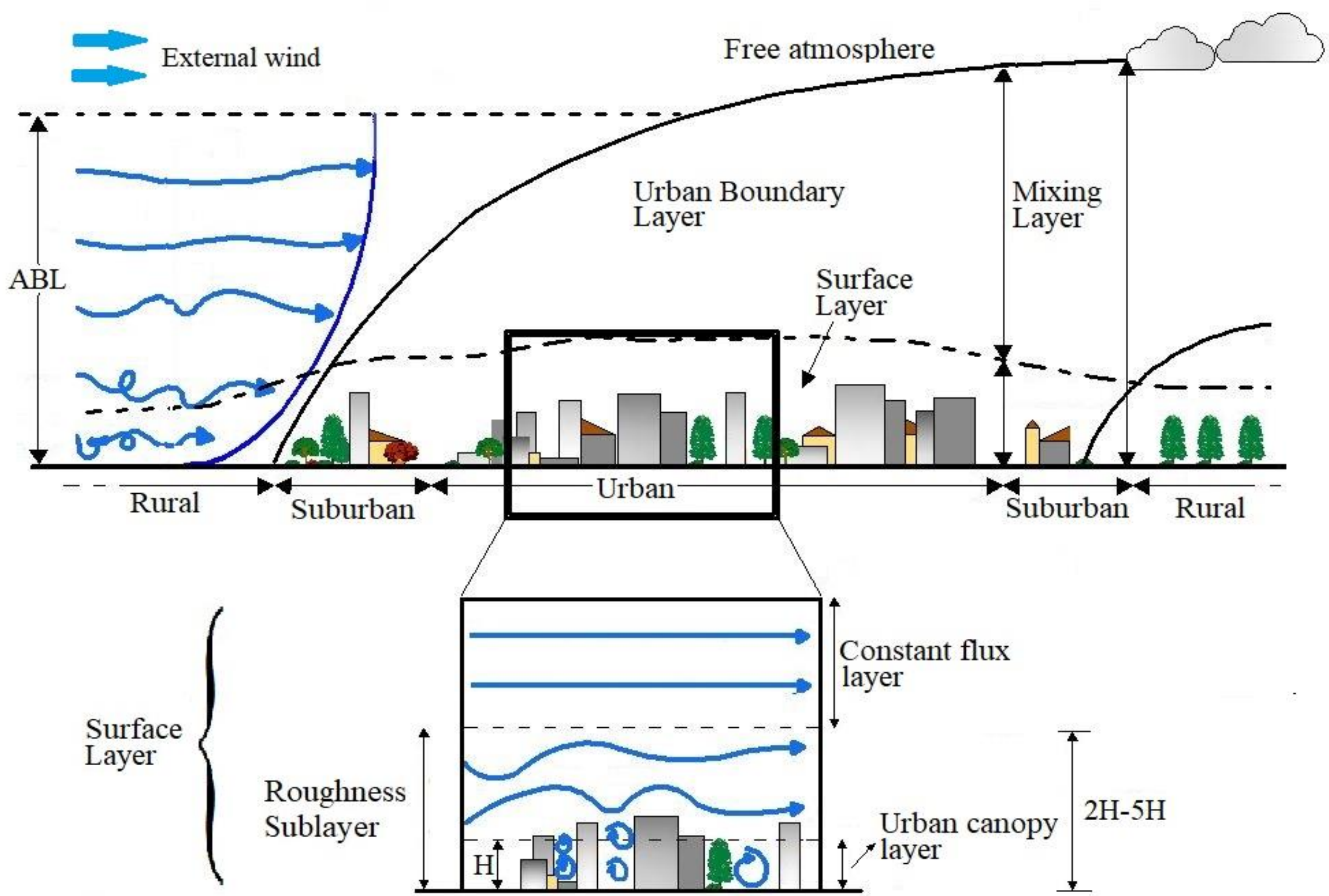
MICROMETEOROLOGY (2)

Micrometeorology is typically studies using:

- stochastic models, are a tool for estimating probability distributions of potential outcomes by allowing for random variation in one or more inputs over time. The random variation is usually based on fluctuations observed in historical data for a selected period using standard time-series techniques and deal with the average statically effects of the eddies
- similarity theory involved the common-behavior exhibited by many empirically-observed phenomena of properly scaled
- phenomenological methods permit to investigate large size structures (as thermals) in a deterministic manner

Alternative approaches regard numerical and laboratory simulations.

VERTICAL STRUCTURE OF PBL

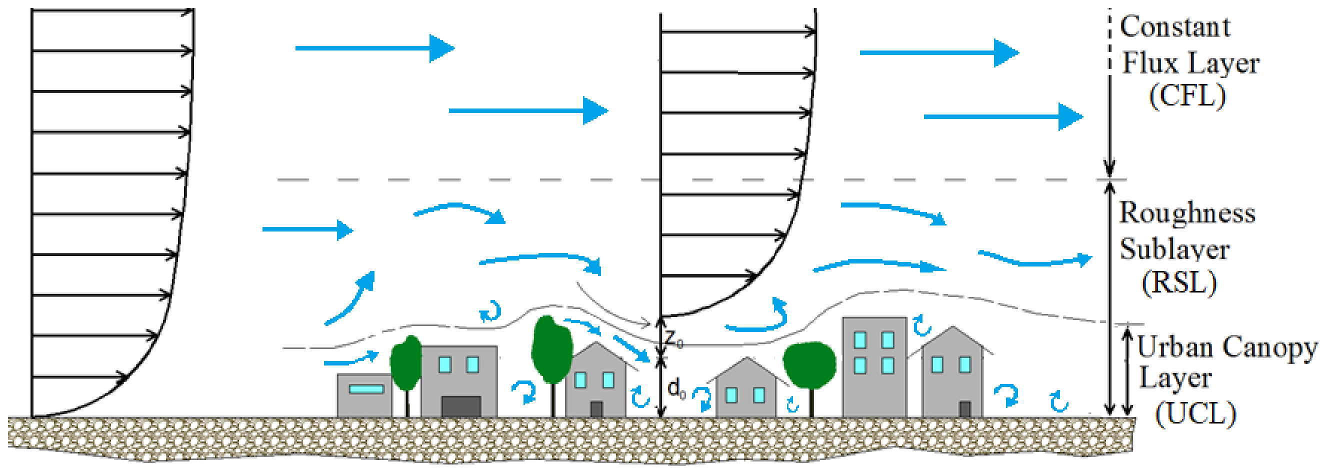


URBAN BOUNDARY LAYER (UBL)

The **urban boundary layer** (**UBL**) is the part of ABL in which most of the Earth's population live and is one of the most complex and least understood topic in environmental fluid mechanics.

