# AATM 505

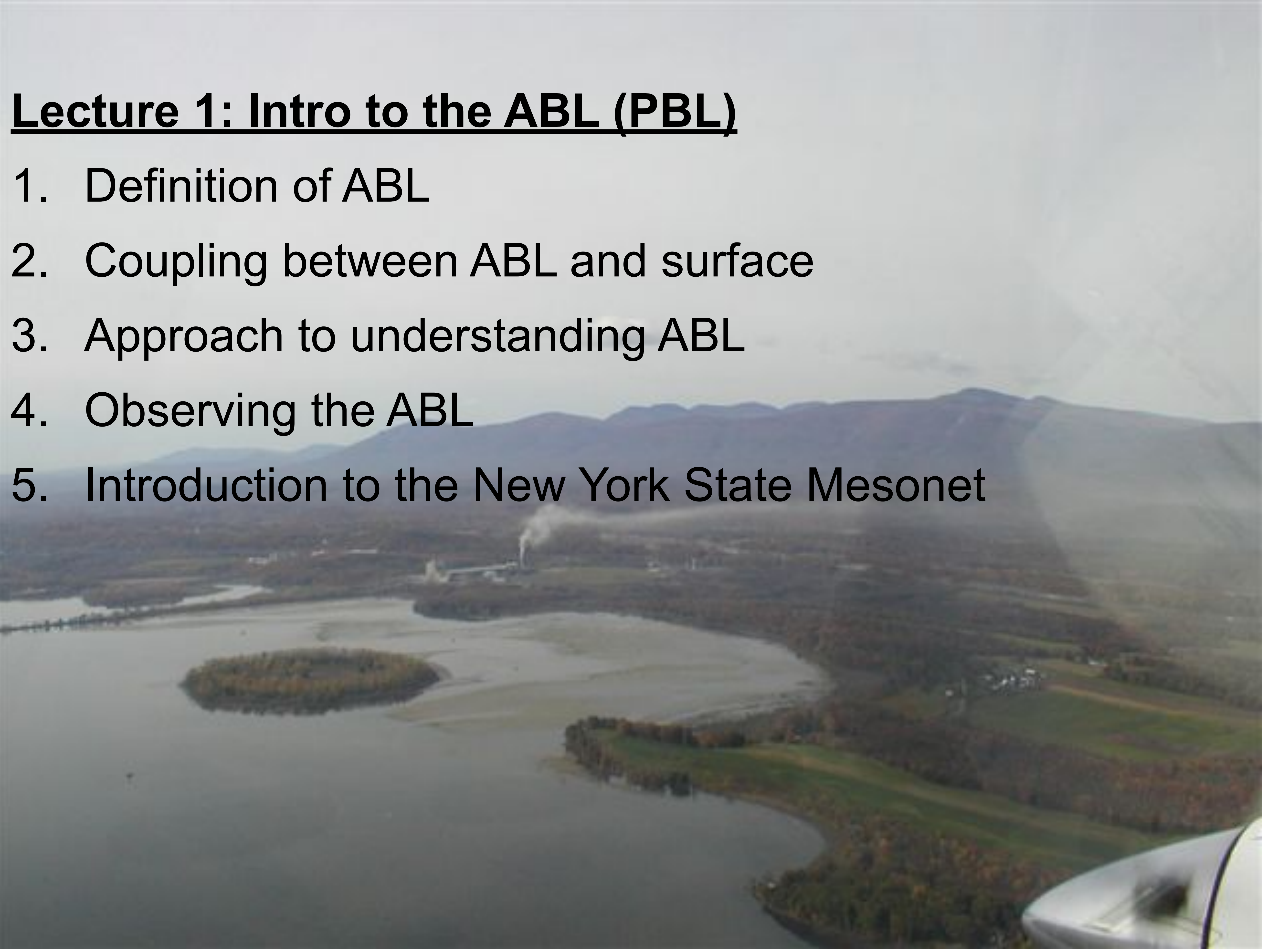
## Boundary Layer Module Review

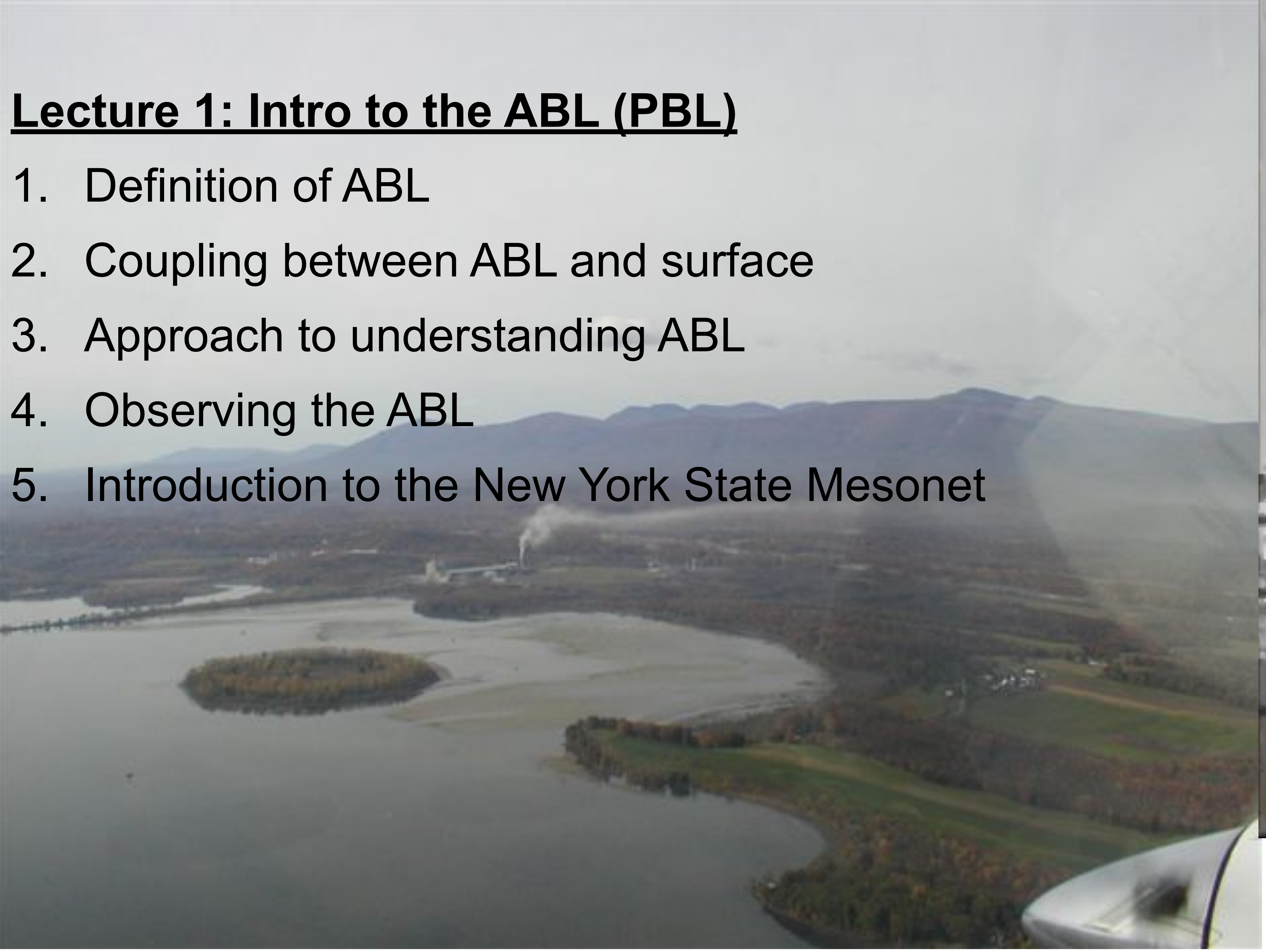
# Exam Monday (3/6/2023)!

(Modules 1 and 2)

**Lecture 1: Intro to the ABL (PBL)**

1. Definition of ABL
2. Coupling between ABL and surface
3. Approach to understanding ABL
4. Observing the ABL
5. Introduction to the New York State Mesonet





