Turbulence and atmospheric boundary layer
Lectures 10 and 11

Marta Wacławczyk, S. P. Malinowski

Institute of Geophysics, Faculty of Physics,
University of Warsaw

marta.waclawczyk@fuw.edu.pl

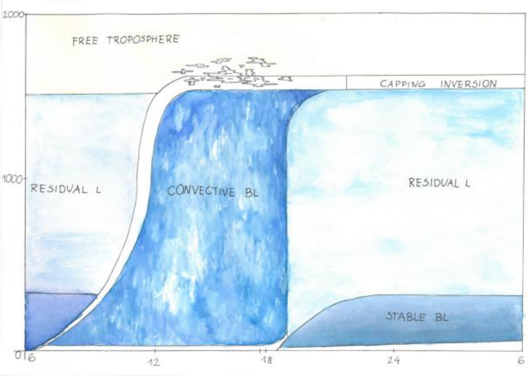
May 20, 2020

Summary of lecture 9

1. Mean winds and temperature profiles in the surface layer
   - Monin-Obukhov theory
2. Mean winds in the outer layer
   - Ekman spiral
3. Budget of turbulence kinetic energy in ABL

Surface energy balance in ABL

Recall the daily changes of the cler-sky ABL (diurnal cycle)



Surface energy balance in ABL

In this daily cycle, the solar radiation plays the role of an external forcing that drives the turbulent motions.

Surface energy balance

$$F = F_s \downarrow - F_s \uparrow + F_L \downarrow - F_L \uparrow$$

F - net radiation flux absorbed by the surface

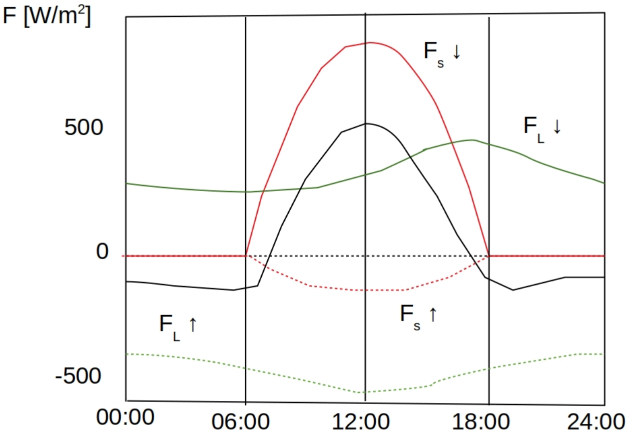
$F_s \downarrow$ - the magnitude of the flux of shortwave solar radiation that reaches the surface,

$F_s \uparrow$ - the magnitude of the flux of shortwave radiation reflected by the surface,

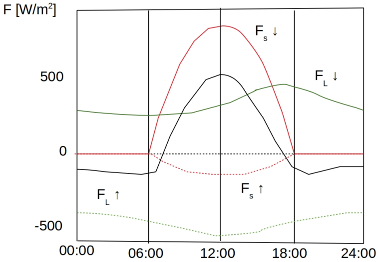
$F_L \downarrow$ - the magnitude of the flux of longwave radiation emitted by atmosphere,

$F_L \uparrow$ - the magnitude of the flux of longwave radiation emitted by the surface,

Surface energy balance in ABL



$F_s \downarrow$ - incoming solar radiation is proportional to the sine of the elevation angle of the sun
$F_s \uparrow$ - the surface reflects some of the sunlight back upward,
$F_L \downarrow$ - depends on the air temperature, which reaches its maximum in late afternoon before sunset and reaches its minimum just after sunrise,
$F_L \uparrow$ - depends on surface temperature. Because of small heat capacity, land surfaces respond almost instantaneously: so $F_L \uparrow$ is in phase with $F_s \downarrow$

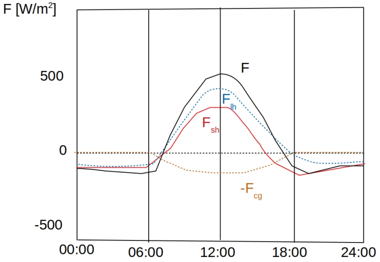


Surface energy balance in ABL

The net heat flux F is next partitioned into

$$F = F_{sh} + F_{lh} + F_{gc}$$

F_{sh} - sensible heat flux
F_{lh} - latent heat flux,
F_{gc} - conduction of heat down into the ground,

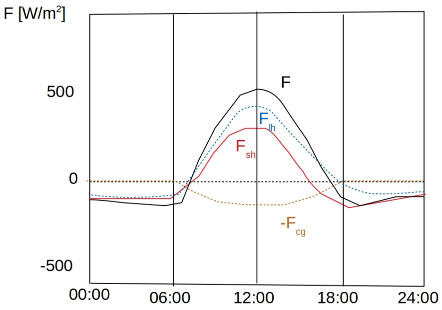


Picture on the RHS presents an exemplary partition of the energy flux for fair weather conditions

F_{sh} - during daytime over a dry, unvegetated land most of the suns energy goes into sensible heat flux,

F_{lh} - during daytime, over moist lawns, forests etc. most of the suns energy goes into evaporation,

F_{gc} - warms up the ground, the surface skin temperature, is inversely proportional to the conductivity of the soil.