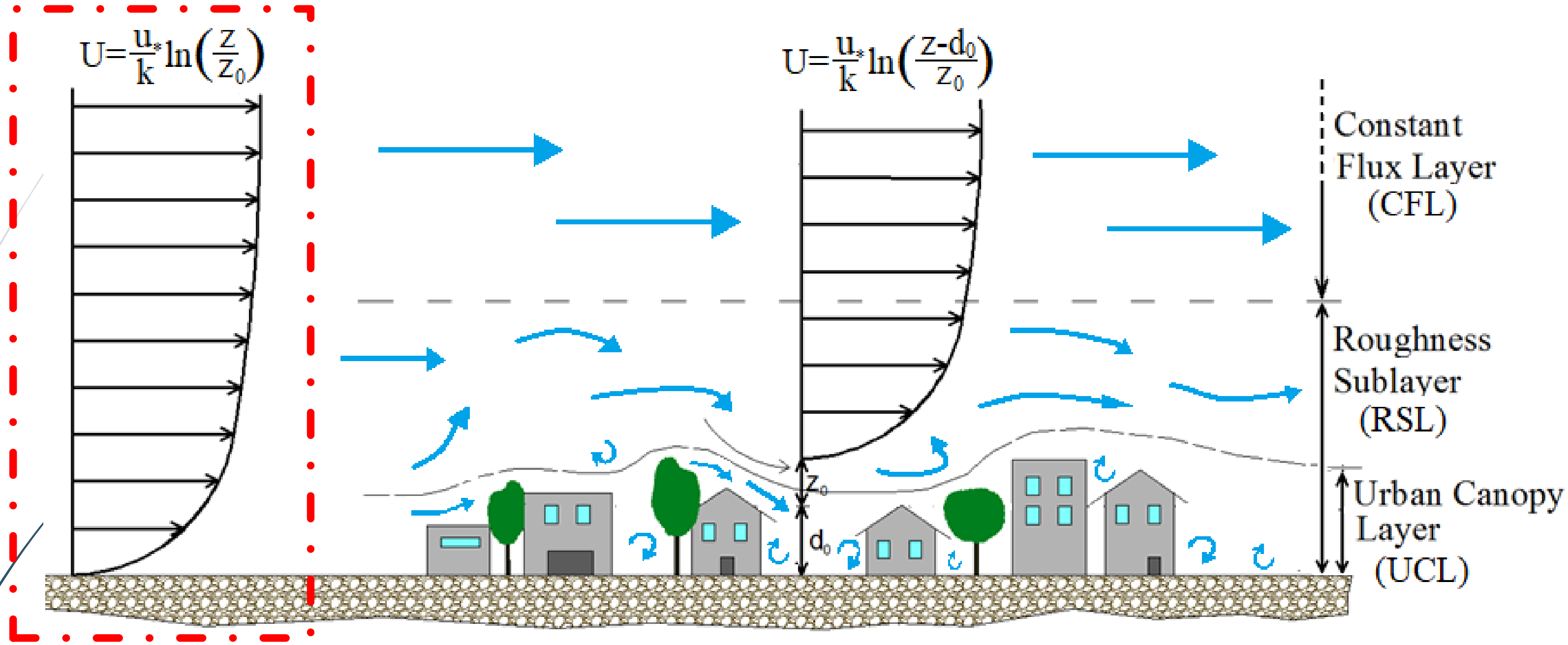
UBL characteristics are strictly linked to weather conditions, urban texture and exchanges of mass, momentum, heat and moisture between the urban canopy and the overlying air layers.

VERTICAL STRUCTURE OF UBL



- *urban canopy layer* (UCL): the layer between the average height of the buildings and the surface
- *roughness sublayer* (RSL): the layer adjacent to the surface wherein the flow is strongly influenced by individual roughness elements and, hence, shows a full three-dimensional structure
- *constant flux layer* (CFL): extends from the top of the roughness sublayer to a few hundred meters above the surface. It is characterized by a nearly constant distribution of turbulence fluxes and the logarithmic velocity profile

VERTICAL WIND PROFILE (1)



Wind profile over flat terrain

$$U(z) = \frac{u_*}{k}\ln\left(\frac{z}{z_0}\right)$$

z height
u_* friction velocity (i.e. the scaling velocity, which is related to the drag force at the surface)
k=0.41 von Karman constant
z_0 aerodynamic roughness length

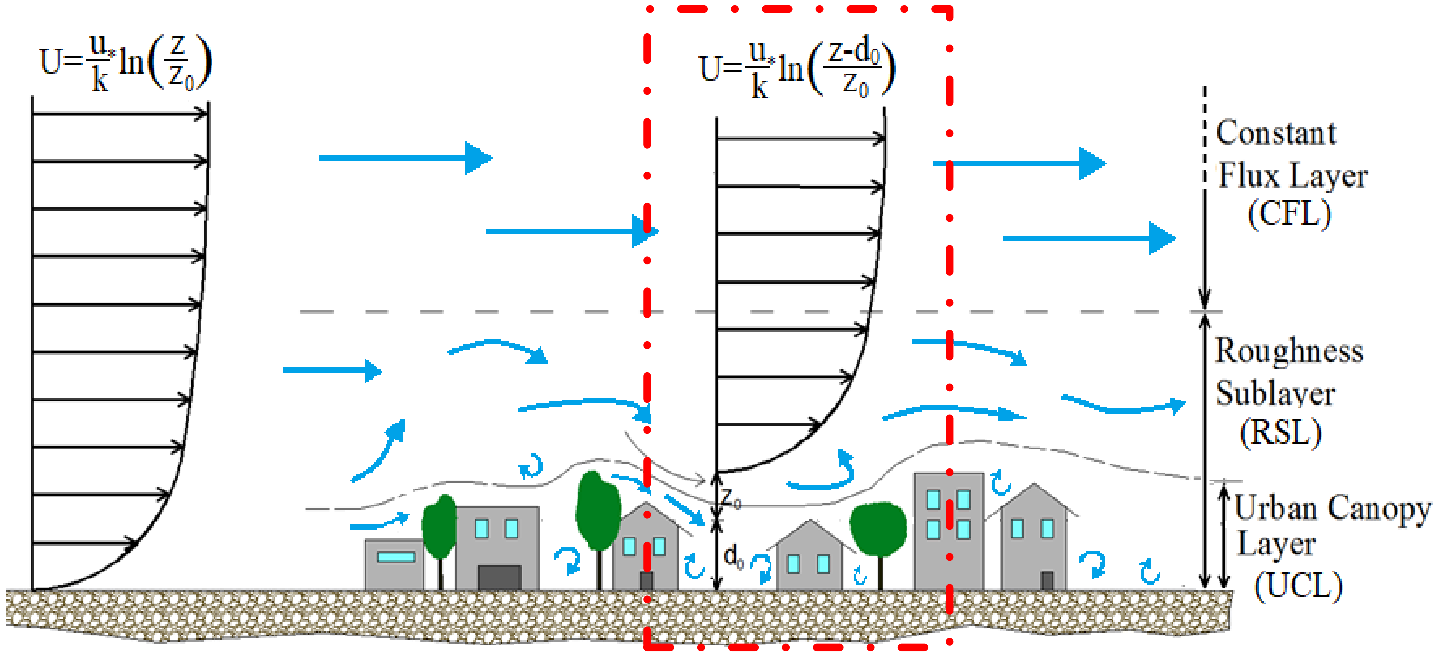
AERODYNAMIC ROUGHNESS LENGHT z_0

Class	Terrain	z_0 [m]
I	muddy terrains, wetlands, icepack	$10^{-5} \div 3\ 10^{-5}$
	water areas*	$3\ 10^{-5} \div 0.0002$
II	sand	$0.0002 \div 0.001$
III	airport runway areas, mown grass	$0.001 \div 0.01$
IV	farmland/airports with very few trees, buildings, etc.	$0.01 \div 0.04$
V	many trees and/or bushes	$0.04 \div 0.1$
VI	forecasts, suburbs	$0.1 \div 1$
VII	cities	$1 \div 4$

* air and sea form a dynamically coupled system, the determination of the roughness length of open sea and water surfaces is usually obtained by models taking into account shapes and dimension of the waves.