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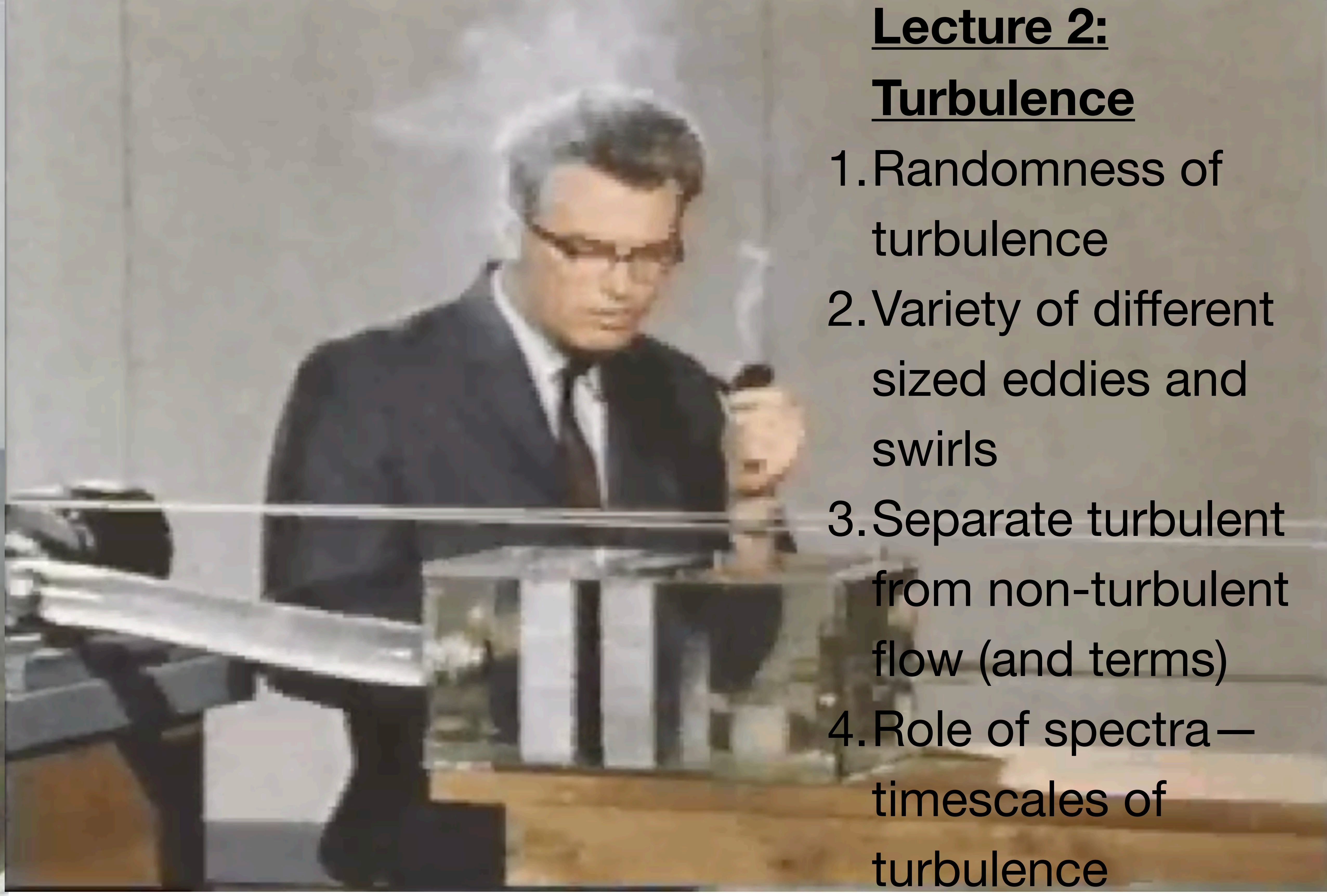
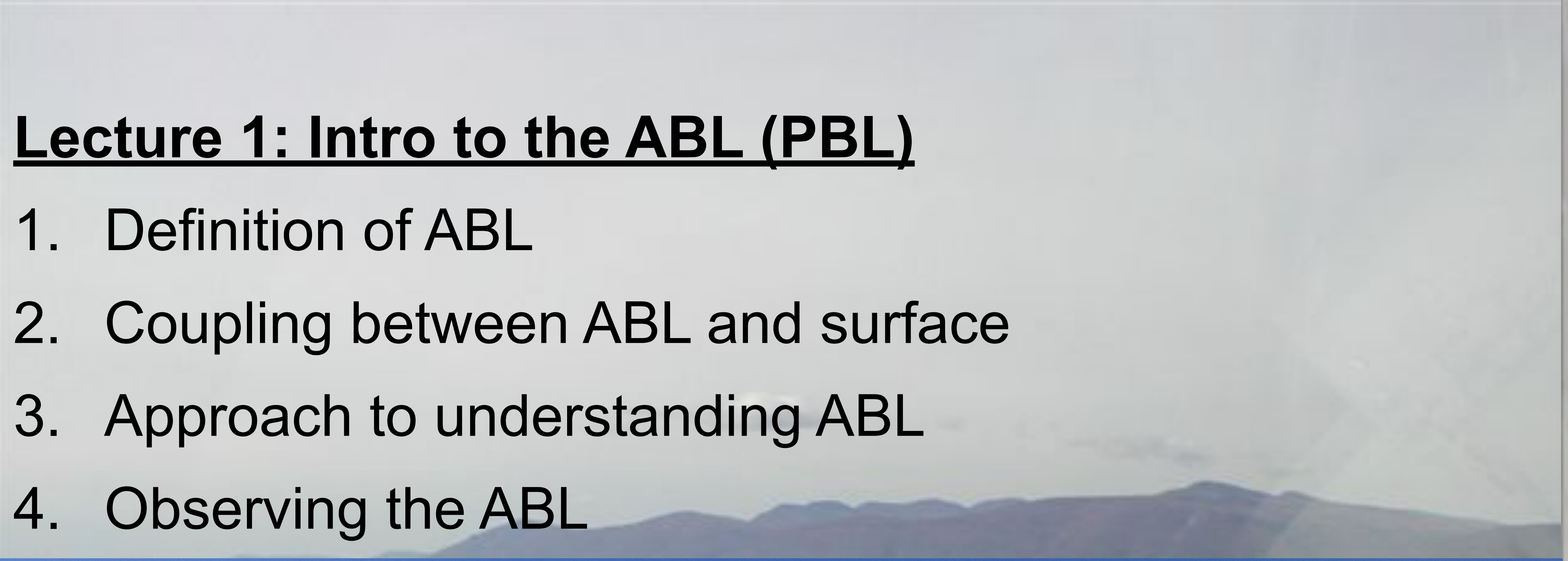


**Lecture 2: Turbulence**

1. Randomness of turbulence
2. Variety of different sized eddies and swirls
3. Separate turbulent from non-turbulent flow (and terms)
4. Role of spectra—timescales of turbulence



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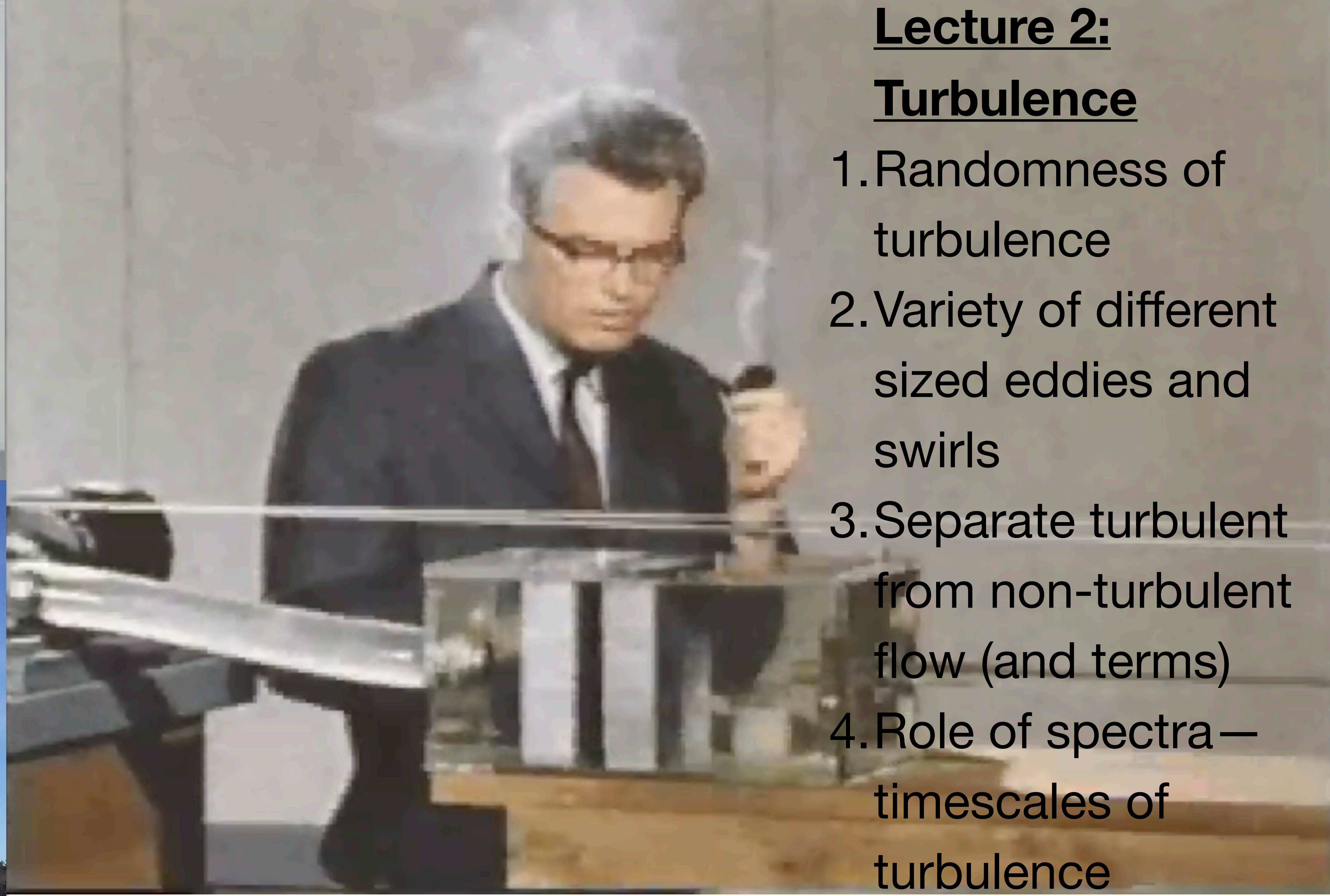

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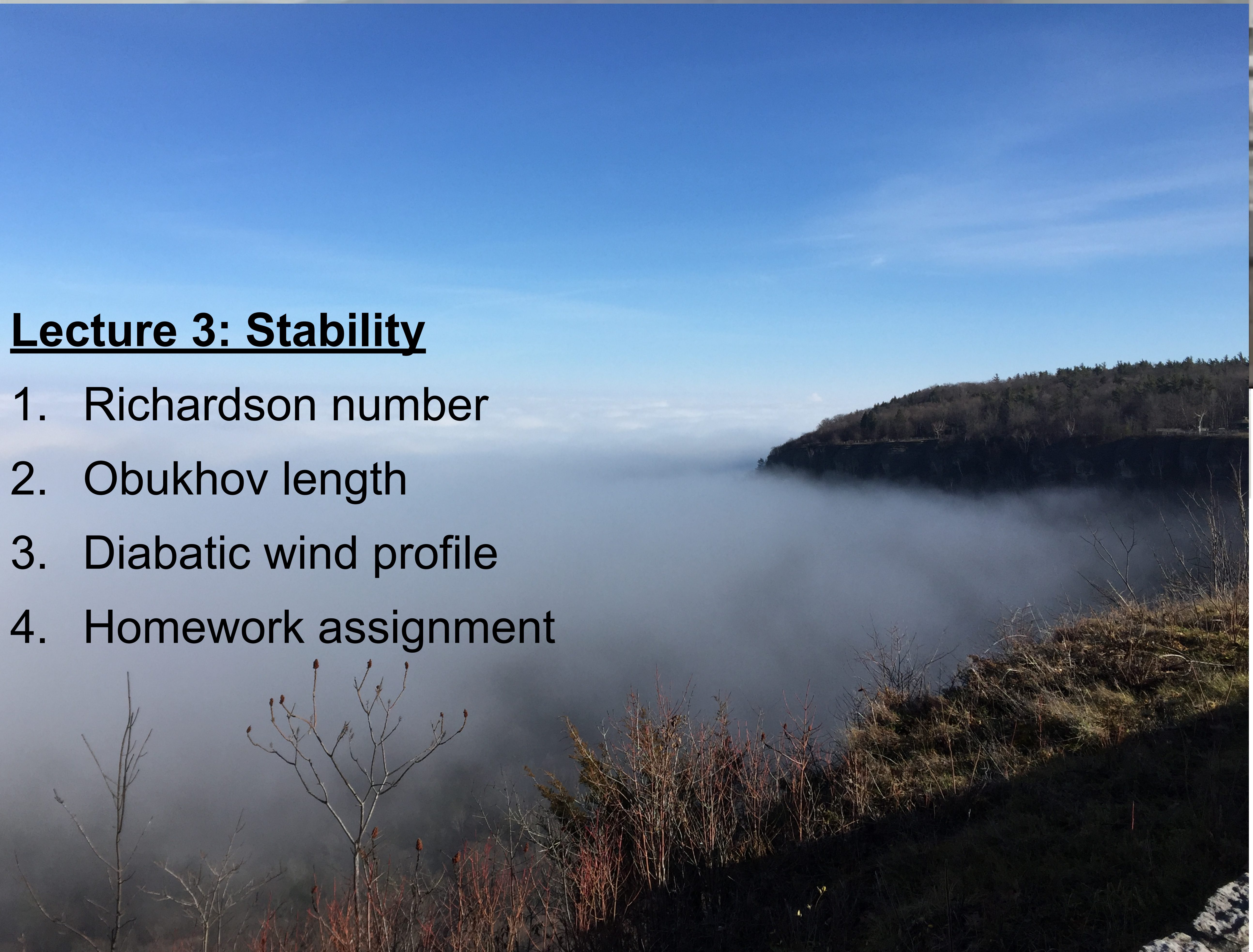

