The atmospheric boundary layer: Where the atmosphere meets the Earth

Outline

- Definitions and description
- Quantifying turbulent exchange:
  - Reynolds decomposition
  - Closure problem
- Different ABL types
  - Neutral (Ekman / Friction) layer
  - Convective boundary layer
  - Stable boundary layer
- Surface energy balance

Characteristics

The atmospheric boundary layer (ABL) is where we spend most of our lives. In contrast to the free atmosphere, it experiences a pronounced daily cycle in nearly all meteorological variables. The ABL is defined as that part of the atmosphere that exchanges information with the surface on short time scales (less than 1 hour).
Close to the surface of the Earth, the atmospheric flow becomes turbulent and breaks up into eddies that effectively vertically transport momentum, heat, and moisture.

The main reasons for turbulence to develop in the ABL are 1. vertical wind shear (wind becomes zero at the surface) which makes the flow hydrodynamically unstable, as well as 2. negative buoyancy ($\frac{\partial}{\partial z} \theta < 0$) through heating of the Earth surface by absorption of solar radiation, which makes the air statically unstable.

These processes are referred to as mechanical production (shear production) of turbulence and buoyancy production/destruction of turbulence, respectively.

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A statically unstable situation occurs in a stratified fluid when a heavy fluid is lifted above a light fluid.

A convective situation develops as the heavier fluid moves into the lighter fluid beneath it.

Characteristics:
- All scales of motion grow without bound, i.e. there is no critical wavenumber that maximizes the growth rate of the instability.
- The growth rate increases with decreasing wavelength.
- Small disturbances quickly grow in amplitude and the flow progresses into large scale turbulence that does not effectively mix out the fluid.

A statically stable stratified fluid can be unstable with the onset of shear across the interface.

If the shearing stress is large enough, the most unstable wavelengths will grow (e.g. generation of Kelvin-Helmholtz waves).

If the shearing stress is not large enough, the stratification prevents disturbances from growing without bound.

The tradeoff between these two forces is quantified by the Richardson number. High Richardson numbers (above 0.25) imply strong stratification and weak susceptibility to shear instabilities, while low Richardson numbers (below 0.25) imply weak stratification and strong shear.

There is a dominant wavelength at which disturbance grow the fastest. This is characteristic of all pure shear instabilities. A disturbance with a small wavelength decays as a result of viscosity, while one that is too large requires too much shear to cause it to overturn. Hence only one wavelength dominates the flow after the instability sets in, and rather than being mixed effectively, only large scale stirring dominates the flow.
Mixed instability

In the above two examples, either a convective instability or a shear instability stirs and mixes the two fluids of differing densities. In either case, with enough time, the fluids in each container would have been completely mixed. But in each case, large scale turbulence dominated the flow field just after the onset of the instabilities.

Small scale turbulence decayed quickly, and mixing occurred through molecular diffusion across relatively weak gradients within the flow field stirred by the large scale turbulence.

Small scale turbulence is what more effectively mixes out two fluids of differing densities, since it increases the gradients by which molecular diffusion acts. But if only small scale turbulence exists, it is quickly overtaken by viscosity.

Hence an ideal situation that maximizes the mixing efficiency is one in which large scales persist throughout the flow field. The large scale motion feeds energy into small scales by lifting heavier fluid above lighter fluid and creating statically unstable regions which are susceptible to convective instabilities.

Buoyancy dominated mixed instability

The mixing efficiency can be increased by decreasing the Richardson number. In this situation mixing occurs more rapidly because small scales persist throughout the flow field. The large scale motion feeds energy into small scales by lifting heavier fluid above lighter fluid and creating statically unstable regions which are susceptible to convective instabilities.

If the shear instability is too strong, then the statically unstable regions do not last long enough for the convective instability to cause them to overturn, as was shown in shear dominated case.

If the time scale of the convective instability is on the order of or shorter than the shear instability, then the flow will be mixed more efficiently. If the time scale of the shear instability is too long, however, then the flow returns to the pure convective instability, which does not effectively mix the fluid.

Shear dominated mixed instability

Although the flow is statically unstable, convective instabilities do not develop because the large scale stirring resulting from the shear instability dominates the flow field.