Classification of terrain based on roughness length, adopted by European Wind Atlas (1989)

VERTICAL WIND PROFILE (2)



<u>Wind profile over complex terrain</u>

$$U(z) = \frac{u_*}{k}\ln\left(\frac{z-d_0}{z_0}\right)$$

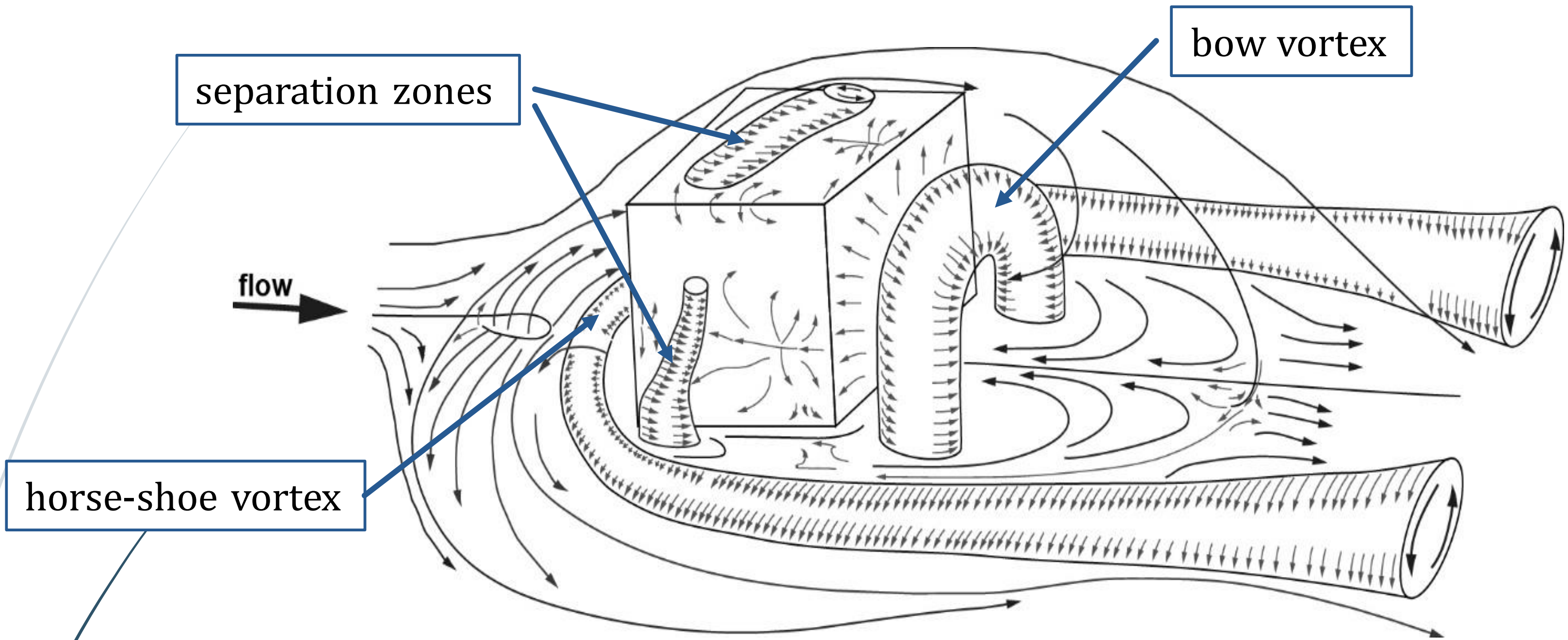
d_0 zero displacement height, i.e. the effective height of the ground due to the vertical flow displacement through the *canopy*

ZERO DISPLACEMENT HEIGHT d_0

Terrain	**d_0 [m]**
water	not available
sand	not available
grass	0.07÷0.66
crops	< 3
orchards	< 4
deciduous forests	< 20
conifer forests	< 30

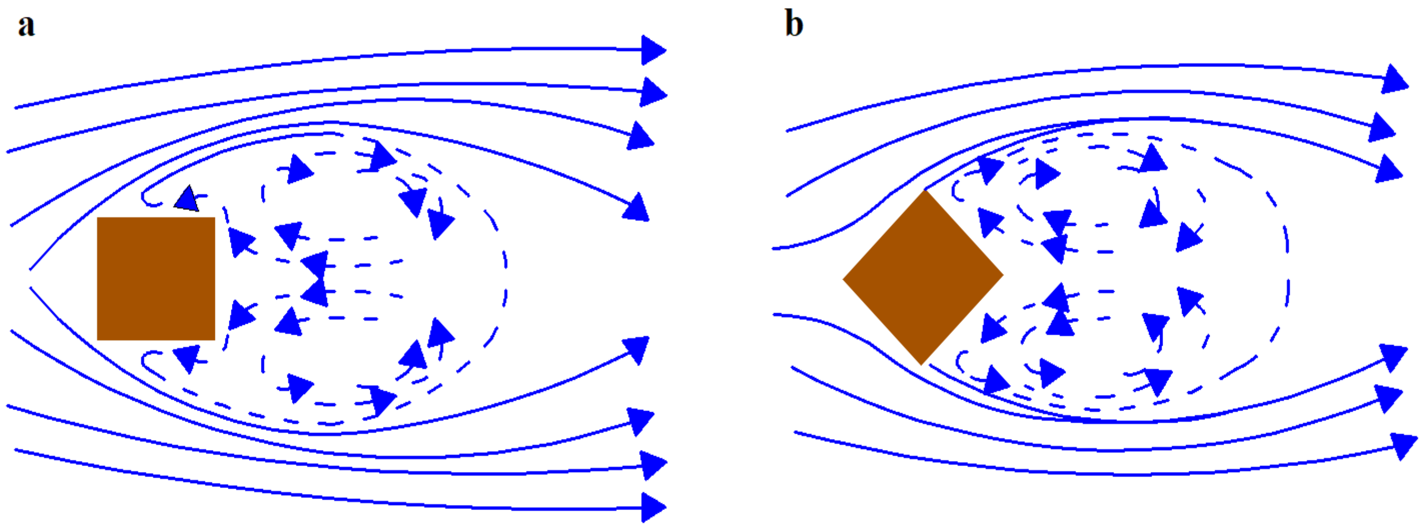
Zero displacement height for different surfaces (from Monteith and Unsworth, 1990)

FLOW AROUND AN ISOLATED BUILDING (1)



The shear in the approaching flow causes a downward flow over the lower portion of the upwind façade of the building and a "*horse-shoe vortex*" at the upwind base, which wraps around the building near the ground extending also further downstream.
The flow separates at the upwind edges, producing *separation zones on the roof top and on the lateral sides of the building*.
At the downwind edges the flow separates again, producing a cavity region and the associated *bow vortex*. The flow there interacts with the current merging from the roof and the side walls as well as with the horseshoe vortex.