**Lecture 3: Stability**

1. Richardson number
2. Obukhov length
3. Diabatic wind profile
4. Homework assignment

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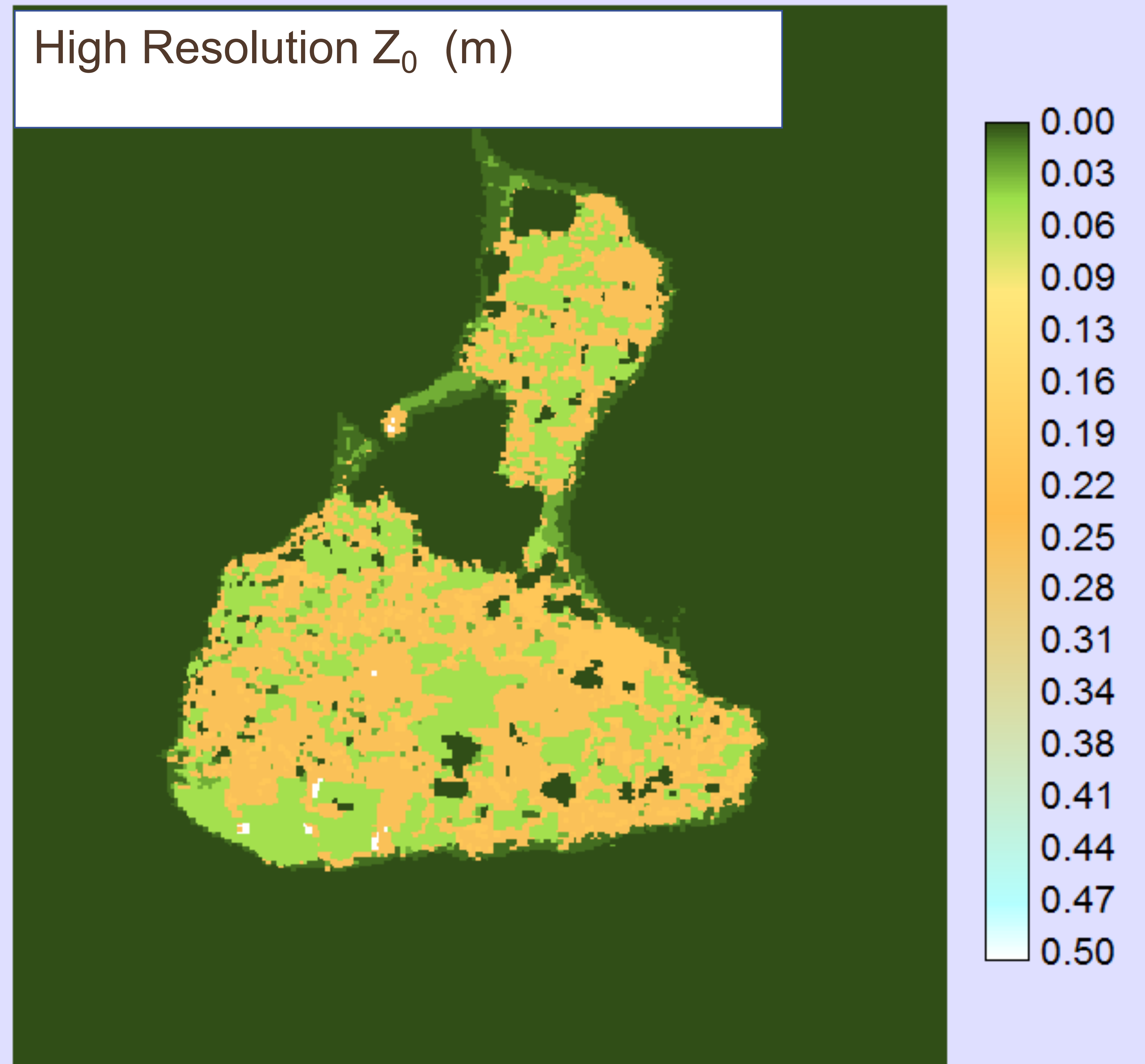


**Lecture 4: surface parameterizations**

1. bulk aerodynamic formulas
2. NYS Mesonet—wind profiles


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NYS Mesonet—wind profiles
Lecture 5: Field measurements (offshore wind | NYSM)
1. NYSM measurements—land-atmosphere exchange
2. Marine Atmospheric Boundary Layer—offshore wind

# Diurnal cycle of the Atmospheric Boundary Layer (ABL)

Dealing mostly with the fair weather (“boring”, quiescent) atmospheric boundary layer—which happens ⅔ of the time around here over land [but we’ll touch upon other boundary layers—over the ocean].

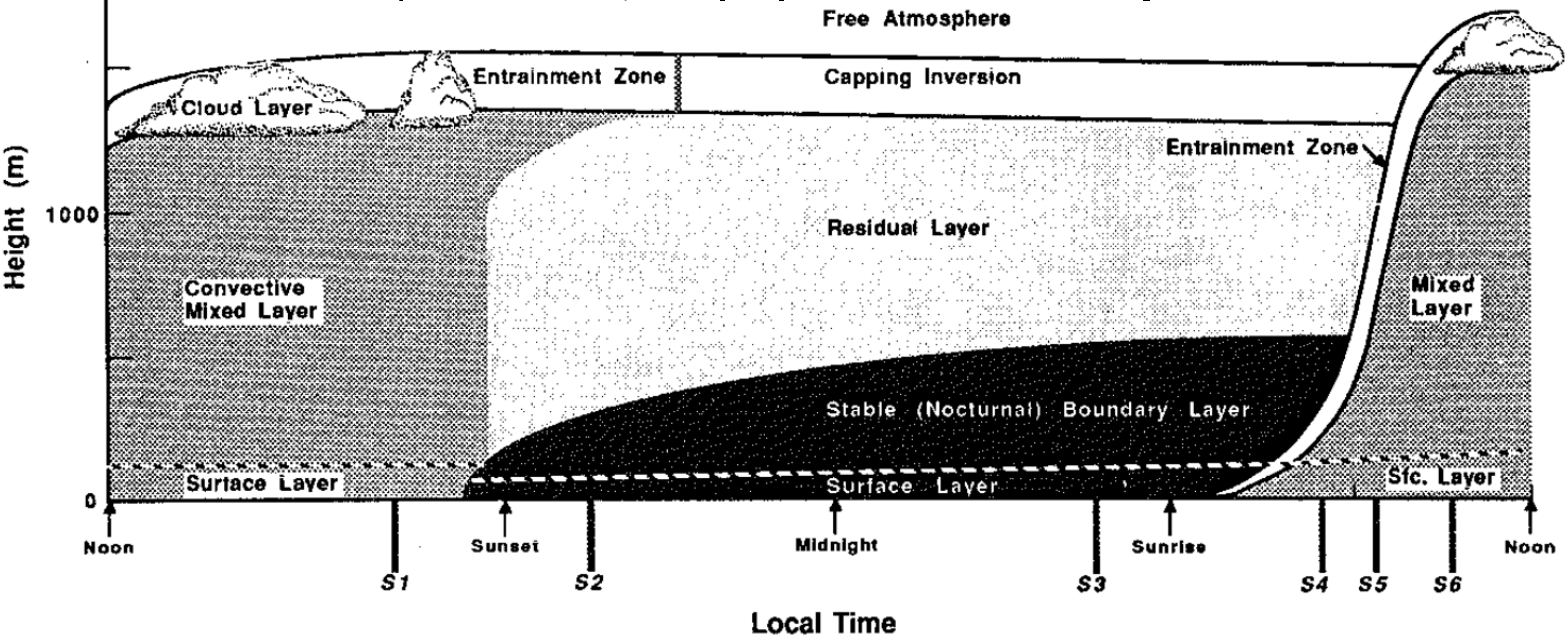
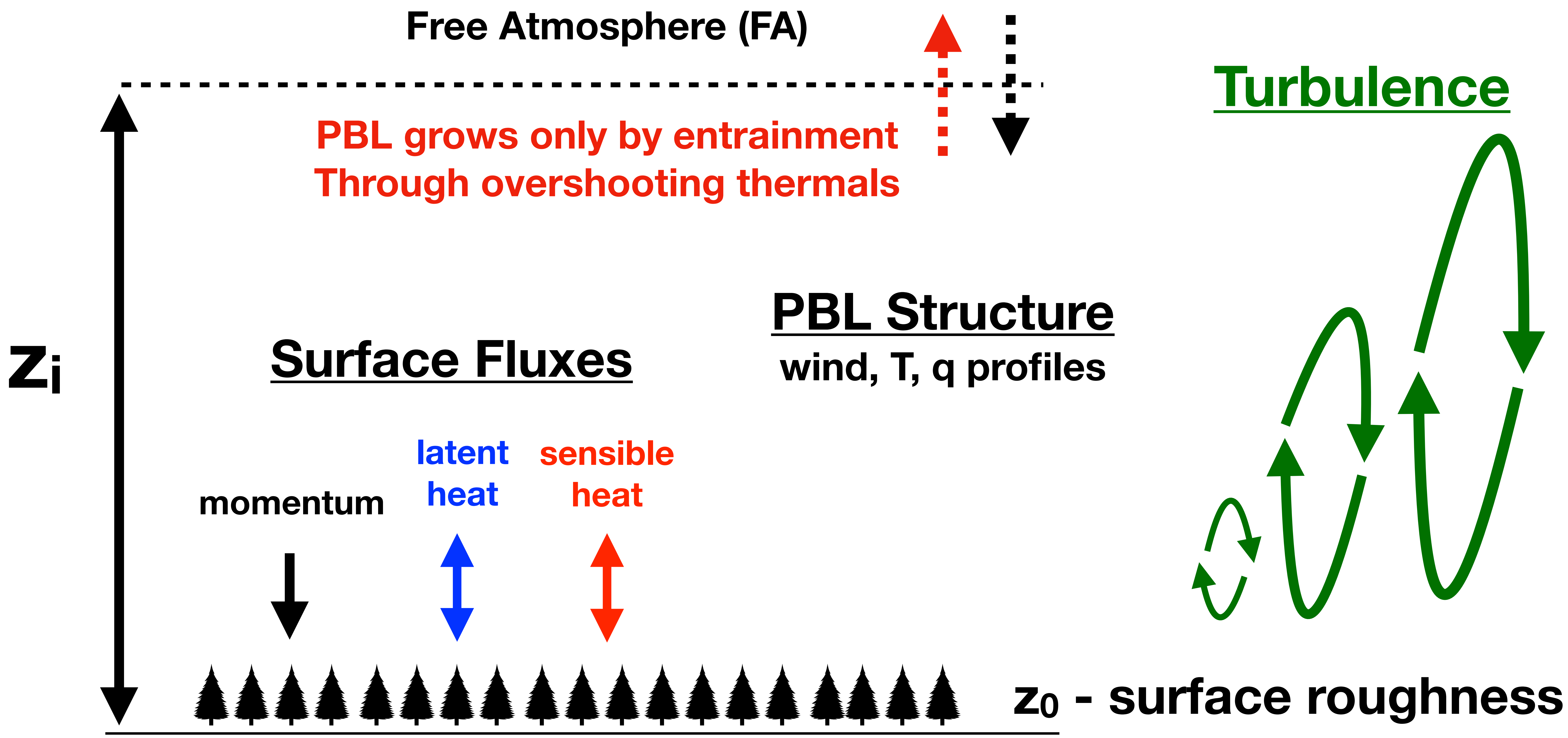