Surface energy balance in ABL

Turbulence in the ocean can quickly mix heat. So, ocean surfaces exhibit larger values of F_{gc}. Moreover, the specific heat of liquid water is also larger than that of soil. Hence, the ocean has much larger heat capacity than the land surface. The ocean can absorb and store solar energy during the day and release it at night. This results in nearly constant ocean surface temperatures through the diurnal cycle and small temperature changes through the annual cycle.

Figure: Author: Tiago Fioreze, CC BY-SA 3.0 source: https://commons.wikimedia.org/wiki/File:Clouds_over_the_Atlantic_C

BL approximation

$$b = g\frac{\theta_v}{\theta_0}, \quad \theta_V = \theta\left(1 + 0.61q_V - q_l\right)$$

where θ_V is the virtual potential temperature, ρ_0 and θ_0 are characteristic ABL density and potential temperature, q_V - water vapor mixing ratio, q_l - liquid water mixing ratio

Balance for the mean θ

$$\frac{\partial\overline{\theta}}{\partial t} + \overline{u_j}\frac{\partial\overline{\theta}}{\partial x_j} = \frac{\partial}{\partial z}\underbrace{\left(\kappa\frac{\partial\overline{\theta}}{\partial z} - \overline{\theta' w'}\right)}_{\sim sensible\quad heat\quad flux} + S_\theta$$

$\rho_0 C_p \kappa \partial\overline{\theta}/\partial z$ - molecular sensible heat flux,
$\rho_0 C_p \overline{\theta' w'}$ - turbulent sensible heat flux,
$S_\theta = 1/(\rho_0 c_p)\partial F_{sh}/\partial z$ - source term

BL approximation

Balance for the mean q_v

$$\frac{\partial \overline{q_v}}{\partial t} + \overline{u_j}\frac{\partial \overline{q_v}}{\partial x_j} = \frac{\partial}{\partial z}\underbrace{\left(\kappa\frac{\partial \overline{q_v}}{\partial z} - \overline{q'_v w'}\right)}_{\sim latent\ \ heat\ \ flux} + S_q$$

$\rho_0 \mathcal{L}\kappa\partial\overline{q_v}/\partial z$ - molecular latent heat flux,
$\rho_0 \mathcal{L}\overline{q'_v w'}$ - turbulent latent heat flux,
$S_q = 1/(\rho_0\mathcal{L})\partial F_{lh}/\partial z$ - source term

BL approximation + gradient diffusion hypothesis

Balance for the mean θ

$$\frac{\partial\overline{\theta}}{\partial t} + \overline{u_j}\frac{\partial\overline{\theta}}{\partial x_j} = \frac{\partial}{\partial z}\underbrace{\left(\kappa\frac{\partial\overline{\theta}}{\partial z} + \kappa_T\frac{\partial\overline{\theta}}{\partial z}\right)}_{\sim sensible\ \ heat\ \ flux} + S_\theta$$

Balance for the mean q_v

$$\frac{\partial\overline{q_v}}{\partial t} + \overline{u_j}\frac{\partial\overline{q_v}}{\partial x_j} = \frac{\partial}{\partial z}\underbrace{\left(\kappa\frac{\partial\overline{q_v}}{\partial z} + \kappa_T\frac{\partial\overline{q_v}}{\partial z}\right)}_{\sim latent\ \ heat\ \ flux} + S_q$$

where κ_T is the eddy diffusivity.

As it is seen, the non-zero heat flux results in gradients of $\overline{\theta}$ and $\overline{q}$.

Bulk aerodynamic method to estimate the fluxes

Instead of solving the equations, we can parametrize the sensible heat flux by the temperature difference between the surface and the air $(\theta_s - \overline{\theta}_{air})$. Let us assume, the temperature changes over a layer of height H. The characteristic length scale of turbulence equals L and the characteristic velocity scale $|V|$ will be proportional to the mean wind speed at height H

$$F_{sh} = \rho_0 c_p \kappa_T \frac{\partial \overline{\theta}}{\partial x_j} \sim \rho_0 c_p L |V| \frac{(\theta_s - \overline{\theta}_{air})}{H}$$

Finally we assume

$$F_{sh} = \rho_0 c_p C_H |V| (\theta_s - \overline{\theta}_{air})$$

where C_H is a dimensionless bulk transfer coefficient for heat.