FLOW AROUND AN ISOLATED BUILDING (2)

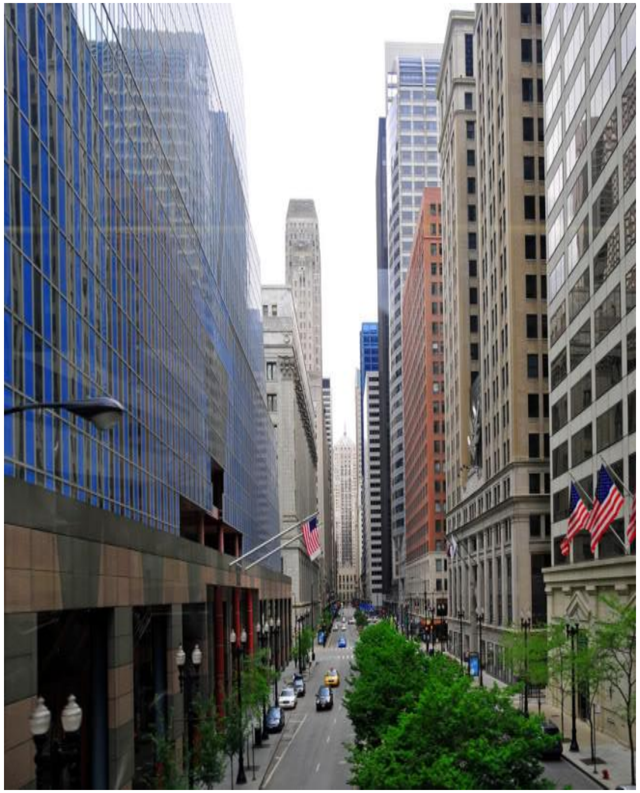


Sketch of the flow patterns around an isolated cubic building along a horizontal plane in the case of approaching wind **a** perpendicular to a façade and **b** forming an angle of 45° with the façade.

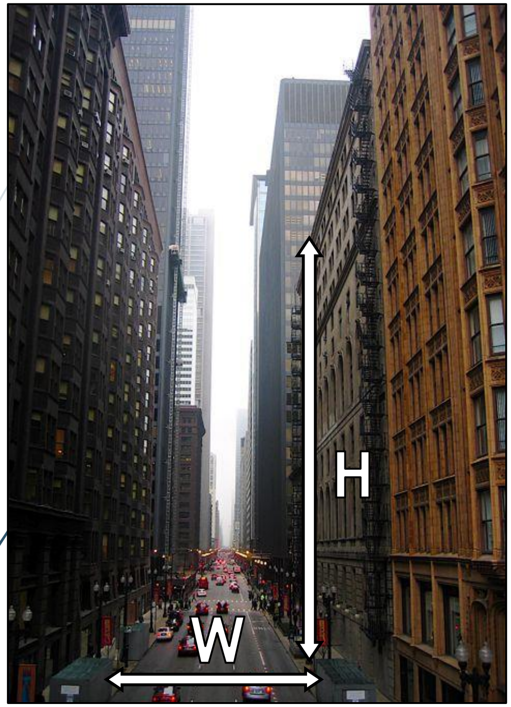
URBAN STREET CANYON

It is the basic geometric unit of urban areas and can be defined as a quasi-narrow street between buildings that line up continuously along both sides.

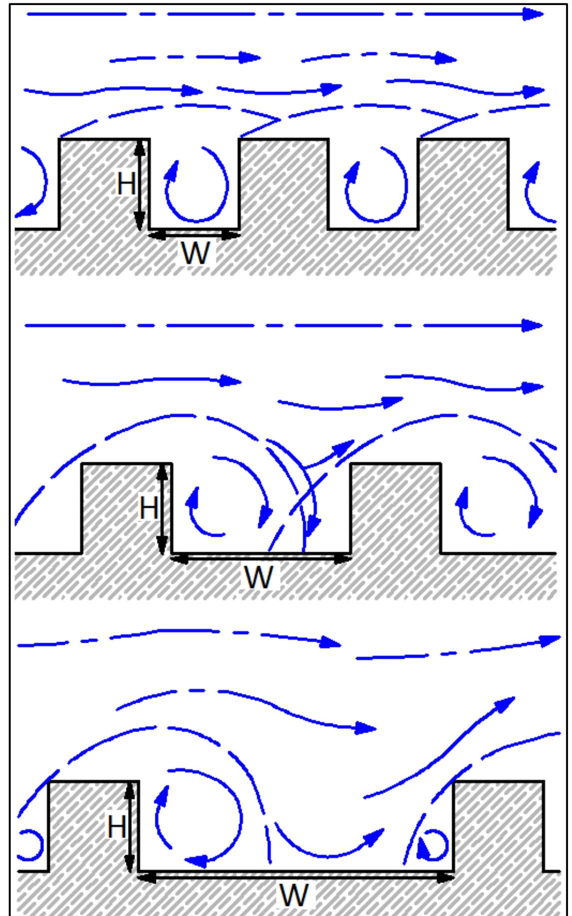
It is assumed as an archetype for more complex and realistic urban fabrics because of its simple and versatile representation.