Figure from Still (1988) Chapter 1

# Conceptual Model of the ABL (PBL)




horizontally homogeneous

# Taylors Frozen Flow Hypothesis

$$time \Leftrightarrow space$$

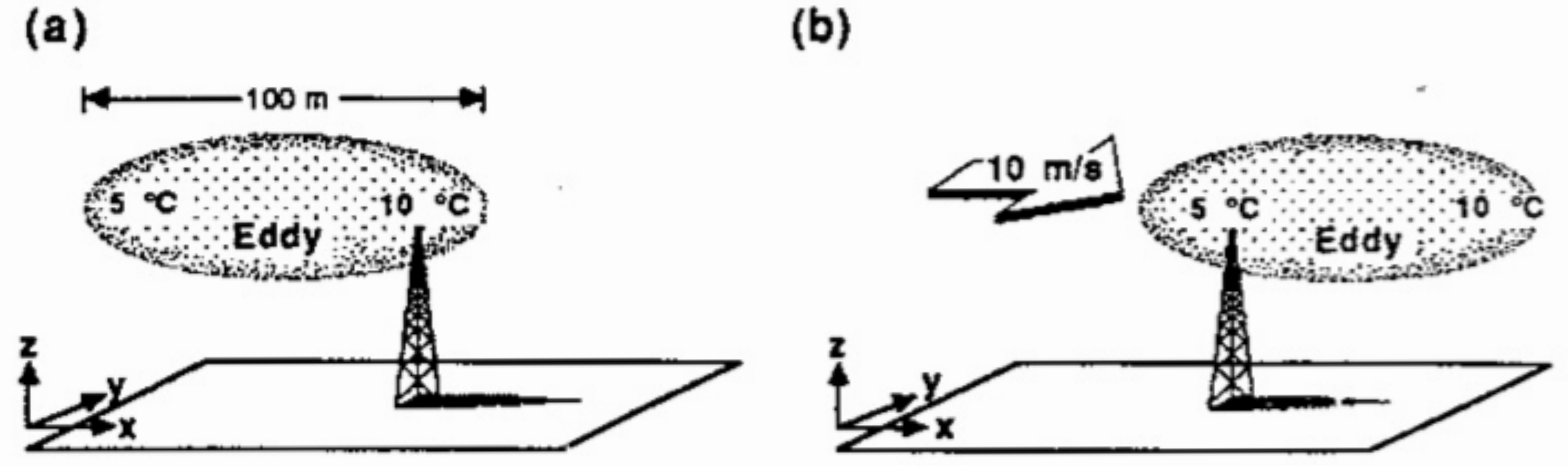



**Fig. 1.4** Illustration of Taylor's hypothesis. (a) An eddy that is 100 m in diameter has a 5 ° C temperature difference across it. (b) The same eddy 10 seconds later is blown downwind at a wind speed of 10 m/s.

$$\frac{d}{dt} = 0 \longrightarrow \frac{\partial}{\partial t} + U\frac{\partial}{\partial x} = 0 \longrightarrow \frac{\partial}{\partial x} = -\frac{1}{U}\frac{\partial}{\partial t}$$

$$\frac{\Delta T}{\Delta x} = -\frac{1}{U}\frac{\Delta T}{\Delta t} = -\frac{1}{10ms^{-1}} \times \frac{-5C}{10s} = 0.05^o Cm^{-1}$$

Stull (1988)

# Spectrum of Wind Speed

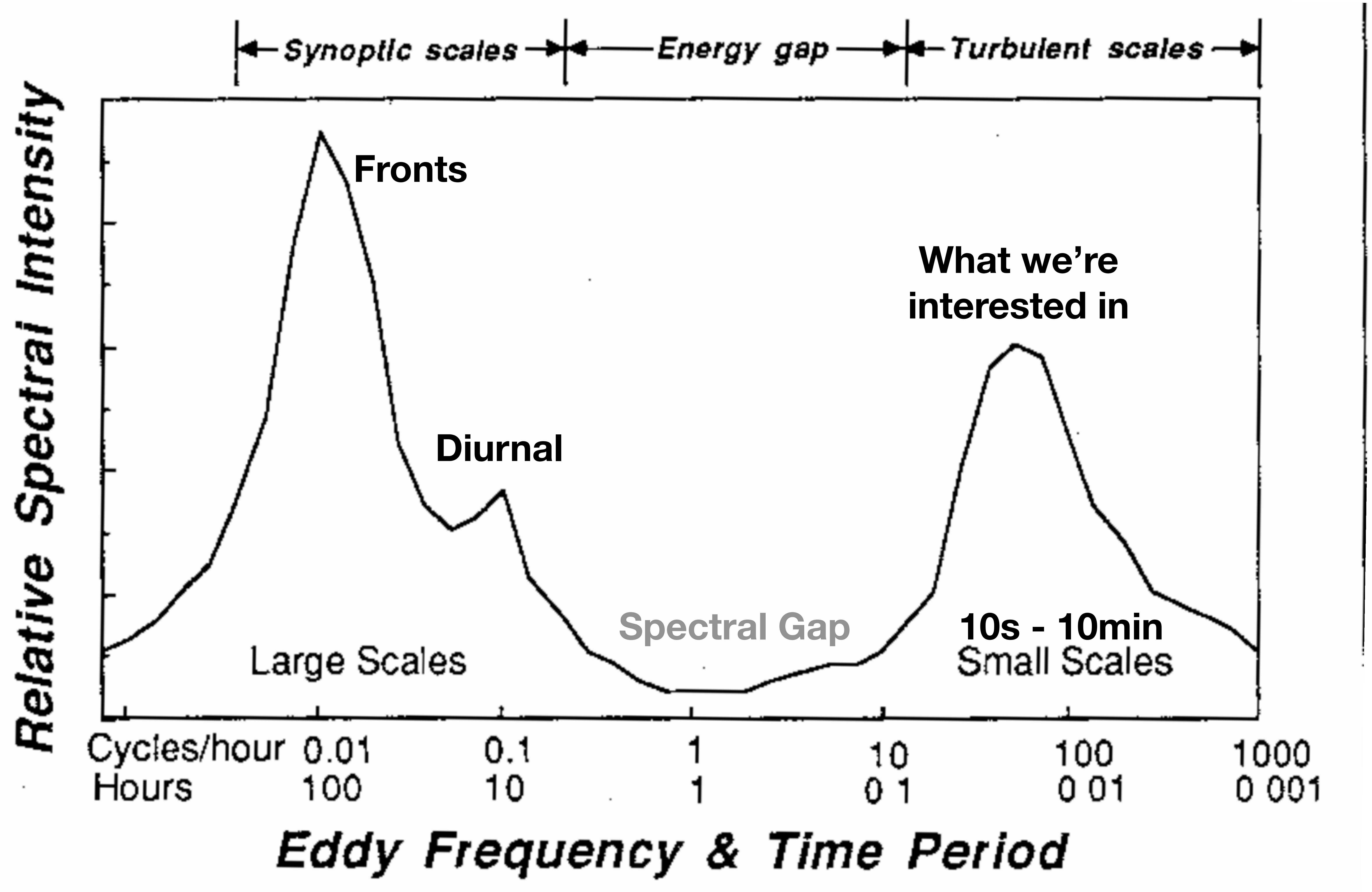



**Fig. 2.2** Schematic spectrum of wind speed near the ground estimated from a study of Van der Hoven (1957)

# scalar flux

$$\frac{J}{m^2 \cdot s} = \frac{W}{m^2}$$

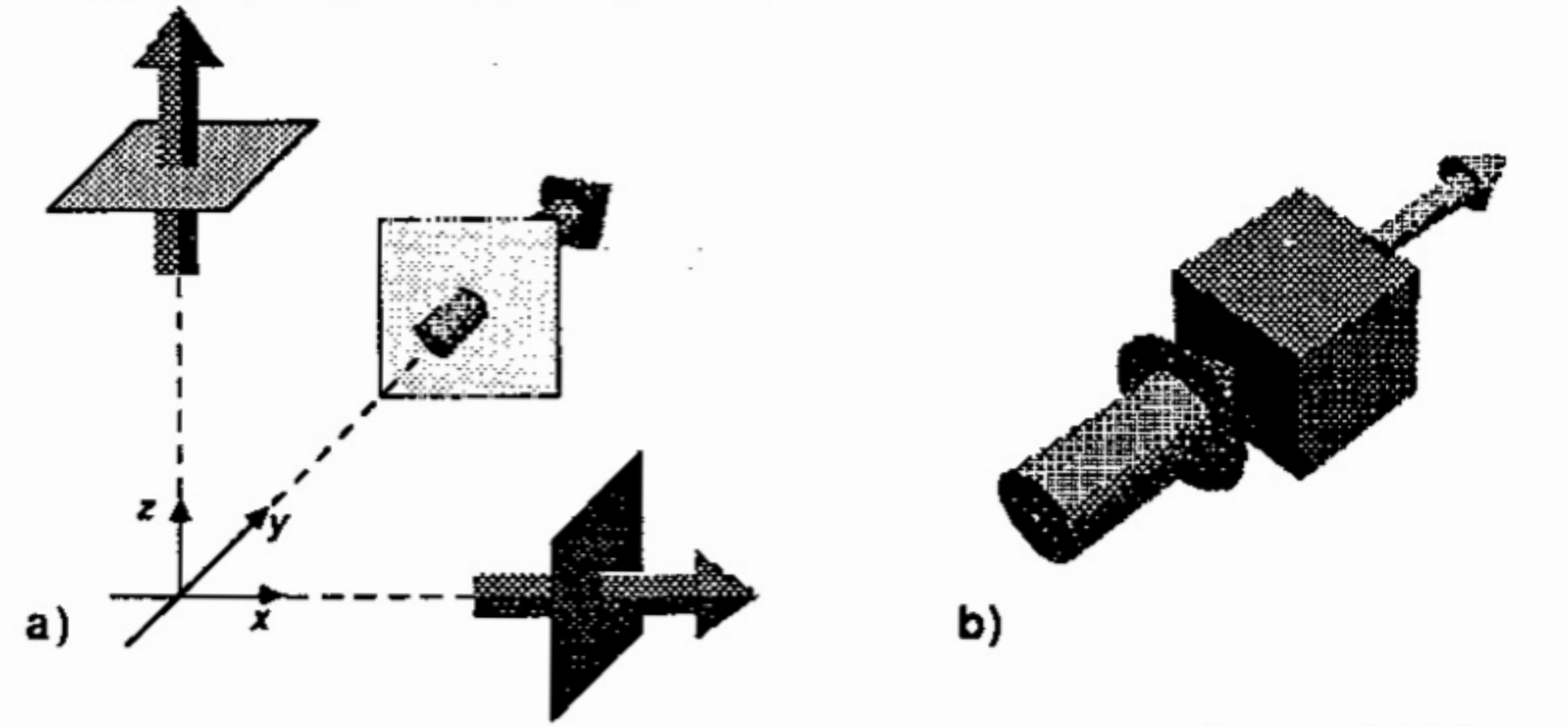



**Fig. 2.10** Flux is the rate of transfer of a quantity across a unit area. (a) The net flux can be split into components in the x,y, and z directions, as shown here. (b) If a greater flux enters a volume than leaves, then there must be a net increase of that quantity within the volume.