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Reynolds decomposition

Reynolds decomposition embodies substituting this in the conservation equations and subsequently taking the time average of the resulting expressions

\[ x = x' + x', \ y = y' + y' \]

Some calculus rules:

\[ \Delta x = \Delta x' + \Delta x', \ y = x' + y' \]

Reynolds decomposition applied to nonlinear term generates additional covariance term: this term is nonzero when turbulent fluctuations in x (e.g. vertical wind speed) and y (e.g. potential temperature) vary together (so vary).

Quantifying turbulence: Reynolds decomposition

Turbulent transport can be quantified by looking at the contributions to time-varying signals with various frequencies:

(a) Mean alone

(b) Waves alone

(c) Turbulence alone

Reynolds decomposition:

\[ p = \nabla \cdot \vec{p}', \ u = \nabla \cdot \vec{u}', \ v = \nabla \cdot \vec{v}', \ w = \nabla \cdot \vec{w}', \ T = \nabla \cdot \vec{T}', \ q = \nabla \cdot \vec{q}' \]

Horizontal components, no tidal forces, neglecting friction

Substitute: \[ p = \nabla \cdot \vec{p}', \ u = \nabla \cdot \vec{u}', \ v = \nabla \cdot \vec{v}', \ w = \nabla \cdot \vec{w}' \]

AND average over time (again)
Continued

In most terms wind speed is simply replaced by its time-averaged value:

\[ \bar{u} = \left( u + u' \right) / 2 \]

In contrast, each nonlinear advection term generates an additional covariance term:

\[ \frac{\partial}{\partial x} \left( \frac{\partial u}{\partial x} \right) = \frac{\partial}{\partial x} \left( \frac{\partial u}{\partial x} \right) \bar{u} + \frac{\partial}{\partial x} \left( \frac{\partial u}{\partial x} \right) u' + \frac{\partial}{\partial x} \left( \frac{\partial u}{\partial x} \right) v' + \frac{\partial}{\partial x} \left( \frac{\partial u}{\partial x} \right) w' \]

The horizontal equations of motion for turbulent flow become:

\[ \frac{\partial u}{\partial t} + u \frac{\partial u}{\partial x} + v \frac{\partial u}{\partial y} + w \frac{\partial u}{\partial z} = -\frac{1}{\rho} \frac{\partial p}{\partial x} + \frac{u'}{\partial x} \frac{\partial u}{\partial x} + \frac{v'}{\partial y} \frac{\partial u}{\partial y} + \frac{w'}{\partial z} \frac{\partial u}{\partial z} \]

\[ \frac{\partial v}{\partial t} + u \frac{\partial v}{\partial x} + v \frac{\partial v}{\partial y} + w \frac{\partial v}{\partial z} = -\frac{1}{\rho} \frac{\partial p}{\partial y} + \frac{u'}{\partial x} \frac{\partial v}{\partial x} + \frac{v'}{\partial y} \frac{\partial v}{\partial y} + \frac{w'}{\partial z} \frac{\partial v}{\partial z} \]

\[ \frac{\partial w}{\partial t} + u \frac{\partial w}{\partial x} + v \frac{\partial w}{\partial y} + w \frac{\partial w}{\partial z} = -\frac{1}{\rho} \frac{\partial p}{\partial z} + \frac{u'}{\partial x} \frac{\partial w}{\partial x} + \frac{v'}{\partial y} \frac{\partial w}{\partial y} + \frac{w'}{\partial z} \frac{\partial w}{\partial z} \]

Continued

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Note: these still include advection of the mean flow.

But we need equations for \((u'w')\) and \((v'w')\) closure problem

Continued

Reynolds’ stress: interpretation

Eddy right side: \( w' < 0 \) and \( u' > 0 \) so \( u'w' < 0 \)

Eddy left side: \( w' < 0 \) and \( u' < 0 \) so \( u'w' < 0 \)

Potential temperature

Fictitious parcel

Start: \( T_p = T_s \)

Then quick displacement upward

Compare parcel temperature with temperature surroundings at new position.

\[ T_p > T_s \quad T_p = T_s \quad T_p < T_s \]

- parcel warmer (less dense) \( \rightarrow \) unstable stratification
- parcel at same temperature (and density) \( \rightarrow \) neutral stratification
- parcel colder (denser) \( \rightarrow \) stable stratification