Bulk aerodynamic method to estimate the fluxes

Similarly, we can calculate the latent heat flux

$$F_{lh} = \rho_0 \mathcal{L} \kappa_T \frac{\partial \overline{q_v}}{\partial x_j} \sim \rho_0 \mathcal{L} L |V| \frac{(\overline{q_{sat}}(\theta_s) - \overline{q}_{air})}{H}$$

Finally we assume

$$F_{lh} = \rho_0 \mathcal{L} C_E |V| (q_{sat}(\theta_s) - \overline{q}_{air})$$

where $C_E \approx C_H$ is the dimensionless bulk transfer coefficient for moisture. For momentum we have

$$u_\tau^2 = C_D |V|^2$$

where C_D is the drag coefficient, u_τ is the friction velocity, $u_\tau^2 = -\overline{u'w'}$

Bulk aerodynamic method to estimate the fluxes

The coefficients C_H, C_E and C_D are not constant, but depend on the external conditions.
For example, $C_H \approx 0.001 - 0.005$ in neutral stratification and flows over flat land. Exact value depends on the surface roughness. In the unstable, convective BL, C_H becomes $2 - 3$ times larger. Conversely, as a stable BL develops, C_H decreases.
Turbulence is parametrized by the coefficient $C_H|V|$ - stronger winds means more shear = stronger turbulence production = stronger mixing.

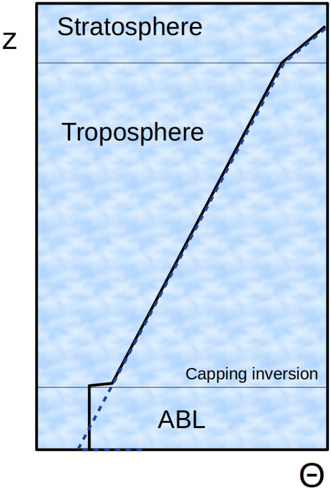
The surface temperature θ_s should be treated as a response to to solar heating. E.g. given fixed F_{sh}, larger $|V|$ results in smaller temperature difference $(\theta_s - \overline{\theta}_{air})$, hence, smaller θ_s. So, under windy conditions the surface is cooler.

Values of drag coefficient for neutral stratification

Surface	C_D
Calm sea, desert, snow-covered flat plain	0.0014
Beaches, ice, snow-covered fields	0.0028
Grass prairie or farm fields, tundra	0.0047
Cultivated area with low crops and single bushes	0.0075
High crops, scattered obstacles (e.g. trees)	0.012
Mixed farm fields and forest clumps, scattered buildings	0.018
Regular coverage with large size obstacles, suburban houses villages, forests	0.030
Centers of large towns and cities, irregular forests	0.062

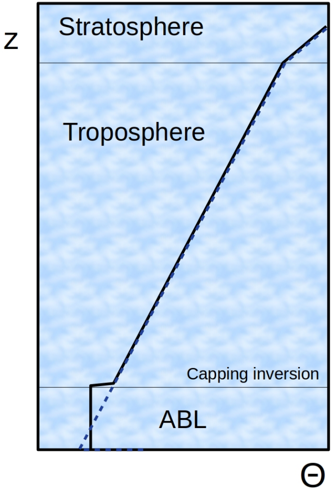
Formation of the capping inversion

- The troposphere is statically stable on average.
- Turbulence generated by processes near the ground mixes surface air of low values of potential temperature θ, with air of higher θ from upper region.



Vertical structure of ABL

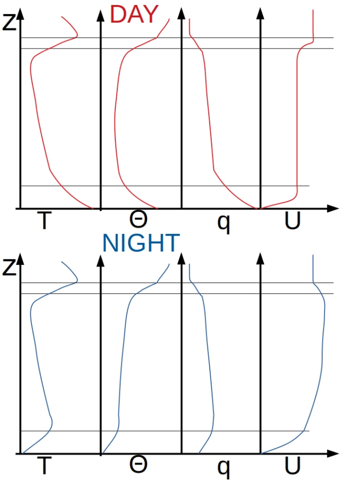
- The resulting mixture within ABL has relatively uniform potential temperature.
- This mixing has also created a temperature jump between the ABL air and the warmer air aloft.
- This is the capping inversion.
- So, turbulence creates the capping inversion, but also the capping inversion traps turbulence in ABL due to its strong stable stratification.



Daily changes of the mean profiles in ABL

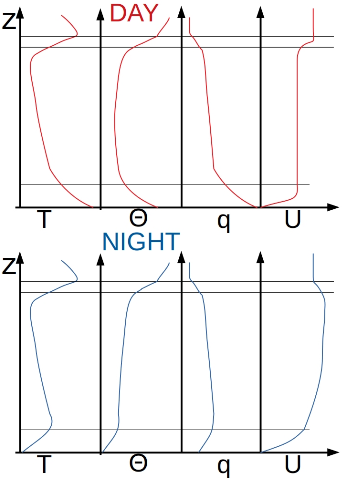
Day time

- Temperature and potential temperature near the ground increases
- Profile of $\overline{\theta}$ is nearly constant within the outer layer due to turbulent mixing (mixed layer)
- Mixing ratio increases near the ground due to evaporation



Night time

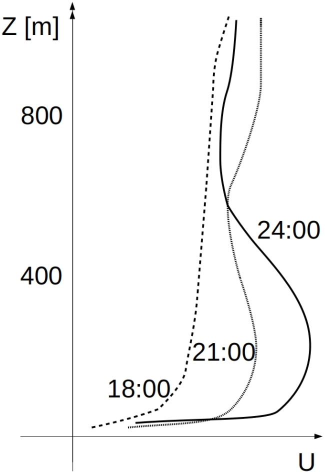
- Stable layer forms at near the ground in response to the cooling of the air by the radiatively cooled surface.
- Capping inversion formed during day is sill present
- Lower values of $\overline{q}$ near the ground due to dew or frost formation.