Stull (1988)

# Turbulent heat flux

$$H_s = \rho_a C_p \overline{w'T'}$$

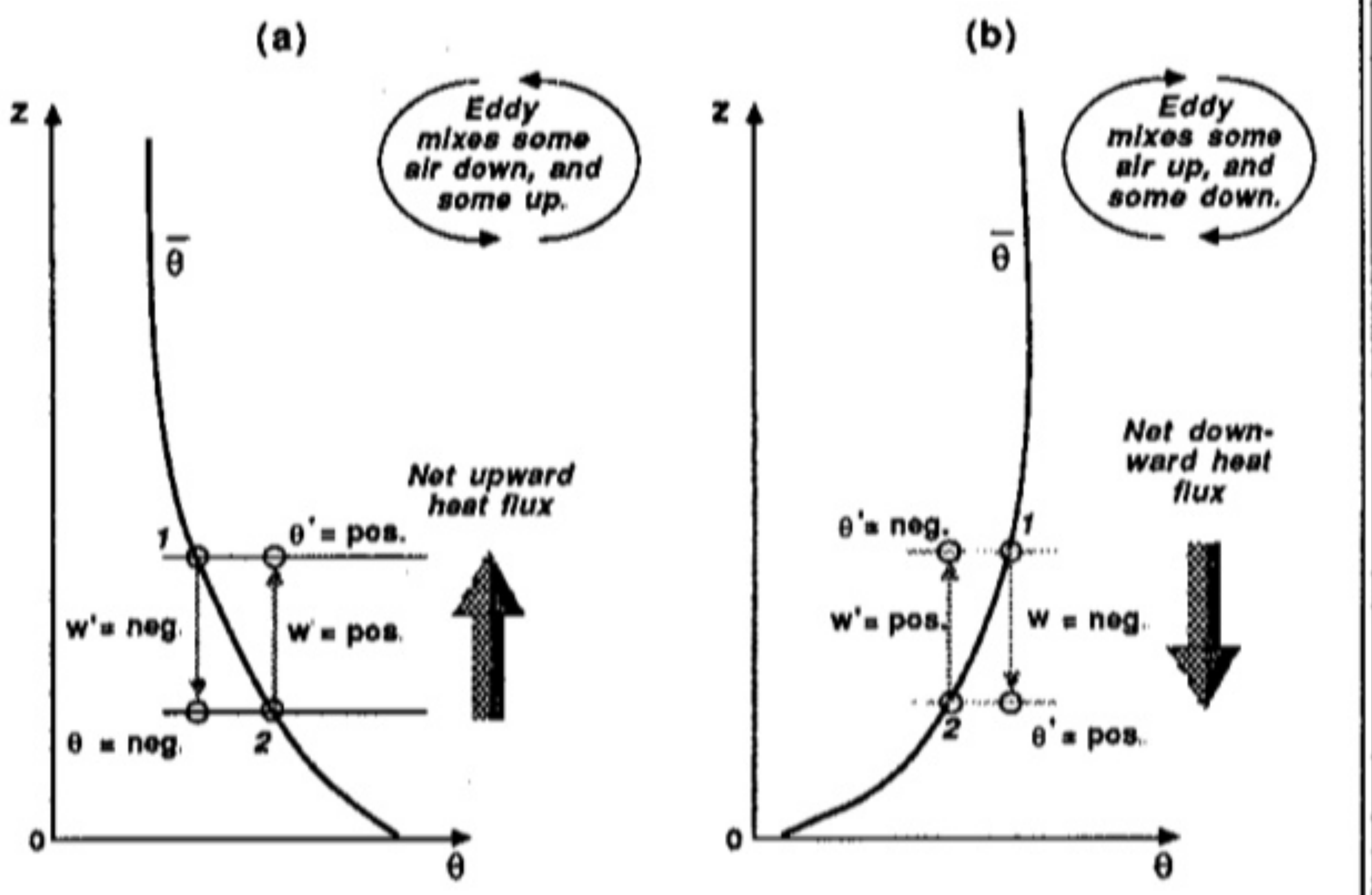
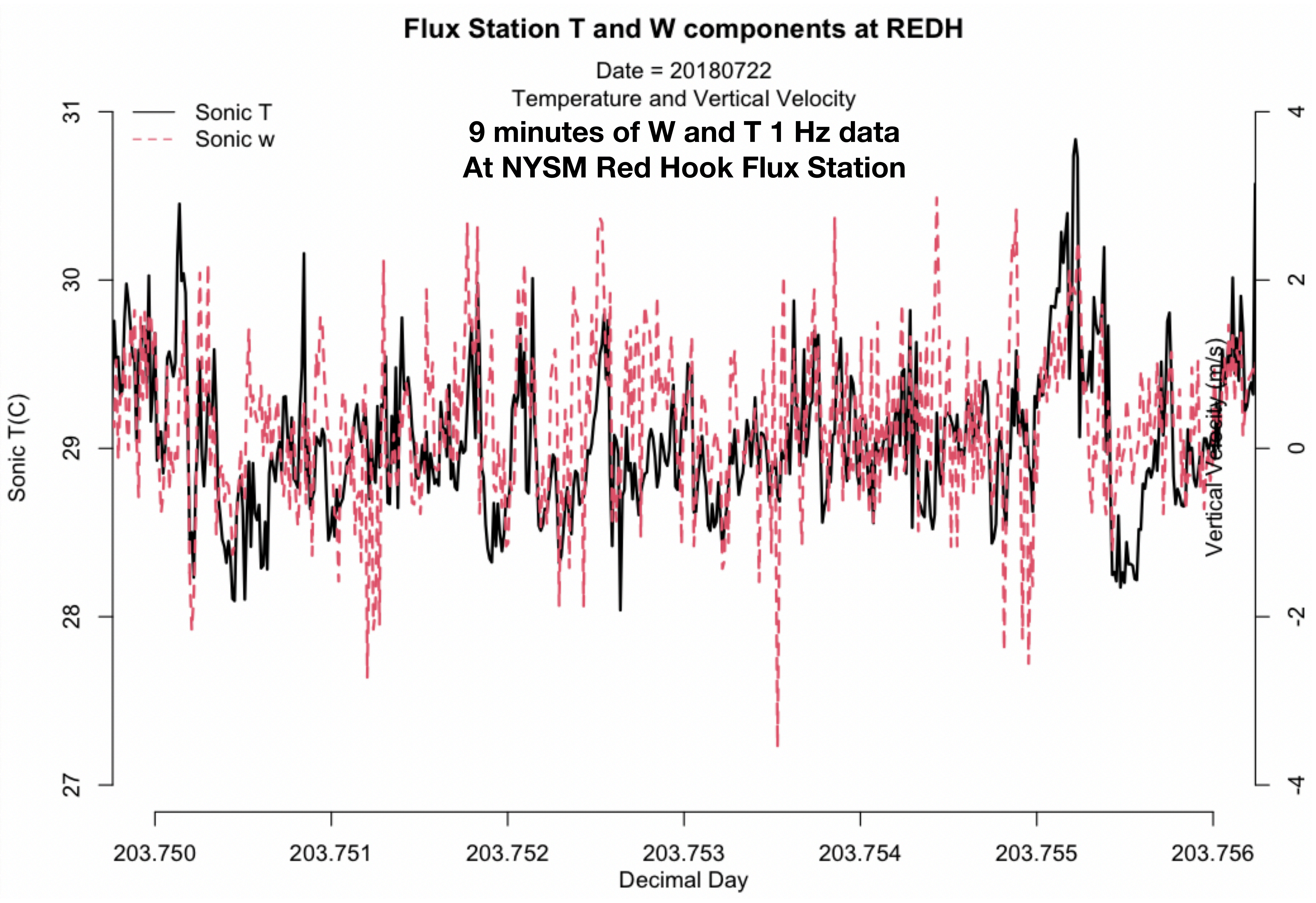


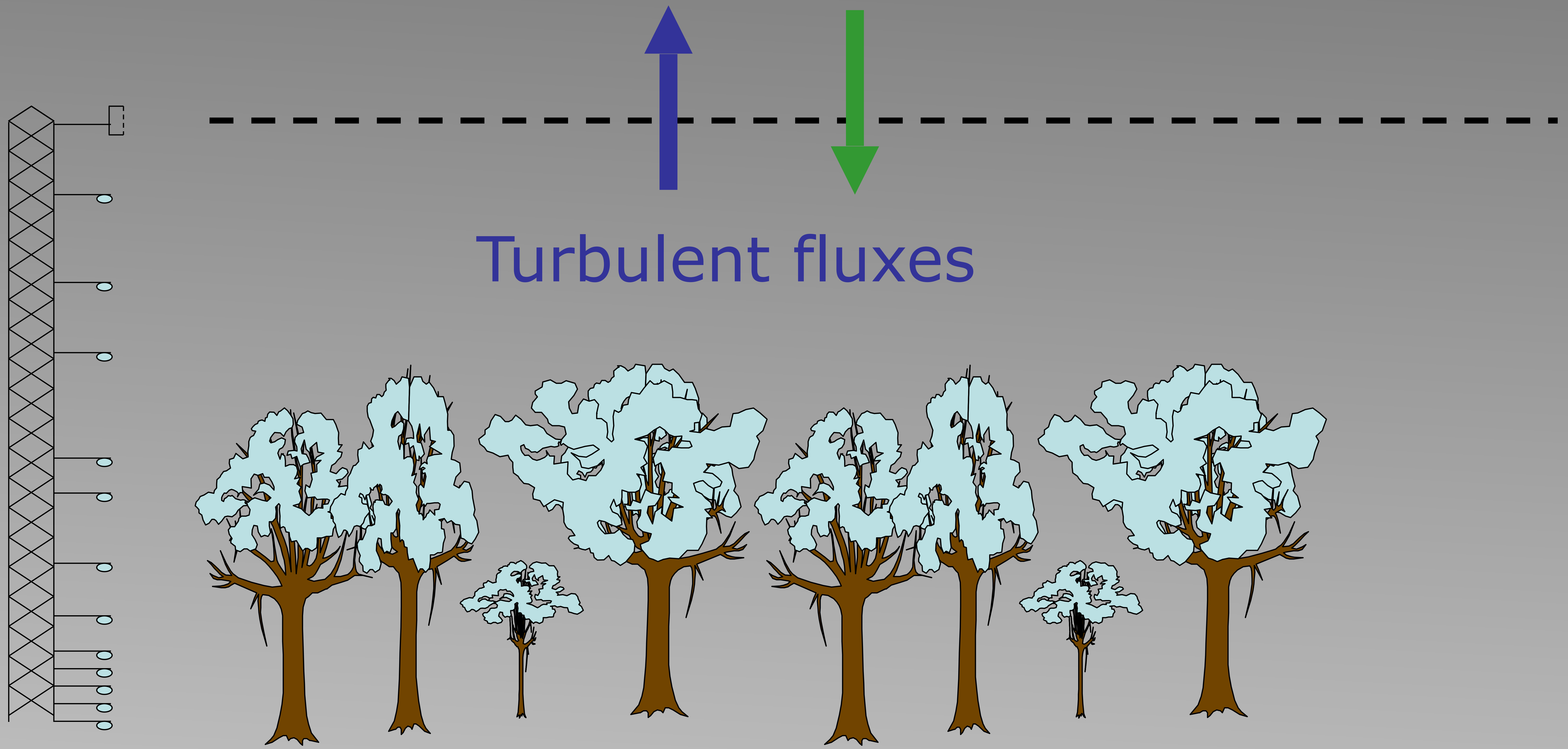

**Fig. 2.12** Idealization of the small eddy mixing process, showing (a) net upward turbulent heat flux in a statically unstable environment, and (b) net downward turbulent heat flux in a stable environment.