FLOW REGIME IN 2D STREET CANYONS



ASPECT RATIO:
AR=W/H

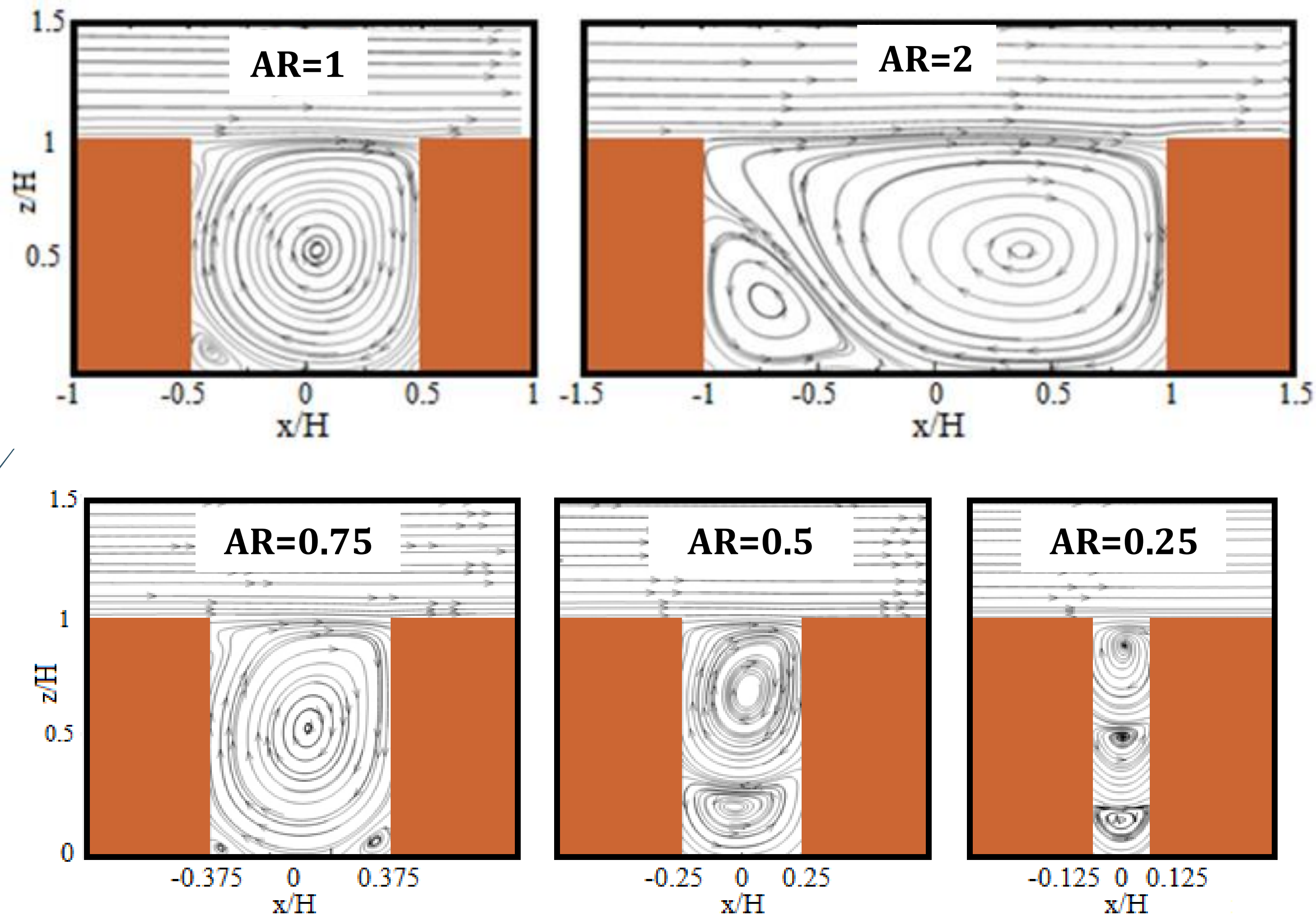


SKIMMING FLOW
AR < 1.5

WAKE-INTERFERENCE REGIME
1.5 ≤ AR ≤ 2.5

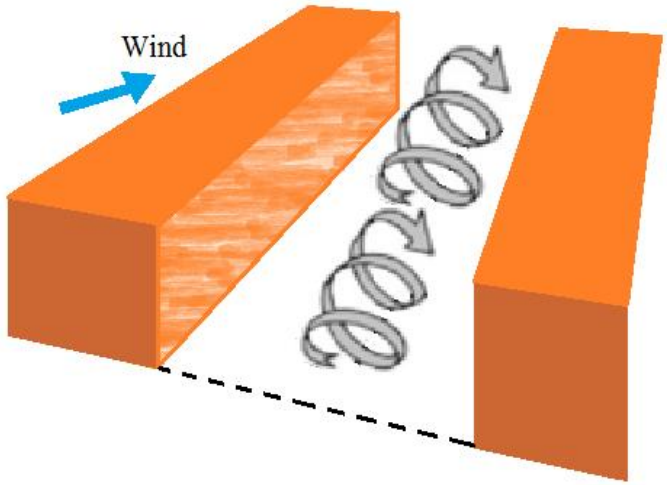
ISOLATED ROUGHNESS FLOW
AR > 2.5

FLOW IN 2D STREET CANYONS (1)

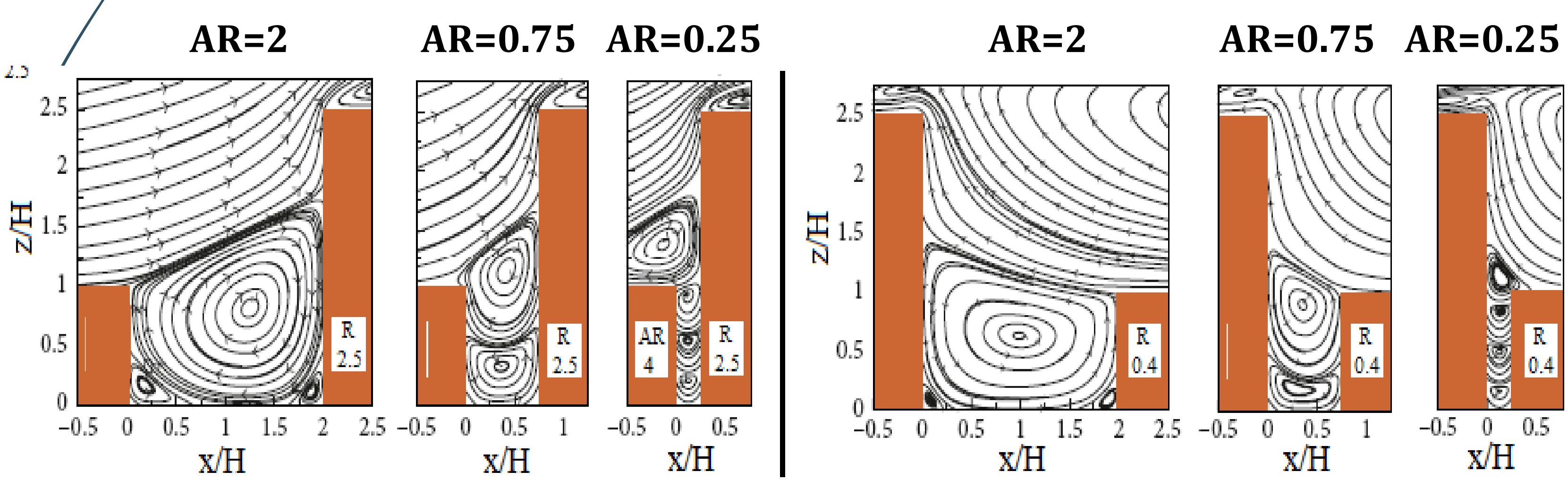


Sketch of flow topology inside a 2D street canyon with <u>external wind perpendicular to the street axis</u> (from left).