Daily changes of the mean profiles in ABL

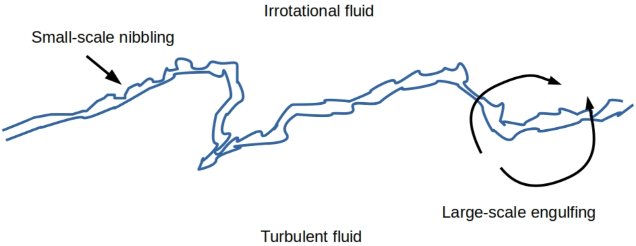
Nocturnal jet

- After the sunset turbulent fluxes across BL suddenly disappear and residual layer is formed.
- These turbulent fluxes reduced and turned the wind (Ekman spiral)
- A sudden imbalance between the main driving forces - Coriolis and the pressure gradient leads to the formation of the nocturnal jet.

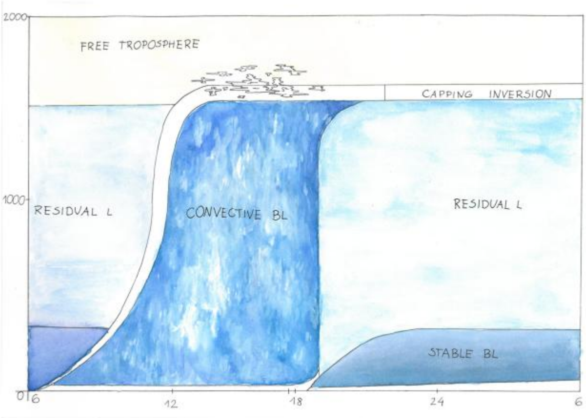


Entrainment

During daytime, thermals rise across BL. Their inertia causes them to overshoot a small distance through the nocturnal inversion.
This phenomenon is accompanied by intensive mixing and **entrainment** of irrotational fluid into BL. As a result, BL grows rapidly.
As it reaches the capping inversion, the entrainment zone becomes new capping inversion.



ABL evolution



BL growth

Let us denote the height of the BL by $h(t)$, the entrainment velocity by w_e and the vertical velocity of the large-scale motion at the top of the boundary layer by w_i (negative for subsidence).
The entrainment velocity is proportional to the sensible heat flux F_{sh} but adversly proportional to the temperature jump across the inversion $\Delta\overline{\theta}_i$.

$$w_e = \beta \frac{F_{sh}}{\rho_0 c_p \Delta\overline{\theta}_i}$$

where β is the Ball parameter

$$\beta = -\frac{F_{shi}}{F_{sh}}$$

and $F_{shi} < 0$ is the sensible heat flux across the capping inversion.

BL growth

So, the mixed layer is warmed up by the upward sensible heat flux from the surface and from the downward entrainment heat flux at the top

$$\rho_0 c_p \frac{d\overline{\theta}}{dt} = \frac{F_{sh} - F_{shi}}{h(t)} = \frac{(1+\beta)F_{sh}}{h(t)}$$

(we assume here that $\overline{\theta}$ is approximately constant within the mixed layer)
We can approximate the height of BL with the so-called encroachment method which neglects penetration into inversion due to convection, that is $\beta = 0$.

BL growth

Assume the stable temperature profile in the early morning

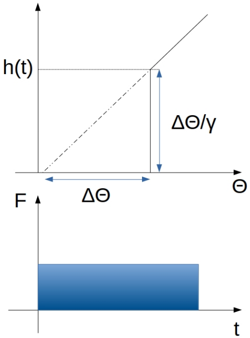
$$\frac{d\theta}{dz} = \gamma \quad \text{or} \quad \frac{dz}{d\theta} = \frac{1}{\gamma}$$

Find the value of the height $h(t)$ which crosses the line $z(\theta)$, that is

$$h(t) = z(\theta)$$

Substitute to the equation from the previous page and integrate.

$$\rho_0 c_p \int z(\theta) d\theta = \int F_{sh} dt$$



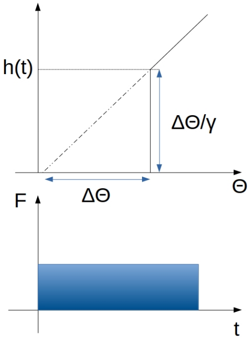
BL growth

It means we calculate the amount of heat needed to increase the BL to the height $h(t)$ from $z(\theta)$ profile and compare it with the the amount of heat calculated from the profile of the flux F

$$\rho_0 c_p \int z(\theta) d\theta = \int F_{sh} dt$$

The triangle

$$\rho_0 c_p \frac{1}{2\gamma} (\Delta\Theta)^2 = F_{sh} t$$



BL growth

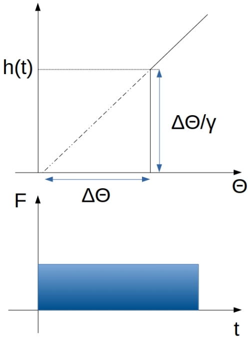
Hence,

$$\Delta\Theta = \sqrt{2\gamma F_{sh} t / \rho_0 c_p}$$

and

$$h_{enc}(t) = \sqrt{2F_{sh} t / (\gamma \rho_0 c_p)}$$

This describes the growth of convective ABL due to solar heating only (no entrainment).