Stull (1988)

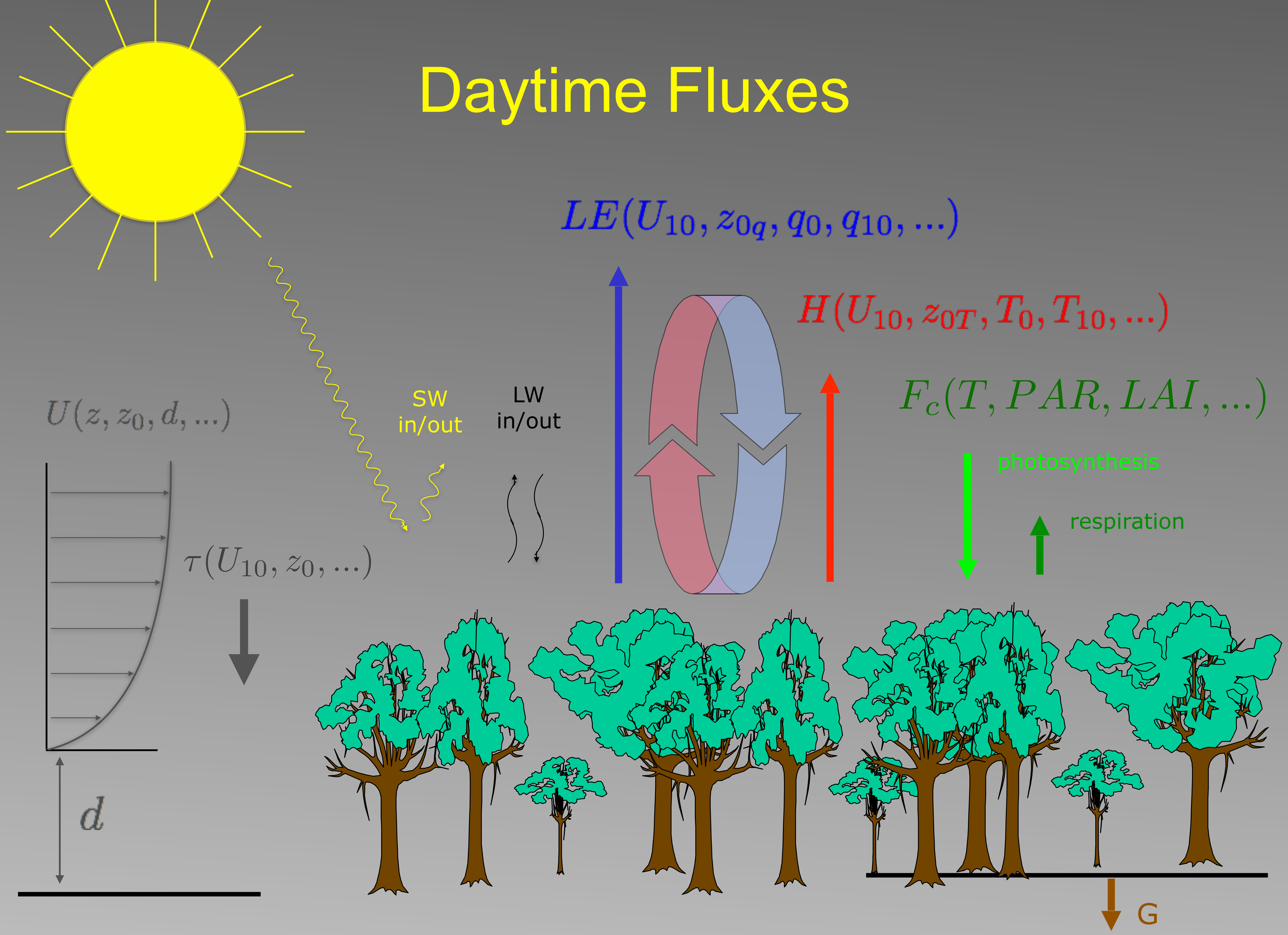

Flux Station T and W components at REDH
Date = 20180722
Temperature and Vertical Velocity
9 minutes of W and T 1 Hz data
At NYSM Red Hook Flux Station
Sonic T
Sonic w
Sonic T(C)
31
30
29
28
27
Vertical Velocity (m/s)
4
2
0
-2
-4
203.750
203.751
203.752
203.753
203.754
203.755
203.756
Decimal Day

# Measurement Approach

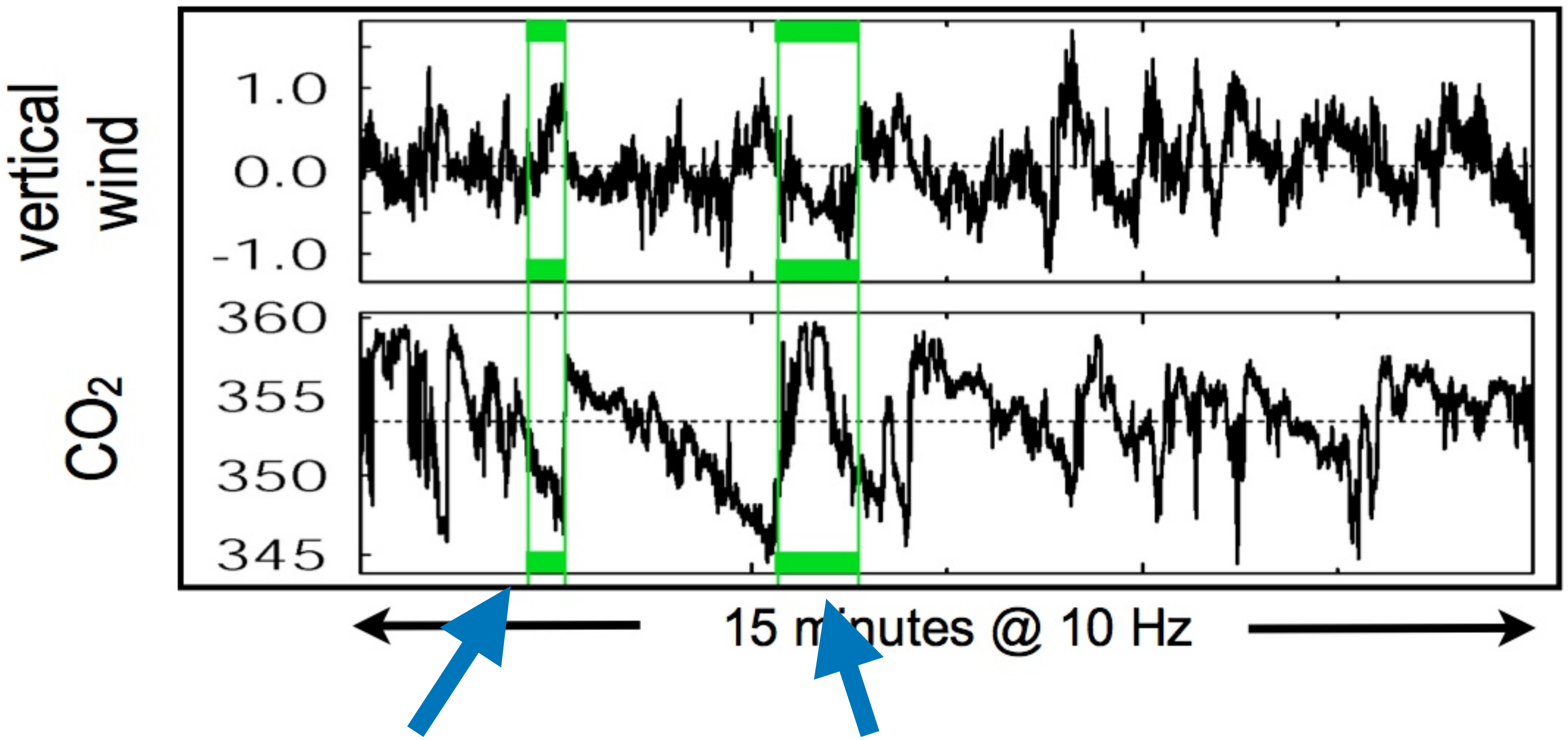
Control volume = atmospheric air below sensors



# Daytime Fluxes



# Example of $CO_2$ exchange above a forest




| **updraft** | **downdraft** | |
|---|---|---|
| $w' > 0$ | $w' < 0$ | |
| $c' < 0$ | $c' > 0$ | |
| $w'c' < 0$ | $w'c' < 0$ | → $\overline{w'c'} < 0$ |
| | | forest absorbing carbon |

# Law of the Wall, aka Logarithmic Wind Profile

$$\tau = \rho K_m \frac{d\bar{u}}{dz}$$

$$\rho u_*^2 = \rho \kappa u_* z \frac{d\bar{u}}{dz}$$

$$\frac{d\bar{u}}{dz} = \frac{u_*}{\kappa z}$$

$$\frac{d\bar{u}}{d\ln z} = \frac{u_*}{\kappa}$$

$$\bar{u}(z) = \frac{u_*}{\kappa} \ln z + C$$ ← const

Define $C$ so that $\bar{u} = 0$ when $z = z_0$ (roughness length). Recall no-slip boundary condition