FLOW IN 2D STREET CANYONS (2)

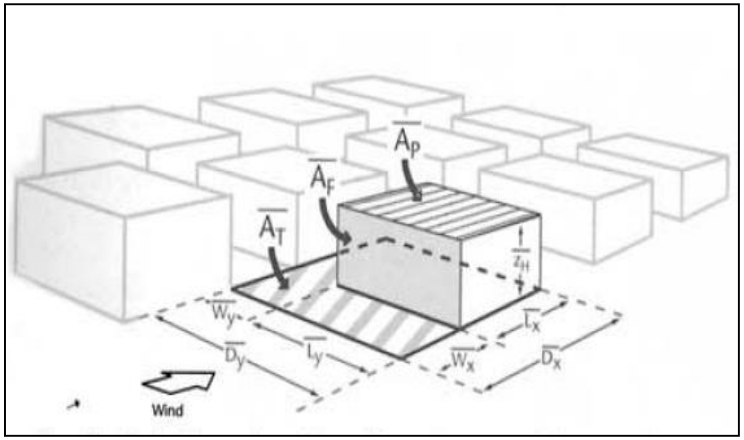


Sketch of flow topology inside a 2D street canyon in the case of external wind not perpendicular to the street axis.



Streamlines of the velocity magnitude for different aspect ratios (AR) and relative height (R) of the buildings.

FLOW REGIME IN 3D STREET CANYONS (1)