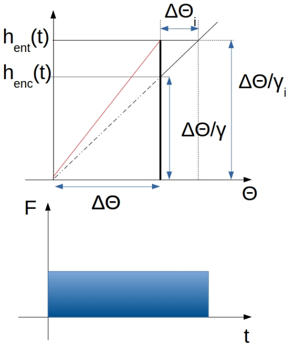
Let us now include entrainment, that is

$$\rho_0 c_p \frac{d\theta}{dt} = \frac{(1+\beta)F_{sh}}{h_{entr}(t)}$$

Similarly as before we should find the point where $h_{entr}(t)$ crosses a line

$$\frac{d\theta}{dz} = \gamma_i \quad \text{or} \quad \frac{dz}{d\theta} = \frac{1}{\gamma_i}$$

where we assume $\gamma_i = const$, so $z = \theta/\gamma_i$, so $h_{entr}(t) = \theta/\gamma_i$



BL growth

We substitute $h_{entr}(t) = \theta/\gamma_i$ into the energy equation and integrate (as we did before)

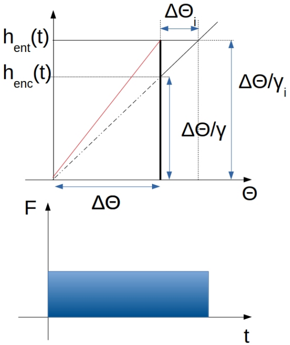
$$\rho_0 c_p \int_0^{\Delta\Theta} \theta d\theta = \gamma_i(1+\beta)F_{sh} \int_0^t dt$$

hence,

$$\rho_0 c_p (\Delta\Theta)^2 = 2\gamma_i(1+\beta)F_{sh} t$$

Recall that for the case without entrainment we obtained

$$\rho_0 c_p (\Delta\Theta)^2 = 2\gamma F_{sh} t$$



BL growth

Comparing the two equalities we obtain

$$\gamma_i = \frac{\gamma}{1+\beta}$$

Now, $h_{entr} = \Delta\Theta/\gamma_i$, so

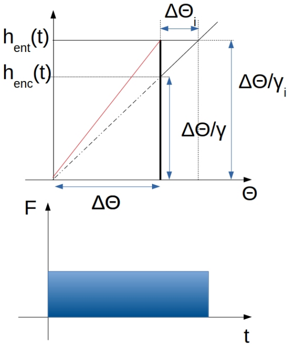
$$h_{entr}(t) = \sqrt{2(1+\beta)F_{sh}t/(\gamma_i \rho_0 c_p)}$$

or with $\gamma_i = \gamma/(1+\beta)$

$$h_{entr}(t) = (1+\beta)\sqrt{2F_{sh}t/(\gamma \rho_0 c_p)}.$$

Hence, finally

$$h_{entr}(t) = (1+\beta)h_{enc}(t).$$



Stable ABL

Stable BL

- one of more difficult types of BL to understand
- BL tends to be 50-300 m depth
- start to form at dusk due to radiative cooling of the Earths surface
- turbulence tends to be intermittent (local re-laminarisations)
- gravity wave-like motions are present especially near the top of BL

Figure: Smog over Almaty city, Kazakhstan, author: Igors Jefimovs, source: Wikipedia, CC BY-3.0

Stable BL

- even the largest eddies do not span entire BL, so chemicals and aerosols have a tendency of layering
- turbulence in SBL is formed mechanically, due to shear
- the buoyant contribution is much smaller and it is generally a destruction process
- since vertical motions are suppressed turbulence is neither homogeneous nor isotropic

Figure: Smog over Almaty city, Kazakhstan, author: Igors Jefimovs, source: Wikipedia, CC BY-3.0

External intermittency

The intermittency coefficient γ measures the fraction of turbulent fluid over the sampling space over which the statistics are taken.
$\gamma = 0$ if the flow is purely laminar,
$\gamma = 1$ if it is fully turbulent and

$$0 < \gamma < 1$$

if it is intermittent.

Figure: Vorticity in fully turbulent vs. intermittent flow [Ansorge & Mellado, J Fluid Mech (805), 611-635, 2016]

Stable BL

Useful parameters to describe SBL

Brunt-Väisälä frequency

$$N = \sqrt{\frac{g}{\overline{\theta}_v}\frac{d\overline{\theta}_v}{dz}}$$

It is the maximum frequency of internal gravity waves. Waves within the SBL are trapped between the ground and the neutral layers above SBL, hence they propagate horizontally.