$$\boxed{\bar{u}(z) = \frac{u_*}{\kappa} \ln\left(\frac{z}{z_0}\right)}$$

law of the wall

# Static Stability

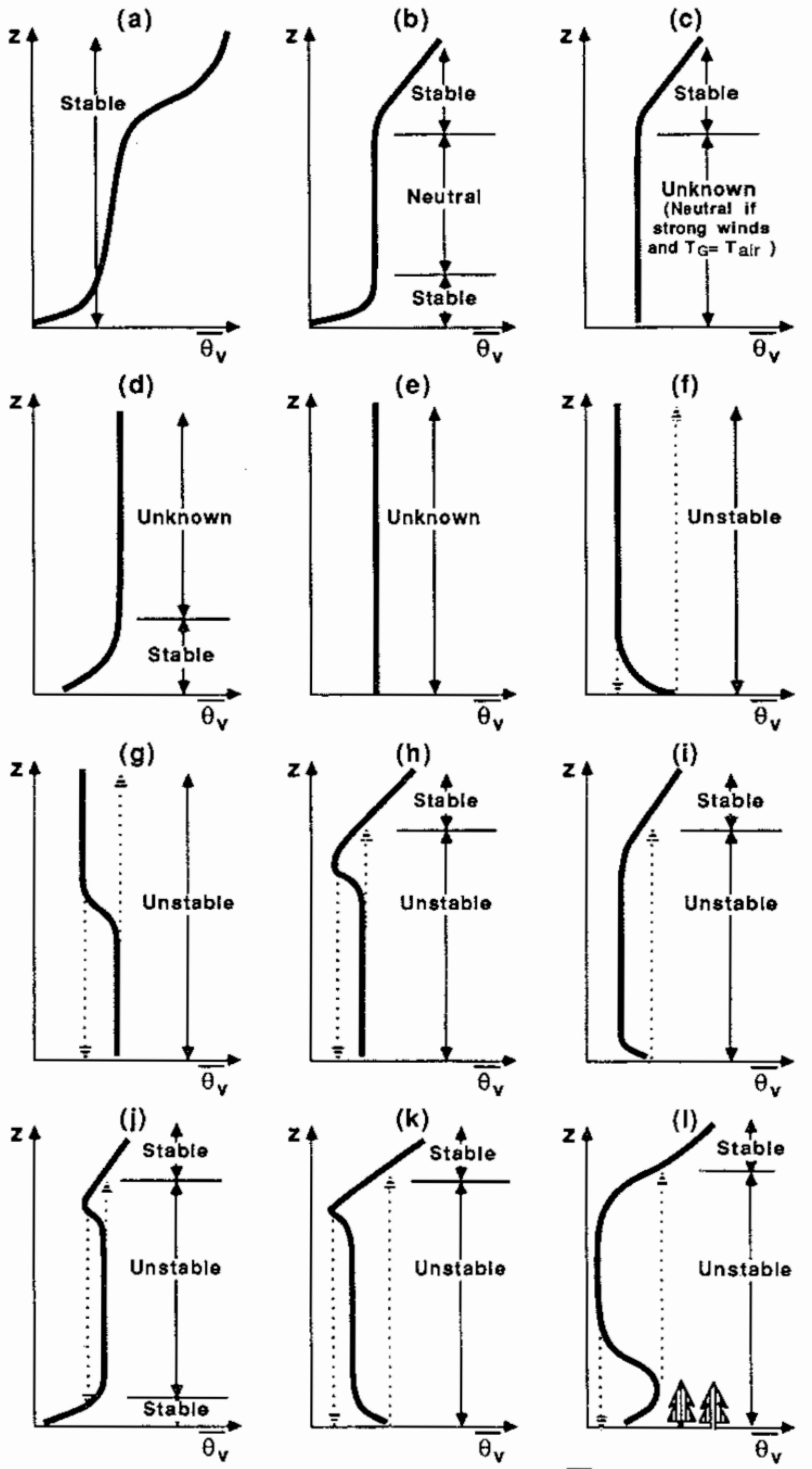
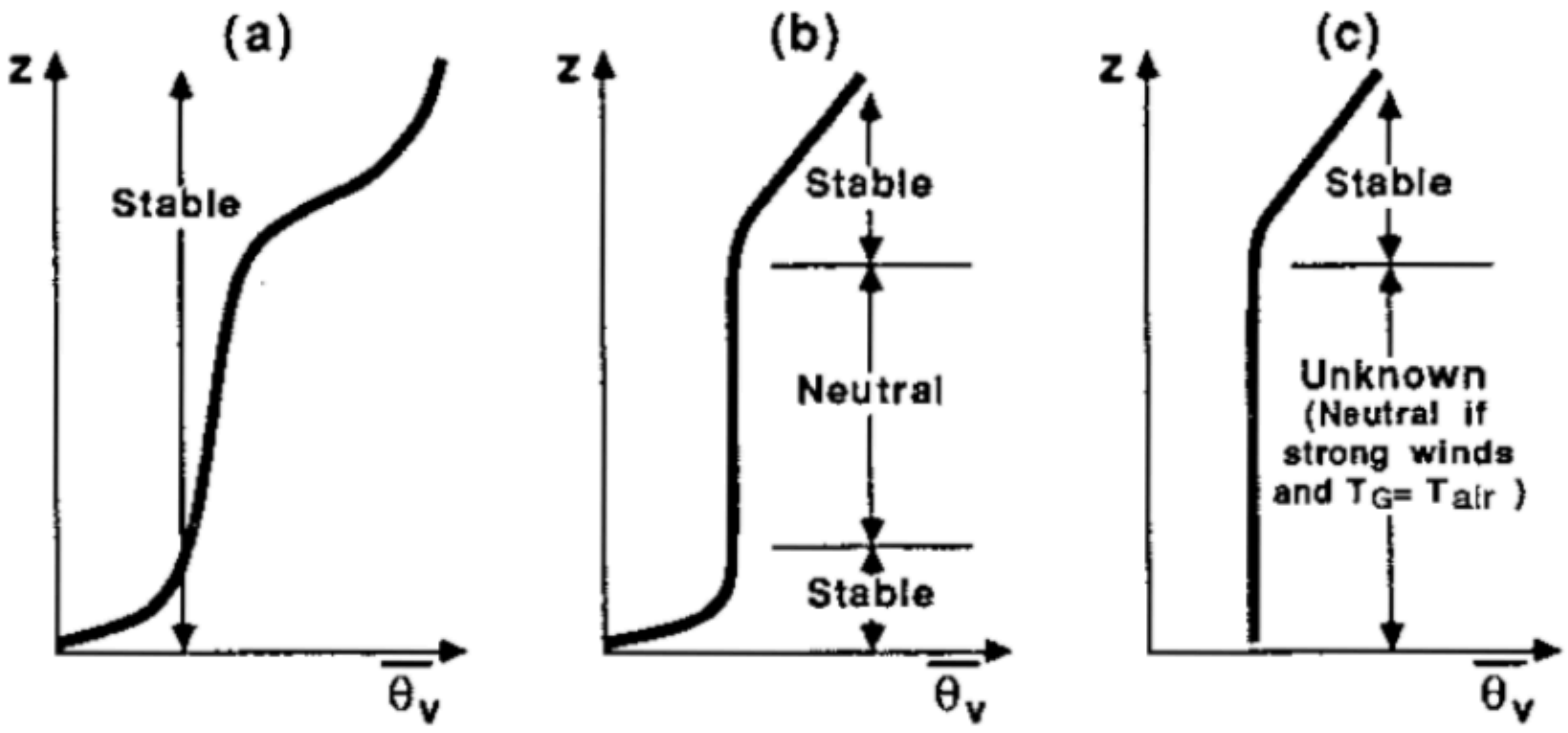


Fig. 5.17 Static stabilitiy as a function of the $\overline{\theta_v}$ profile
Dotted lines denote parcel movement.

- measure of capability for convection
- considers buoyancy only (does not consider wind/mechanical turbulence)
- local lapse rate (stability) insufficient - need the look at the whole profile *or* measure the buoyancy flux

Stull 1988

# Static Stability

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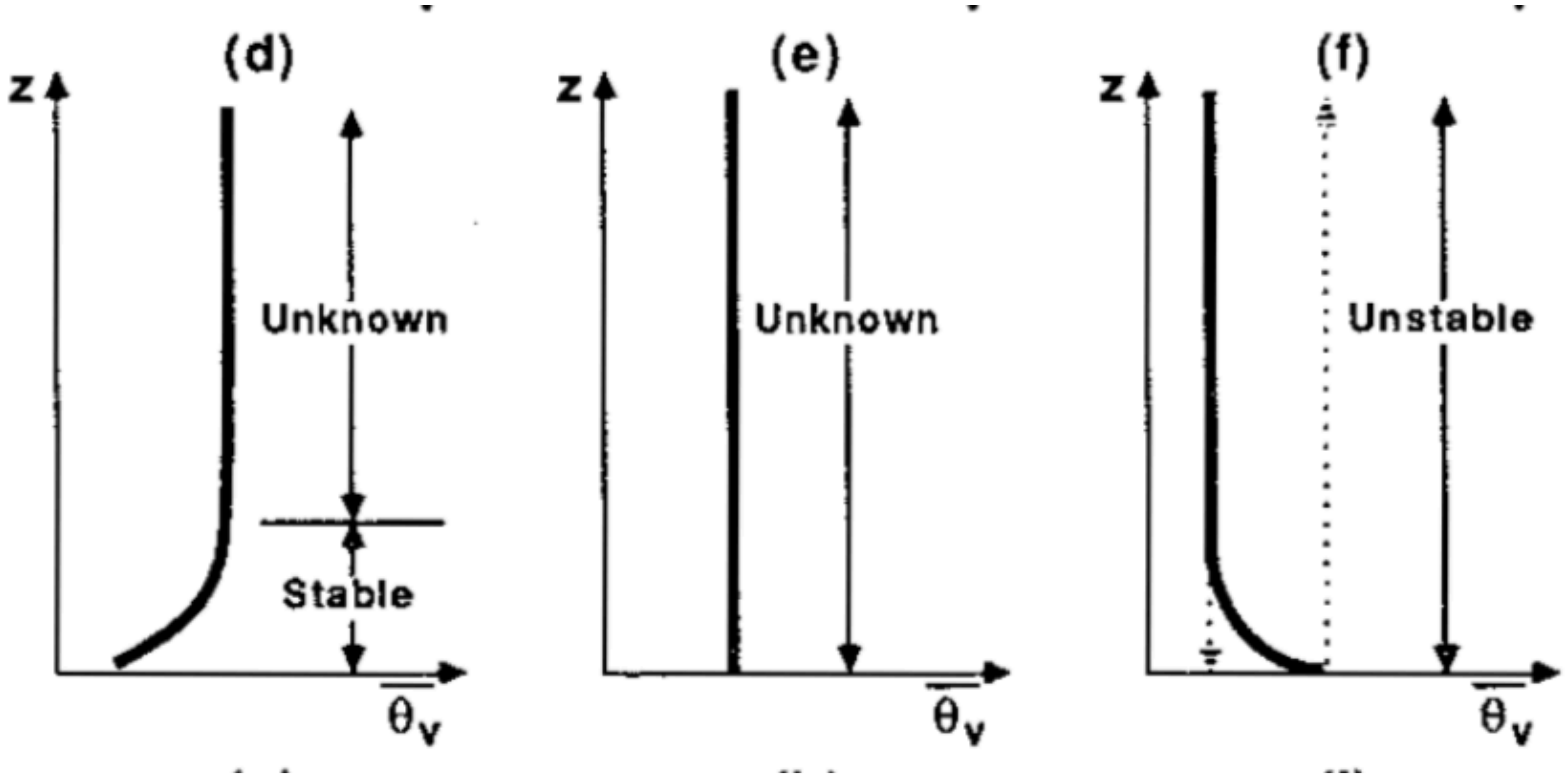