PLANAR AREA INDEX: $\lambda_P = \frac{\overline{A_P}}{\overline{A_T}}$

Mean λ_p value for European cities ≈ 0.33

SKIMMING FLOW

$\lambda_p < 0.13$

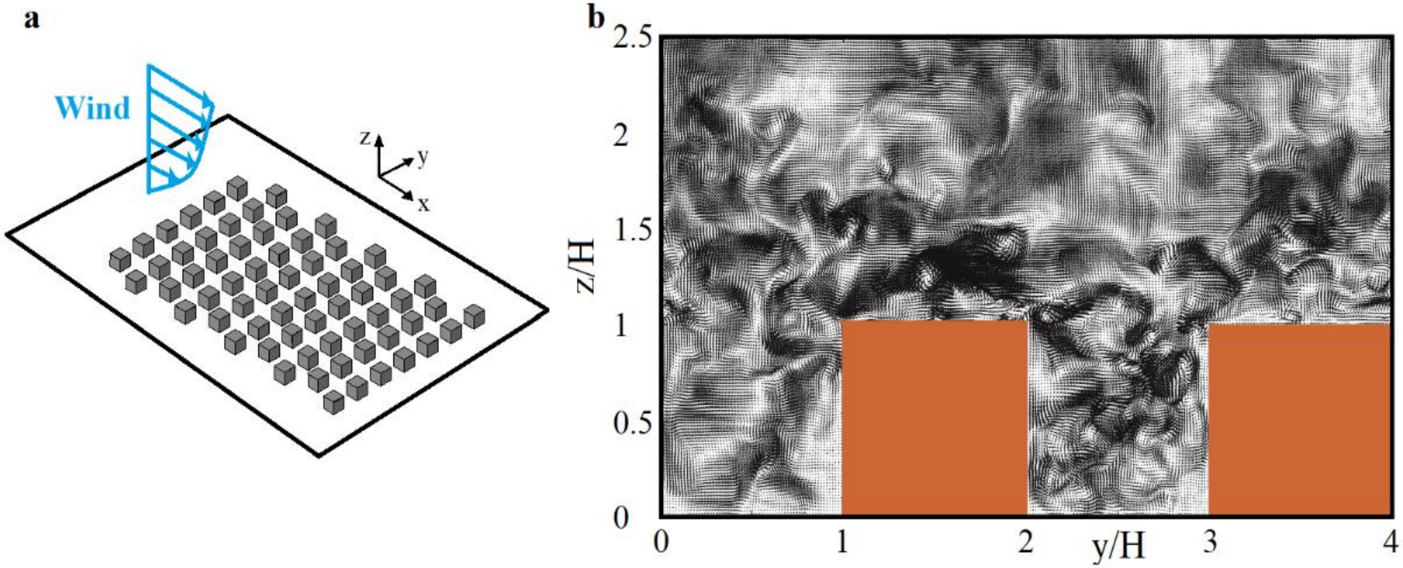
WAKE-INTERFERENCE REGIME

$0.13 \leq \lambda_p \leq 0.35$

ISOLATED ROUGHNESS FLOW

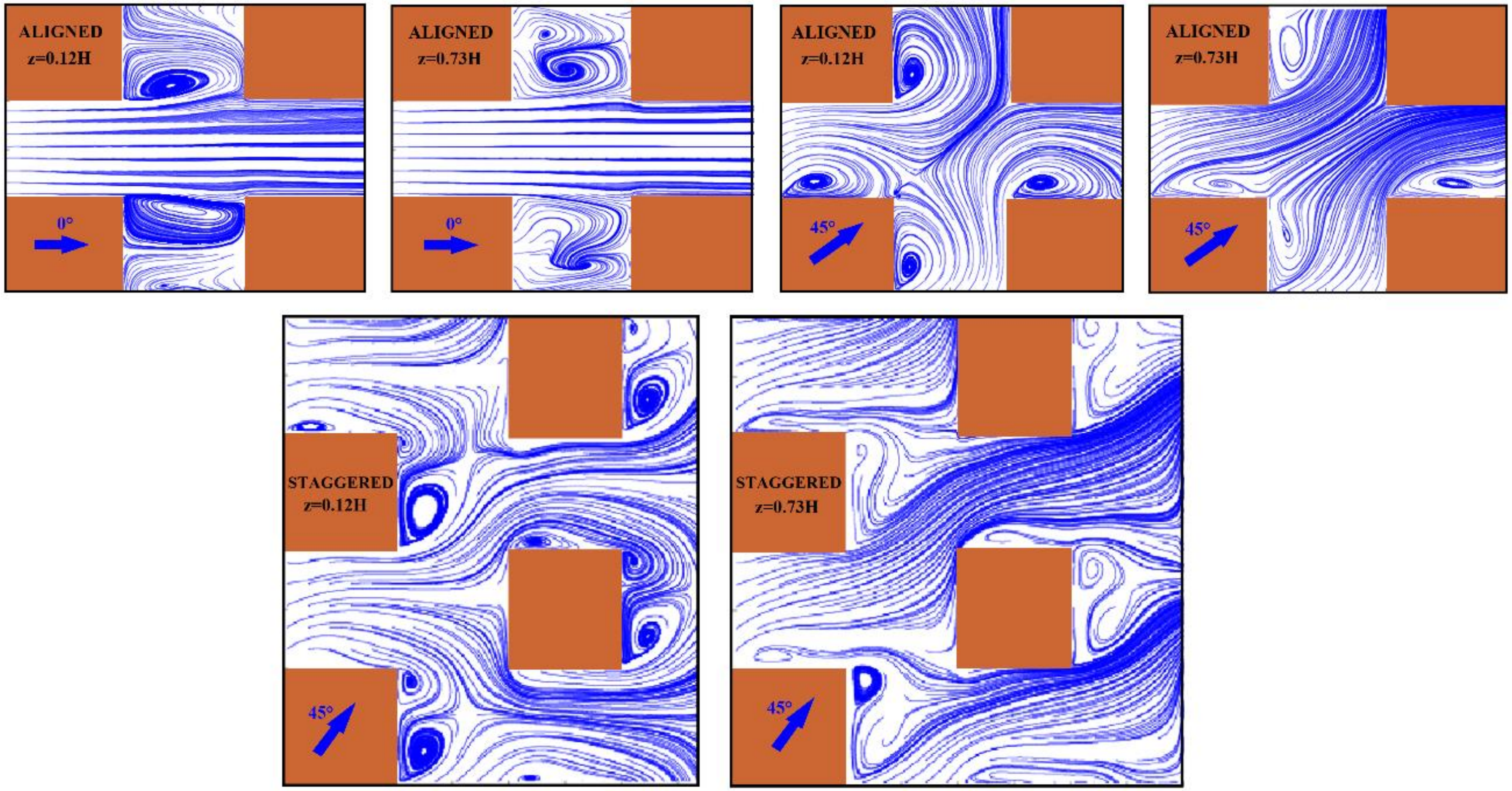
$\lambda_p > 0.35$

FLOW IN 3D STREET CANYONS (2)



a Sketch of archetypal an array of cubes usually adopted in CFD simulations and laboratory experiments
b instantaneous snapshot of the velocity field – obtained for a staggered building array with $\lambda_P = 0.25$

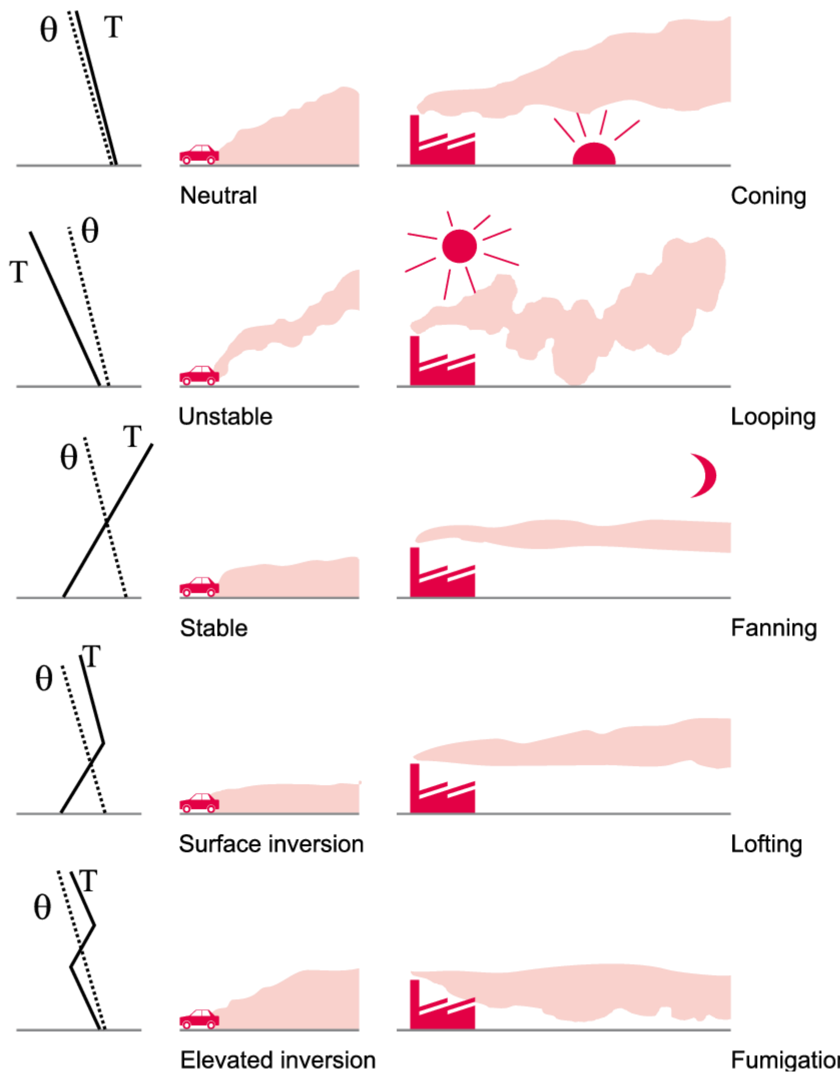
FLOW IN 3D STREET CANYONS (3)



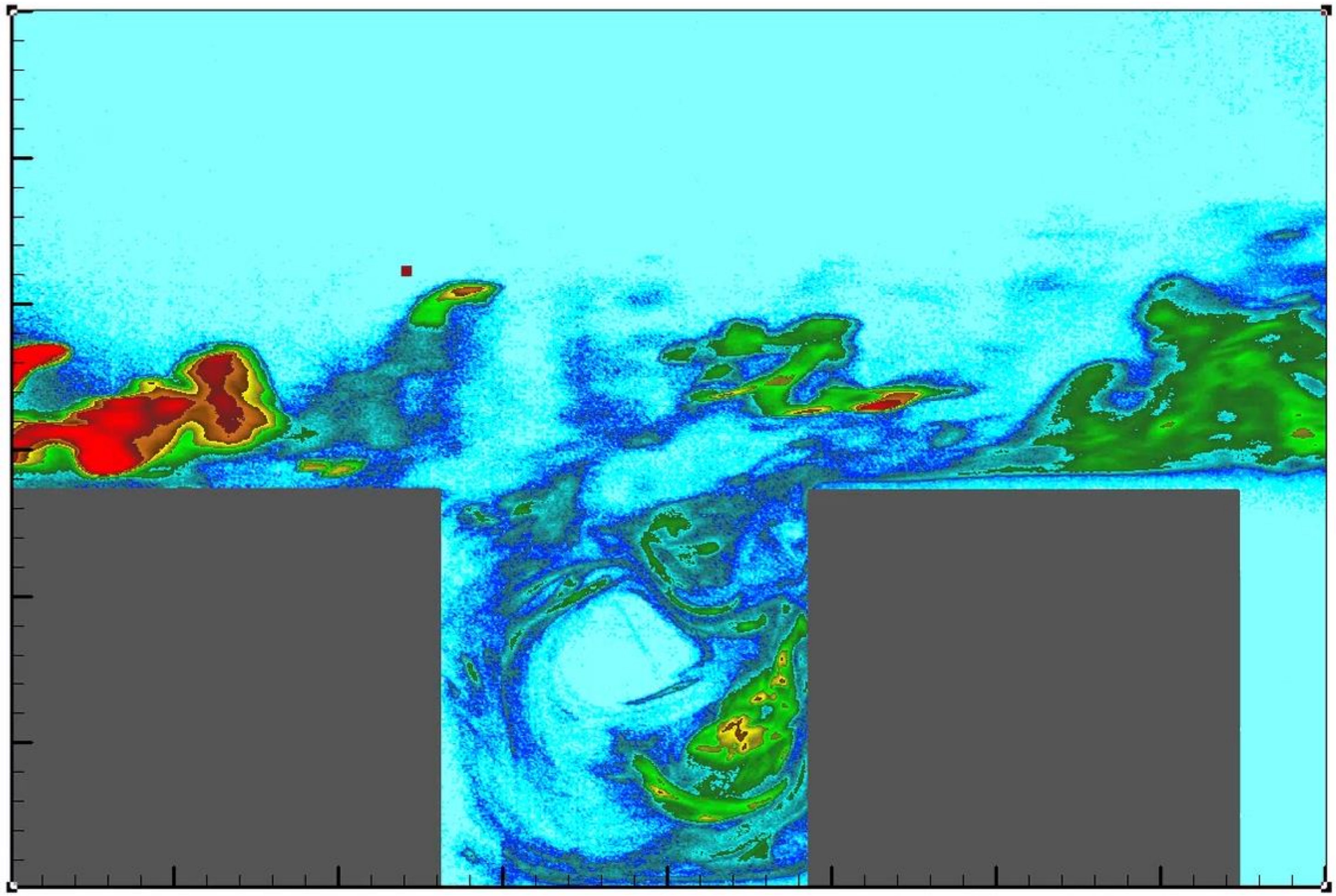
Streamlines along horizontal planes for <u>aligned and staggered regular arrays of cubical buildings for two directions of the approaching flow</u>

