Parametrization of the planetary boundary layer (PBL)

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Introduction
Surface layer and surface fluxes
Outer layer
PBL& Cloud evaluation
Exercises

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Why studying the Planetary Boundary Layer?

- Natural environment for human activities
- Understanding and predicting its structure
  - Agriculture, aeronautics, telecommunications, Earth energetic budget
- Weather forecast, pollutants dispersion, climate prediction
Outline

Definition

Turbulence/Equations

Stability

Classification

Clear convective boundary layers

Cloudy boundary layers (stratocumulus and cumulus)

Summary
The PBL is the layer close to the surface within which vertical transports by turbulence play dominant roles in the momentum, heat and moisture budgets.

- The layer where the flow is turbulent.
- The fluxes of momentum, heat or matter are carried by turbulent motions on a scale of the order of the depth of the boundary layer or less.
- The surface effects (friction, cooling, heating or moistening) are felt on timescales < 1 day.

Composition

- atmospheric gases ($N_2$, $O_2$, water vapor, …)
- aerosol particles
- clouds (condensed water)
PBL: Governing equations for the mean state

- gas law (equation of state)
- momentum (Navier Stokes)
- continuity eq. (conservation of mass)
- heat (first principle of thermodynamics)
- total water

Reynolds averaging $A = \bar{A} + A'$

Averaging (overbar) is over grid box, i.e. sub-grid turbulent motion is averaged out.

**Simplifications**

- **Boussinesq** approximation (density fluctuations non-negligible only in buoyancy terms)
- **Hydrostatic** approximation (balance of pressure gradient and gravity forces)
- **Incompressibility** approximation (changes in density are negligible)
PBL: Governing equations for the mean state

Reynolds averaging \( A = \overline{A} + A' \)

1. **Gas law**
   \[
   \overline{p} = \overline{\rho} R_d \overline{T_v}
   \]
   virtual temperature

2. **Momentum**
   \[
   \frac{\partial \overline{u}_i}{\partial t} + \overline{u}_j \frac{\partial \overline{u}_i}{\partial x_j} = -\delta_{i3} g + f_c \varepsilon_{ij3} \overline{u}_j - \frac{1}{\rho} \frac{\partial \overline{P}}{\partial x_i} + \nu \frac{\partial^2 \overline{u}_i}{\partial x_j^2} - \frac{1}{\partial x_j} \frac{\partial (u'_i u'_j)}{\partial x_j}
   \]
   mean advection  gravity  Coriolis  Pressure gradient  Viscous stress  Turbulent transport

3. **Continuity equation**
   \[
   \frac{\partial \overline{u}_i}{\partial x_j} = 0
   \]

4. **Heat**
   \[
   \frac{\partial \overline{\theta}}{\partial t} + \overline{u}_j \frac{\partial \overline{\theta}}{\partial x_j} = - \frac{1}{\rho c_p} \frac{\partial \overline{F}_j}{\partial x_j} - \frac{\partial \overline{u}_j' \theta'}}{\partial x_j} - \frac{L_v E}{\rho c_p}
   \]
   mean advection  radiation  turbulent transport  Latent heat release

5. **Total water**
   \[
   \frac{\partial \overline{q}_t}{\partial t} + \overline{u}_j \frac{\partial \overline{q}_t}{\partial x_j} = \frac{S_{qt}}{\rho} - \frac{\partial \overline{u}_j' q'_t}{\partial x_j}
   \]
   mean advection  precipitation  turbulent transport
TKE: a measure of the intensity of turbulent mixing

\[ e = \frac{1}{2} (u'^2 + v'^2 + w'^2) \]

**PBL: Turbulent kinetic energy equation**

\[ \frac{\partial e}{\partial t} + \mathbf{u}_j \frac{\partial e}{\partial x_j} = \left( \frac{g}{\theta_0} \frac{w' \theta_v'}{\theta_v} \right) - \frac{u_i' u_j'}{\partial x_j} - \frac{\partial u_i' e}{\partial x_j} - \frac{1}{\rho} \frac{\partial u_i' p'}{\partial x_i} - \epsilon \]

- Buoyancy production
- Mechanical shear
- Turbulent transport
- Pressure transport
- Dissipation

\[ \theta_v = \theta \left( 1 + 0.61 q_v - q_1 \right) \]

**An example:**

- \( \theta_v' < 0 \), \( w' < 0 \) \( \Rightarrow \) \( w' \theta_v' > 0 \) \( \text{source} \)
- \( \theta_v' > 0 \), \( w' > 0 \) \( \Rightarrow \) \( w' \theta_v' > 0 \)
- \( \theta_v' < 0 \), \( w' > 0 \) \( \Rightarrow \) \( w' \theta_v' < 0 \) \( \text{sink} \)
- \( \theta_v' > 0 \), \( w' < 0 \) \( \Rightarrow \) \( w' \theta_v' < 0 \)
Traditionally stability is defined using the temperature gradient.