But: when moving the parcel up it will change temperature!
The first law of thermodynamics:
\[ dQ = d(E) + dW \]

1. Heat added to the parcel: \( Q \)
2. Internal energy of the parcel: \( E \)
3. Work done by the parcel: \( W \)

Adiabatic process: \( dQ = 0 \)

Result: dry-adiabatic lapse rate \( \gamma_d \)

\[ \gamma_d = -\frac{g}{c_p} = -\frac{9.81 \text{ m/s}^2}{1005 \text{ J/kg} \cdot \text{K}} = -9.8 \text{ K/km} \]

Define: potential temperature:

\[ \theta = \frac{T}{p} \]

Continued

Static stability of the unsaturated (dry) atmosphere

Compare the (potential) temperature lapse rate of a fictitious air parcel:

\[ \gamma_p = 9.8 \text{ K/km} \]

with the observed atmospheric (potential) temperature lapse rate:

\[ \gamma = \frac{\partial \theta}{\partial z} \]

Result:

- \( \gamma_p < \gamma \) unstable stratification
- \( \gamma_p = \gamma \) neutral stratification
- \( \gamma_p > \gamma \) stable stratification

1) Convective boundary layer (CBL): buoyancy-driven heat and moisture exchange in statically unstable conditions, caused by a) absorption of solar radiation at surface or b) advection of cold air over warm surface (sea ice edge). Depth: up to 2000 m.

2) Stable boundary layer (SBL): shear-driven turbulent exchange in statically stable conditions, caused by a) longwave cooling at surface at night or b) advection of warm air over cold surface (warm sea cold continent). Depth: 100-500 m.

3) Neutral ABL over land ("Ekman layer"). Clouds and strong wind prevent strong daily cycle. Turbulent heat and moisture exchange at the surface is small. The Ekman ABL acts merely as a friction layer (exchange of momentum). Depth: typically 1000 m depending on wind speed.

4) Marine ABL: very important, because it determines atmosphere ocean coupling. Resembles Ekman layer, because mixing in the upper ocean layers transports absorbed solar energy away from the surface, so the daily cycle of surface temperature is small. Variable roughness (waves) and potentially strong advection effects because of long response time of water surface temperature and strong land-ocean temperature contrast.

Type of boundary layers

Basic temperature profiles

Potential temperature
Conserved for adiabatic (fast) vertical motion

Background
Start with a stable background potential temperature profile \( \theta(z) \)
Neutral boundary layer

A neutral, Ekman or friction BL is created by adiabatic mixing (strong winds, no additional energy source): upper ABL cools, lower ABL warms, but average ABL potential temperature remains unchanged. Because by definition ABL vertical motions must be fast, they are in good approximation adiabatic and the potential temperature is constant.

\[ \theta(z) \]

In the neutral ABL (d\(q\)/dz = 0) no exchange of heat takes place with the surface, only exchange of momentum, hence the term friction layer.

\[ \frac{Du}{Dt} = -\frac{1}{\rho} \frac{\partial p}{\partial x} + f v - \frac{\partial (u'w')}{\partial z} \]

\[ \frac{Dv}{Dt} = -\frac{1}{\rho} \frac{\partial p}{\partial y} - f u - \frac{\partial (v'w')}{\partial z} \]

Influence of friction on geostrophic flow along isobars is to deflect flow in the direction of the low pressure, which follows from a balance of forces in stationary flow.

Simple friction in the horizontal plane

Assume friction linearly dependent on wind speed: kW

\[ \frac{\partial (u'w')}{\partial z} = ku \]

Otherwise stationary flow (U/V=0) and pressure gradient in x-direction:

1. \( 0 = -\frac{1}{\rho} \frac{\partial p}{\partial x} + vu - ku \)
2. \( 0 = -fu - kv \)

\[ u = \frac{1}{k^2 + f^2} < 0 \]

\[ v = \frac{f}{k} \frac{1}{k^2 + f^2} < 0 \]
The Ekman spiral

\[ \text{Ekman spiral} \]

\[ \dot{u}_E = \frac{1}{\rho} \frac{\partial \mathbf{p}}{\partial x} + f v = \frac{1}{\rho} \frac{\partial (u'w')}{\partial z} \]

\[ \dot{v}_E = \frac{1}{\rho} \frac{\partial \mathbf{p}}{\partial x} - f u = \frac{1}{\rho} \frac{\partial (v'w')}{\partial z} \]

\[ \text{(See lecture notes)} \]

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Daytime mixed ABL: vigorous vertical mixing through strong surface heating, buoyant and mechanical production of turbulence. Maximum depth 1500 m, capping surface inversion.