DISPERSION AND STABILITY

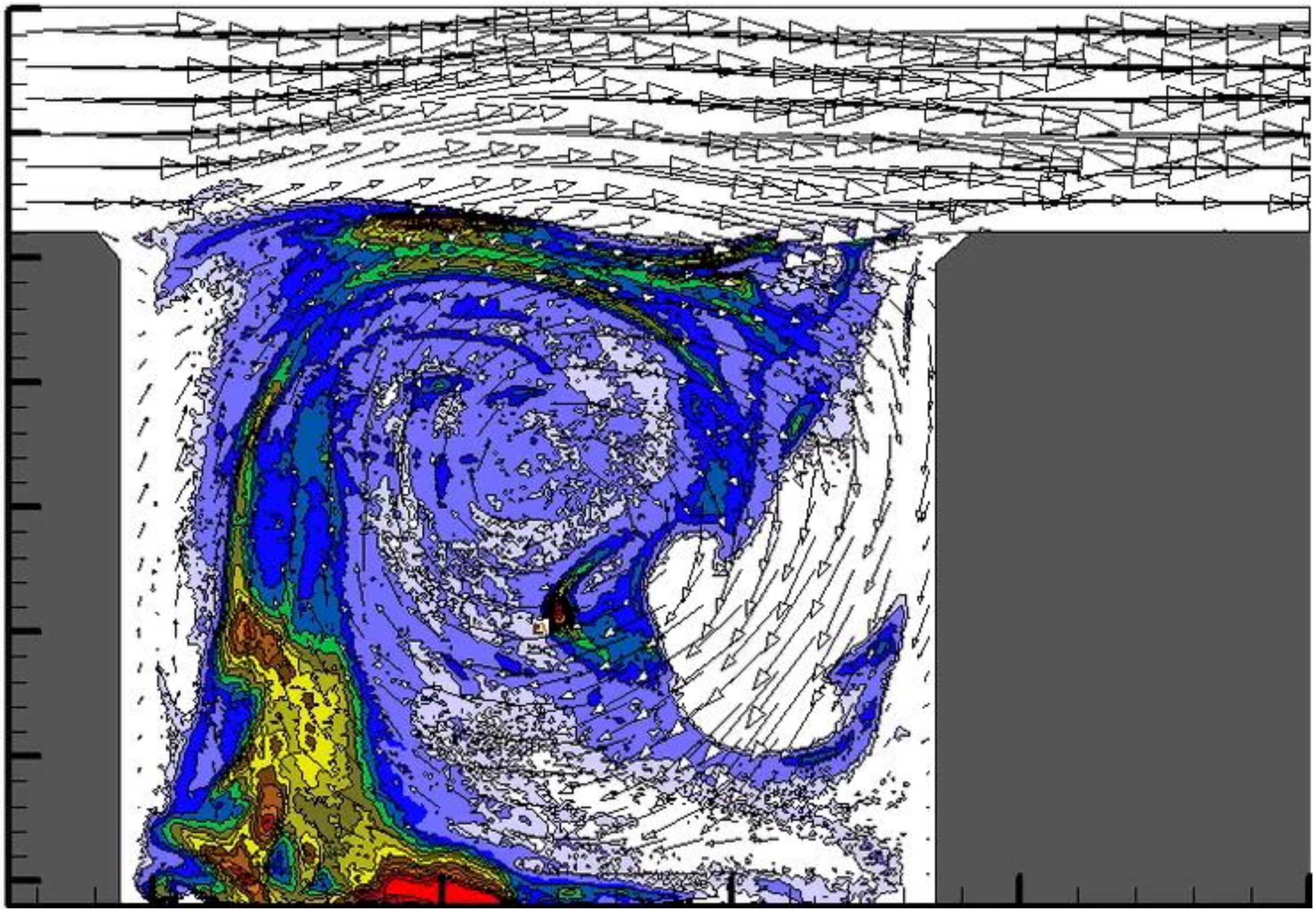
- Mean wind dominates horizontal transport.
- Turbulence dominates vertical transport.



DISPERSION IN 2D STREET CANYON



DISPERSION AND VELOCITY IN 2D STREET CANYON



PBL – IMPORTANCE AND APPLICATIONS (1)

- It is the part of the atmosphere where we live
- Almost 50% of the turbulent kinetic energy is dissipated in the PBL
- Turbulent transport and advection move water and oxygen to/from plants
- Location of source and sink of many trace gases (e.g. $CO_{2,}$ ozone, methane, water vapor), dust and pollutants
- It is very important for mesoscale weather conditions and climate change
- Thunderstorms and hurricane evolution are tied to the inflow of moist in PBL
- Crops are grown in the PBL, pollen is distributed by the PBL circulation
- Cloud nuclei are stirred into the air from the surface by PBL processes

PBL – IMPORTANCE AND APPLICATIONS (2)

- Air quality, atmospheric transport and diffusion of pollutants (local, urban and regional air quality)
- Mesoscale meteorology
- Urban planning, through flow dispersion around buildings, prediction and abatement of pollutant gases, prediction of road surface temperatures, icing
- Human comfort, heat waves and optimal design of structures and buildings
- Interaction with urban heat island
- Optimal design of wind turbines