$\theta_v$ gradient (local definition):

- $\frac{\partial \theta_v}{\partial z} > 0$ stable layer
- $\frac{\partial \theta_v}{\partial z} < 0$ unstable layer
- $\frac{\partial \theta_v}{\partial z} = 0$ neutral layer

How to determine the stability of the PBL taken as a whole?

- In a mixed layer the gradient of temperature is practically zero.
- Either $\theta_v$ or $w^' \theta_v$ profiles are needed to determine the PBL stability state.
Bouyancy flux at the surface:

\[ w' \theta'_v > 0 \quad \text{unstable PBL (convective)} \]

\[ w' \theta'_v < 0 \quad \text{stable PBL} \]

\[ w' \theta'_v = 0 \quad \text{neutral PBL} \]

Or dynamic production of TKE integrated over the PBL depth stronger than thermal production

Monin-Obukhov length:

\[ L = \frac{-\overline{\theta'_v u^3}}{\overline{kg(w'_v \theta'_v)_s}}, \quad u_*^2 = (u'w')_s \]

\[-10^5m \leq L \leq -100m \quad \text{unstable PBL}\]

\[-100m < L < 0 \quad \text{strongly unstable PBL}\]

\[0 < L < 10 \quad \text{strongly stable PBL}\]

\[10m \leq L \leq 10^5m \quad \text{stable PBL}\]

\[|L| > 10^5m \quad \text{neutral PBL}\]
PBL: Classification and scaling

Neutral PBL:
- turbulence scale $l \sim 0.07 \, H$, $H$ being the PBL depth
- Quasi-isotropic turbulence
- Scaling - adimensional parameters: $z_0$, $H$, $u^*_w$

Stable PBL:
- $l \ll H$ (stability embeds turbulent motion)
- Turbulence is local (no influence from surface), stronger on horizontal
- Scaling: $(w' \theta')_s \, (u'w')_s \cdot H$

Unstable (convective) PBL
- $l \sim H$ (large eddies)
- Turbulence associated mostly to thermal production
- Turbulence is non-homogeneous and asymmetric (top-down, bottom-up)

Scaling: $H$, $w_* = \left( \frac{g}{\theta_v} (w' \theta'_v)_s H \right)^{1/3}$, $z = \frac{z}{H}, q_* = \frac{E_0}{w_*}, \theta_* = \frac{Q_0}{w_*}$
PBL: Diurnal variation

- **Clear convective PBL**
- **Cloudy convective PBL**
- **Inversion**
- **Residual Layer**
- **Convective Mixed Layer**
- **Stable (nocturnal) Layer**

Adapted from *Introduction to Boundary Layer Meteorology* - R.B. Stull, 1988
PBL: State variables

**Clear PBL**

Specific humidity

\[ q_v = \frac{m_v}{m_d + m_v} \]

Potential temperature

\[ \theta = T \left( \frac{p}{p_0} \right)^{-R_d / c_p} \]

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**Cloudy PBL**

Total water content

\[ q_t = \frac{m_v + m_c}{m_d + m_v + m_c} \]

Liquid water potential temperature

\[ \theta_l \approx \theta - \frac{L_v}{c_p} q_l \]

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no liquid water is condensed \( q_l = 0 \)

[Diagram showing no liquid water is condensed (left side) and liquid water is condensed (right side).]

Conserved variables

Conserved variables

Cloud base

[Diagram showing Cloud base.]
Clear Convective PBL

- Buoyantly-driven from surface

- 0.7 H

- Turbulent transport and pressure term

- Level of neutral buoyancy

- Main source of TKE production

- PBL height: \( \frac{dH}{dt} = \bar{w} + W_e \)

- Entrainment rate: \( w_e = A \frac{w^3}{g \frac{\theta_0}{H} \Delta \theta_v} \) (a possible parameterization)

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- Fluxes at PBL top: \( w' \psi_H' = -W_e \Delta \psi \)

- Key parameters: \( W_e, \Delta \theta_v, H, (w' \theta'_v)_0 \)
Clouds effects on climate

