CHAPTER 4: ATMOSPHERIC TRANSPORT

Forces in the atmosphere:

- Gravity \( g \)
- Pressure-gradient \( \ddot{a}_p = -\left(\frac{1}{\rho}\right) \nabla P \)
- Coriolis \( \gamma_c = 2\omega v \sin \lambda \) to R of direction of motion (NH) or L (SH)
- Friction \( \ddot{a}_f = -kv \)

Equilibrium of forces:

In vertical: barometric law

In horizontal: geostrophic flow parallel to isobars

In horizontal, near surface: flow tilted to region of low pressure
CIRCULATIONS AROUND HIGHS AND LOWS

Northern hemisphere

Surface High = fine weather;
Surface Low = precipitation

CONVERGENCE  DIVERGENCE
THE HADLEY CIRCULATION (1735): global sea breeze

Explains:
• Intertropical Convergence Zone (ITCZ)
• Wet tropics, dry poles

Problem: does not account for Coriolis force. Meridional transport of air between Equator and poles would result in unstable longitudinal motion.
TROPICAL HADLEY CELL

- Easterly “trade winds” in the tropics at low altitudes
- Subtropical anticyclones at about 30° latitude

Fig. 4-11 Northern hemisphere Hadley cell.
CLIMATOLOGICAL SURFACE WINDS AND PRESSURES
(January)
CLIMATOLOGICAL SURFACE WINDS AND PRESSURES (July)
TIME SCALES FOR HORIZONTAL TRANSPORT (TROPOSPHERE)

- 2 weeks
- 1-2 months
- 1 year

(a) January
VERTICAL TRANSPORT: BUOYANCY

Balance of forces:

\[ \tilde{a}_{\text{buoyancy}} = \tilde{a}_p - g \]
\[ = \frac{\rho' - \rho}{\rho} g \]

Note: Barometric law assumed a neutrally buoyant atmosphere with \( T = T' \)

\( \left( \tilde{a}_p = -g \right) \)

\( T \neq T' \) would produce buoyant acceleration
Consider an air parcel at $z$ lifted to $z+dz$ and released. It cools upon lifting (expansion). Assuming lifting to be adiabatic, the cooling follows the adiabatic lapse rate $\Gamma$:

$$\Gamma = -\frac{dT}{dz} = \frac{g}{C_p} = 9.8 \text{ K km}^{-1}$$

What happens following release depends on the local lapse rate $-dT_{\text{ATM}}/dz$:

- $-dT_{\text{ATM}}/dz > \Gamma \Rightarrow$ upward buoyancy amplifies initial perturbation: atmosphere is **unstable**
- $-dT_{\text{ATM}}/dz = \Gamma \Rightarrow$ zero buoyancy does not alter perturbation: atmosphere is **neutral**
- $-dT_{\text{ATM}}/dz < \Gamma \Rightarrow$ downward buoyancy relaxes initial perturbation: atmosphere is **stable**
- $dT_{\text{ATM}}/dz > 0$ ("inversion"): very stable

The stability of the atmosphere against vertical mixing is solely determined by its lapse rate.
Fig. 4-19 Vertical profiles of temperature $T$, potential temperature $\theta$, water vapor (dew point), and ozone measured by aircraft in early afternoon in August over eastern Canada.
WHAT DETERMINES THE LAPSE RATE OF THE ATMOSPHERE?

- An atmosphere left to evolve adiabatically from an initial state would eventually tend to neutral conditions (-dT/dz = \( \Gamma \)) at equilibrium.
- Solar heating of surface and radiative cooling from the atmosphere disrupts that equilibrium and produces an unstable atmosphere:

\[
\begin{align*}
\text{Initial equilibrium state: } & - \frac{dT}{dz} = \Gamma \\
\text{Solar heating of surface/radiative cooling of air: } & \text{unstable atmosphere} \\
\text{buoyant motions relax unstable atmosphere back towards } & -\frac{dT}{dz} = \Gamma
\end{align*}
\]

- Fast vertical mixing in an unstable atmosphere maintains the lapse rate to \( \Gamma \).
Observation of \(-\frac{dT}{dz} = \Gamma\) is sure indicator of an unstable atmosphere.
IN CLOUDY AIR PARCEL, HEAT RELEASE FROM $\text{H}_2\text{O}$ CONDENSATION MODIFIES $\Gamma$

Wet adiabatic lapse rate $\Gamma_w = 2-7 \text{ K km}^{-1}$

“Latent” heat release as $\text{H}_2\text{O}$ condenses

$\Gamma = 9.8 \text{ K km}^{-1}$

RH $> 100\%$: Cloud forms

RH $100\%$
VERTICAL PROFILE OF TEMPERATURE
Mean values for 30°N, March

Radiative cooling (ch.7)

- 3 K km⁻¹

- 6.5 K km⁻¹

Radiative heating:
\[ \text{O}_3 + h\nu \Rightarrow \text{O}_2 + \text{O} \]
\[ \text{O} + \text{O}_2 + \text{M} \Rightarrow \text{O}_3 + \text{M} \]

Latent heat release

Surface heating
DIURNAL CYCLE OF SURFACE HEATING/COOLING: ventilation of urban pollution
Fig. 4-17 Formation of a subsidence inversion. Temperature profiles on the right panel are shown for the upwelling region $A$ (thin line) and the subsiding region $B$ (bold line). It is assumed for purposes of this illustration that regions $A$ and $B$ have the same surface temperature $T_0$. The air column extending up to the subsidence inversion is commonly called the planetary boundary layer (PBL).
FRONTS

WARM FRONT:
- Warm Air
- Cold Air
- Wind
- Front boundary; inversion

COLD FRONT:
- Cold Air
- Warm Air
- Wind
- Inversion
TYPICAL TIME SCALES FOR VERTICAL MIXING

- Estimate time $\Delta t$ to travel $\Delta z$ by turbulent diffusion:

$$\Delta t = \frac{(\Delta z)^2}{2K_z}$$

with $K_z \sim 10^5 \text{cm}^2\text{s}^{-1}$