Flux Richardson number

$$Ri = \frac{\frac{g}{\overline{\theta}_v}\overline{w'\theta'_v}}{\overline{u'w'}\frac{\partial\overline{u}}{\partial z} + \overline{v'w'}\frac{\partial\overline{v}}{\partial z}}$$

Ri increases with stratification, when $Ri > Ri_c$ flow is assumed to re-laminarise. $Ri_c \sim 1$.

Cloud-topped boundary layers

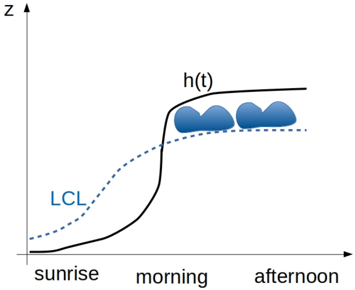
Cloud-Topped Boundary Layer over Land

On days when sufficient moisture is present in the BL, the lifting condensation level (LCL) can be below $h(t)$. In such case, the tops of the thermals that extend above the LCL are filled with cumulus clouds.

Figure: Cumulus humilis clouds in the foreground, taken at Swifts Creek, in the Great Alps of East Gippsland, Victoria, Australia. Author: Fir0002, Fir0002/Flagstaffotos, source: Wikipedia, CC BY-NC

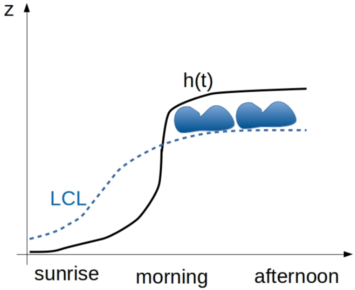
Cloud-Topped Boundary Layer over Land

- short after sunrise BL top rises slowly while nocturnal inversion is burned off
- but the LCL level rises rapidly due to sun's warming (lowering of the relative humidity)
- as a result, BL top is cloudless in the morning



Cloud-Topped Boundary Layer over Land

- when the nocturnal inversion disappears in the morning BL top rises rapidly
- it jumps to the level $\sim 1 - 2km$ which can be above LCL level
- then fair-weather cumulus clouds are most likely to form
- the top of these clouds does not penetrate beyond the capping inversion



Cloud-Topped Boundary Layer over Land

- if there is sufficient vertical wind shear $\partial\overline{u}/\partial z$, $\partial\overline{v}/\partial z$ horizontal roll vortices can be formed
- These circulations organize the stronger thermals into lines, oriented parallel to the mean wind direction

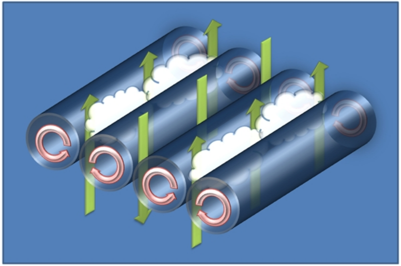


Figure: Horizontal convective rolls, author: Daniel Tyndall, Departmant of Meteorology, University of Utah, source: https://pl.m.wikipedia.org/wiki/Plik:Convrolls.PNG (public domain)

Cloud-Topped Boundary Layer over Land

- if there is sufficient vertical wind shear $\partial \overline{u}/\partial z$, $\partial \overline{v}/\partial z$ horizontal roll vortices can be formed
- These circulations organize the stronger thermals into lines, oriented parallel to the mean wind direction
- water vapor condenses on the upward side of the roll, evaporation happens on the downward side of the roll.

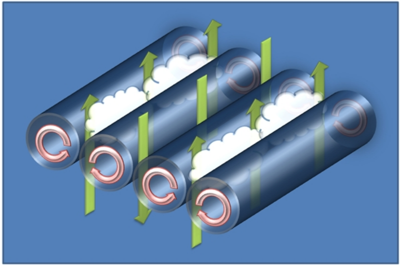


Figure: Horizontal convective rolls, author: Daniel Tyndall, Departmant of Meteorology, University of Utah, source: https://pl.m.wikipedia.org/wiki/Plik:Convrolls.PNG (public domain)

Cloud-Topped Boundary Layer

- height of the rolls $\sim 1km$, width $\sim 1-10km$
- rolls form the cloud-streets
- formation mechanism is not well understood, but is related to the Ekman spiral and the presence of unstable convective boundary layer below and stable inversion layer above

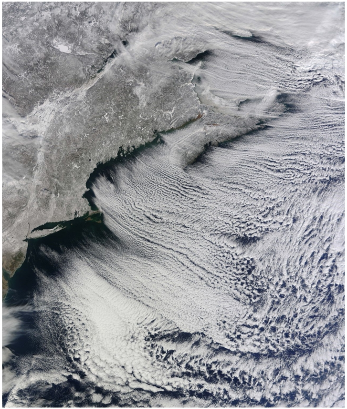


Figure: Horizontal convective rolls, NASA image by Jeff Schmaltz source: https://earthobservatory.nasa.gov/images/49254/winter-cloud-streets-north-atlantic (public domain)