Greenhouse effect: warming
- High clouds, like cirrus
- Infrared radiation

Umbrella effect: cooling
- Boundary layer clouds (low clouds)
- Solar radiation
- Shallow cumulus
- Stratocumulus
Boundary layer clouds over oceans

- Cloud Clusters
- Hadley/Walker Circulation
- Land/Sea Circulation

EQ warm western tropical oceans

- Shallow cumulus
- Stratocumulus

EQ cold eastern subtropical ocean

SST & surface wind
Cloudy boundary layers

Stephan de Roode, Ph.D thesis

Free atmosphere

Inversion layer

mixed layer

surface layer

Stratocumulus topped PBL

Cumulus PBL

Dry-adiabatic lapse rate

wet-adiabatic lapse rate

Conditionally unstable cloud layer

Subcloud mixed surface layer
Stratocumulus – Why are they important?

- Cover in (annual) mean 29% of the planet (Klein and Hartmann, 1993)
- Cloud top albedo is 50-80% (in contrast to 7% at ocean surface).
- A 4% increase in global stratocumulus extent would offset 2-3K global warming from CO$_2$ doubling (Randall et al. 1984).
- Coupled models have large biases in stratocumulus extent and SSTs.

Stratocumulus cloud cover (annual mean)

Wood, 2012, based on Han and Waren, 2007
Characterisation of a stratocumulus topped PBL
Subsidence

\[ \theta_l(K) \]

\[ r_t(g \text{ kg}^{-1}) \]

\[ r_{sat}(g \text{ kg}^{-1}) \]

\[ h_{STBL}(m) \]

\[ LWP(\text{kg m}^{-2}) \]

(Liquid Water Path)

Such a cloudy system is extremely sensitive to thermodynamical conditions
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Characterisation of a stratocumulus topped PBL

Such a cloudy system is extremely sensitive to thermodynamical conditions.
Processes controlling the evolution of a non-precipitating stratocumulus

Night-time

Entrainment warming, drying

Well-mixed

Surface heat and moisture fluxes

LW radiative cooling

Warm, dry, subsiding free-troposphere

The cloud deepens

Courtesy of Bjorn Stevens (data from DYCOMS-II)
Processes controlling the evolution of a non-precipitating stratocumulus

**Daytime**

- **Entrainment warming, drying**
- **Stabilisation of the subcloud layer**
- **Surface heat and moisture fluxes**
- **Warm, dry, subsiding free-troposphere**
- **SW absorption**
- **LW radiative cooling**
- **Decoupled**
- **The cloud thins**

*Courtesy of Bjorn Stevens (data from DYCOMS-II)*
Diurnal cycle during observed during FIRE-I experiment

Hignett, 1991 (data from FIRE-I)
Precipitation flux

Under cloud evaporation affects the dynamics of the boundary layer.
Stratocumulus topped boundary layer

- Complicated turbulence structure
  - Cloud top entrainment
  - Long-wave cooling
  - Condensation
  - Long-wave warming

- Buoyantly driven by radiative cooling at cloud top
- Surface latent and heat flux play an important role
- Cloud top entrainment an order of magnitude stronger than in clear PBL
- Solar radiation transfer essential for the cloud evolution
- Key parameters: $w_e, \Delta \theta_v, H, (w'\theta'_v)_0, (w'q'_v)_0, \Delta q_t, z_b, \Delta F$
Cloudy boundary layers

Stratocumulus topped PBL

Cumulus PBL
Cumulus capped boundary layers

Buoyancy is the main mechanism that forces cloud to rise
Cumulus capped boundary layers

- Buoyancy is the main mechanism that forces cloud to rise

- Represented by mass flux convective schemes: \( M_c (\psi_u - \psi_d) = kw' \psi' \)

- Decomposition: cloud + environment

- Lateral entrainment/detainment rates prescribed

- Key parameters: \( H, z_b, (w' \theta_v')_0, (w' q_v')_0 \)
  \[ \left( \frac{\partial \theta_v}{\partial z} \right)_{environ}, \left( \frac{\partial q_v}{\partial z} \right)_{environ} \]
Characteristics:
- several thousands of meters – 2-3 km above the surface
- turbulence, mixed layer
- convection
- Reynolds framework

Classification:
- neutral (extremely rare)
- stable (nocturnal)
- convective (mostly diurnal)

Clear convective

Cloudy (stratocumulus or cumulus)
- Importance of boundary layer clouds (Earth radiative budget)
- Small liquid water contents, difficult to measure

P. Bougeault, V. Masson, Processus dynamiques aux interfaces sol-atmosphere et ocean-atmosphere, cours ENM