Nighttime stable ABL: stable stratification through longwave cooling of the surface; turbulence generated by wind shear, destroyed by buoyancy. Slow growth. Poorly defined depth.

CBL temperature profile

**Connective Boundary Layer**

A CBL is created by heating the surface or transporting cold air over a warm surface (sea ice edge!). This heat is transferred to the near-surface air, which becomes statically absolutely unstable, so that parcels may rise unimpeded to ABL top until they meet potentially warmer air. The CBL has constant potential temperature except near the surface. Temperature inversion (potential temperature jump) sharply marks ABL depth. Large eddies with size up to ABL depth.

SBL temperature profile

**Stable Boundary Layer**

A SBL is created by cooling the surface. Heat is transferred from air to surface, creating a statically stable air layer in which turbulence is suppressed and can only be maintained by shear instability (strong wind). In absence of a clear inversion, the top of the SBL is often poorly defined. Sometimes a neutral residual layer is present above the SBL. Small eddies.

Example: clear sky daily q cycle

**Interpretation (Stable, night)**

Vertical profiles of turbulent sensible heat flux

Stable (blue) profiles:
- Eddy right side: \( u' > 0 \) and \( v' > 0 \) so \( w' < 0 \) and \( \dot{q} < 0 \)
- Eddy left side: \( u' < 0 \) and \( v' > 0 \) so \( w' > 0 \) and \( \dot{q} > 0 \)

Cooling
Interpretation (Convective, day)

Vertical profiles of turbulent sensible heat flux

Unstable (red) profiles:
Eddy right side: \( w' > 0 \) and \( q' > 0 \) so \( \left( \frac{w'}{w} \right) > 0 \)
Eddy left side: \( w' < 0 \) and \( q' < 0 \) so \( \left( \frac{w'}{w} \right) < 0 \)

Global and annual mean energy fluxes

The surface energy balance

Sensible and latent heat exchange at Earth’s surface

Surface energy balance

All energy fluxes through the surface should balance, unless energy is used for phase changes of the surface (like melt). This surface energy balance can be written as:

\[
M = SW \left( 1 - \alpha \right) + LW = LW_{net} + SHF + LHF + G
\]

where:
- \( M \) is the net surface heat flux
- \( SW \) is the shortwave flux
- \( LW \) is the longwave flux
- \( LW_{net} \) is the net longwave flux
- \( SHF \) is the sensible heat flux
- \( LHF \) is the latent heat flux
- \( G \) is the ground heat flux

Surface temperature is fully determined by the surface energy balance.

Sign convention:
Fluxes are defined positive when directed towards the surface

Surface short wave radiation balance

Net surface shortwave radiation \( SW_{net} = SW_{in} + SW_{out} \)

NH summer (W m\(^{-2}\)), June, July, August

NH winter (W m\(^{-2}\)), December, January, February

Surface long wave radiation balance

Net surface longwave radiation \( LW_{net} = LW_{in} + LW_{out} \)

NH summer (W m\(^{-2}\)), June, July, August

NH winter (W m\(^{-2}\)), December, January, February
**Surface net radiation balance**

Net surface radiation $R_{net} = S_{wnet} + L_{wnet}$

NH summer (W m$^{-2}$), June, July, August

NH winter (W m$^{-2}$), December, January, February

**Surface turbulent fluxes, latent heat**

$LHF = \rho C_v \left( q_h - q_s \right)$

NH summer (W m$^{-2}$), June, July, August

NH winter (W m$^{-2}$), December, January, February

**Surface turbulent fluxes, sensible heat**

$SHF = \rho C_P \left( \theta_a - \theta_s \right)$

NH summer (W m$^{-2}$), June, July August

NH winter (W m$^{-2}$), December, January, February

**To conclude**

$L_{win}$ $S$ $(1-\alpha)$ $L_{out}$ $G$