Marine boundary layers

The cloud-topped boundary layer over the oceans are different from their land counterparts due to

- higher relative humidities
- larger cloud cover
- more important and complex role of radiative transfer
- significant role of drizzle plays in the boundary-layer heat and water balance
- diurnal cycle governed by different physics

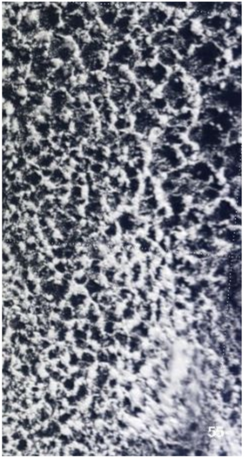


Figure: Cellular convection, author: Imaleaper source: https://en.wikipedia.org/wiki/File:Open_Cellular_Convection.JPG,

Marine boundary layers

- Cloud cover reduces insolation...
- but also reduces outgoing longwave radiation from the underlying water.
- Cloud cover also emits the longwave radiation.
- For thin BL ($\sim 1km$) clouds are not much cooler than the Earth surface, so the net effect is the radiative cooling of the BL.
- This keeps relative humidity inside BL high.

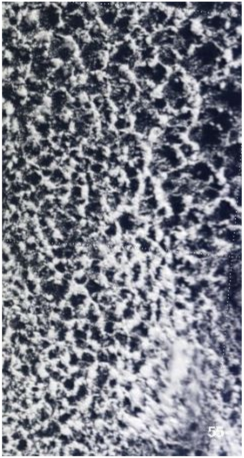


Figure: Cellular convection, author: Imaleaper source: https://en.wikipedia.org/wiki/File:Open_Cellular_Convection.JPG,

Marine boundary layers

- As a result of net radiative cooling, the capping inversion strengthens.
- This in turn reduces entrainment of dry air into the cloud layer

$$w_e = \beta \frac{F_{sh}}{\rho_0 c_p \Delta\overline{\theta}_i}$$

(larger $\Delta\overline{\theta}_i$ means smaller w_e)

- This contributes to the maintenance of the cloud deck.

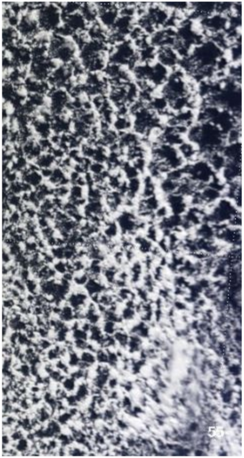


Figure: Cellular convection, author: Imaleaper source: https://en.wikipedia.org/wiki/File:Open_Cellular_Convection.JPG, CC BY 3.0

Marine boundary layers

Convection can be driven by

- heating from below by the surface buoyancy flux
- cooling from above by the flux of longwave radiation at the top of the cloud layer.

This induces formation of cell-like structures (open or closed). More about in the following lectures...

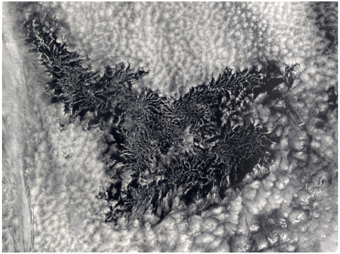


Figure: Open-cell and closed-cell clouds off California, Pacific Ocean, credit: Jacques Descloitres, MODIS Rapid Response Team, NASA/GSFC, source: https://visibleearth.nasa.gov/images/61786/open-cell-and-closed-cell-clouds-off-california-pacific-ocean

Boundary layer - terrain effects

Thermally driven circulations

- During fair-weather conditions, when mountain slopes are heated by the sun, the warm air rises along the slope forming an *anabatic wind*.

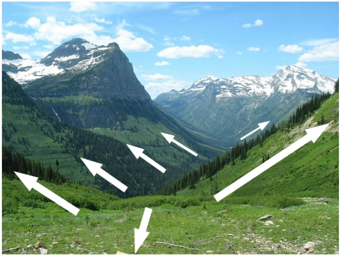


Figure: Modified picture of Glacier National Park in Montana, U.S., author: BorisFromStockdale source: https://commons.wikimedia.org/wiki/File:Glacier_park1.jpg, CC BY-SA 3.0

Thermally driven circulations

- After the sunset, when the mountain slopes are cooled by longwave radiation, the cold, heavier air sinks downslope as a cold *katabatic wind*.

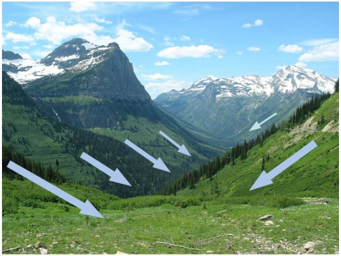


Figure: Modified picture of Glacier National Park in Montana, U.S., author: BorisFromStockdale source: https://commons.wikimedia.org/wiki/File:Glacier_park1.jpg, CC BY-SA 3.0

Sea and land breezes

- Due to small heat capacity land surfaces react much faster to changes of insolation than the ocean surface - the air above land heats up and cools down faster than the air over ocean.
- Lands are hotter in the afternoons and cool down faster at night.

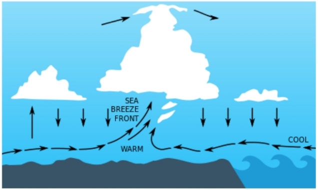


Figure: Schematic cross section through a sea breeze front, source: U.S. FAA AC 00-6A, Chapter 16, Figure 165. (public domain)

Sea and land breezes

- The resulting temperature difference between land and ocean cause horizontal pressure gradients that drive shallow, diurnally varying winds: daytime sea breezes and land breezes at nighttime.

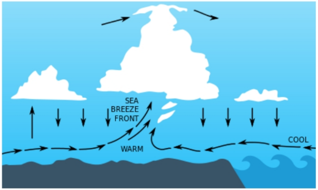


Figure: Schematic cross section through a sea breeze front, source: U.S. FAA AC 00-6A, Chapter 16, Figure 165. (public domain)

Consider a slope of angle α. Let us rotate the coordinate system such that x and y axes are parallel to the slope (x is the downslope direction) and z is perpendicular to it.

Simplified momentum balance - x direction

$$\frac{\partial \overline{u}}{\partial t} + \overline{u_j}\frac{\partial \overline{u}}{\partial x_j} = -\frac{1}{\rho_0}\frac{\partial \overline{p}}{\partial x} + \overline{b} sin\alpha$$

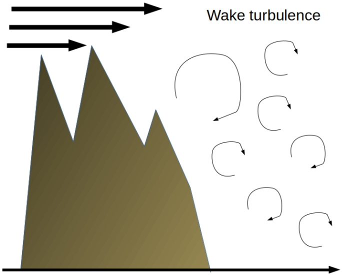
$$b = g\frac{\Delta\theta_v}{\theta_0},$$

$\overline{b} sin\alpha$ acts as a driving force.

Boundary layer - terrain effects

Strong wind conditions

- Stronger winds evolve due to the synoptic-scale forcing.
- They cause a number of phenomena on smaller scales, including local accelerations of the wind over crests, blocking of low-altitude winds, cloud formation, wake turbulence, Karman vortex streets, downslope wind storms, and hydraulic jumps.

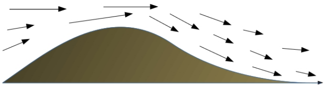


Boundary layer - terrain effects

Strong wind conditions

- local accelerations of the wind over crests

Recall mass flux across the stream $US = const$, where U is the velocity and S is the cross-sectional area.



Strong wind conditions

- Von Kármán vortex sheets form in regions where fluid flow is disturbed by an object. In the atmosphere, the resulting vortex streets can be accompanied by the characteristic formations of stratocumulus cloud sheets.

$$St = \frac{fL}{U}$$

where St is the Strouhal number, f is the shedding frequency, L is the characteristic length.

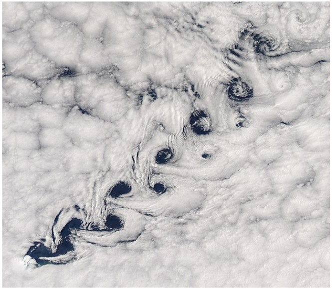
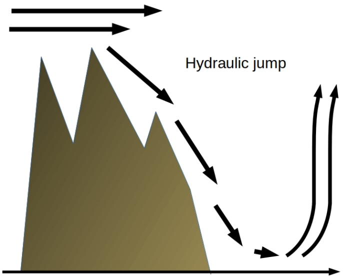


Figure: Cloud vortices in the cloud layer off Heard Island, author: NASA / GSFC / Jeff Schmaltz / MODIS Land Rapid Response Team, (public domain)

Strong wind conditions

- hydraulic jumps are formed when high velocity fluid flows into a zone of lower velocity
- the fluid horizontal velocity rapidly decreases and the fluid ascends



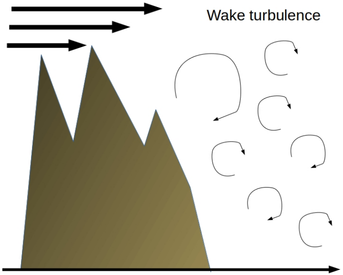
Boundary layer - terrain effects

Strong wind conditions

- Flow configuration which can occur can be predicted by the Froude number

$$Fr = \frac{U}{NL}$$

where N is the Brunt-Väisälä frequency, L is the height of the mountain. At small Fr, the airflow is forced to go around the mountain and/or through gaps. At larger values of Fr more flow over the top of the mountain occurs.



References

J.R. Garratt (1992)
The atmospheric boundary layer
Cambridge University Press

R. Stull (2005)
The atmospheric boundary layer In: Atmospheric Science (Eds. J. Wallace,P. V. Hobbs)
Elsevier

The End