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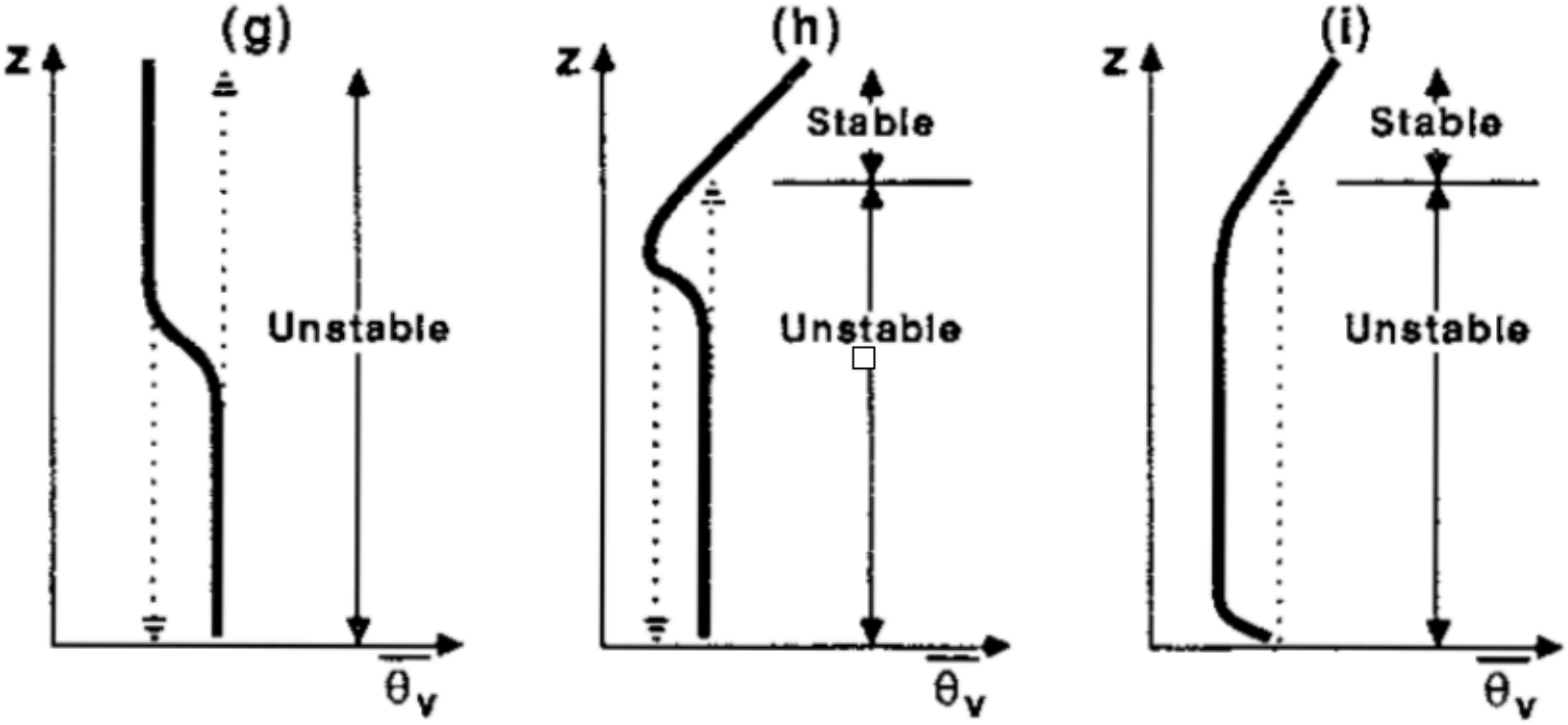


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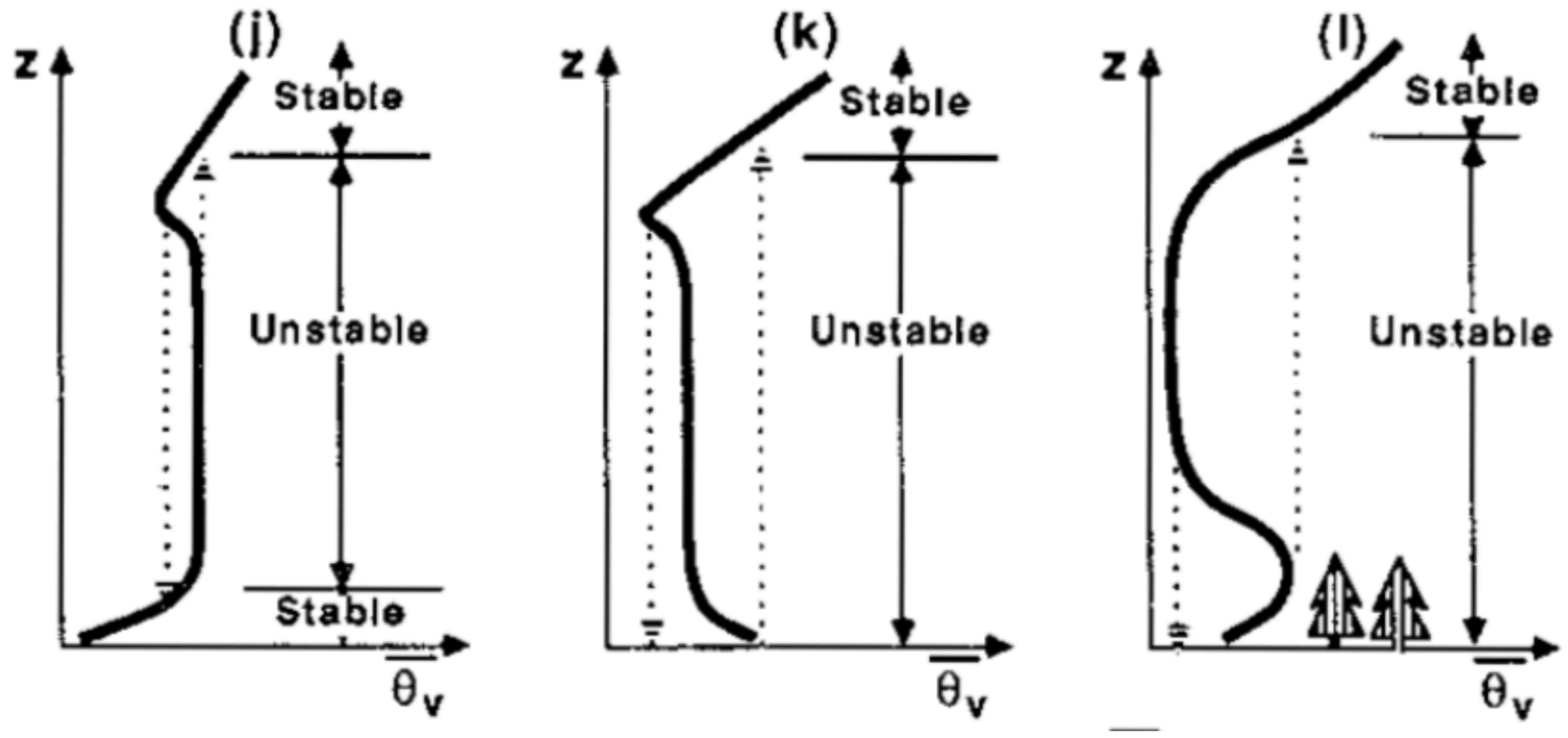


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Stull 1988

# Obukhov Length

$$L = -\frac{\bar{\theta}_v u_*^3}{g k (\overline{w'\theta_v'})_s}$$ **meters**

- L negative during daytime (unstable) and positive at night (stable)
- larger L magnitude corresponds to more shear and/or less heat flux
- L blows up when surface heat flux transitions pos/neg or neg/pos

**Physical interpretation: scale height where buoyancy dominates over shear**

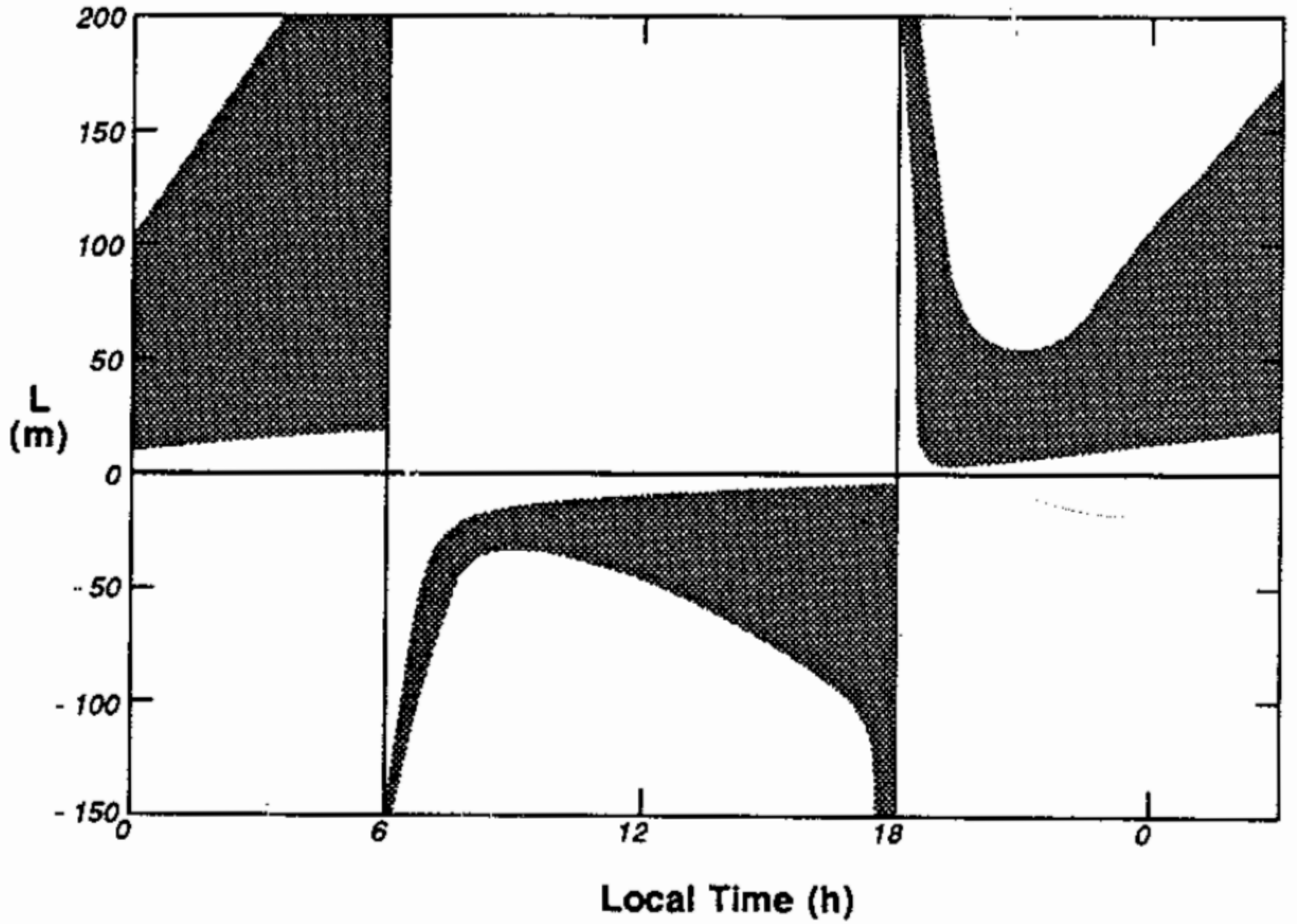
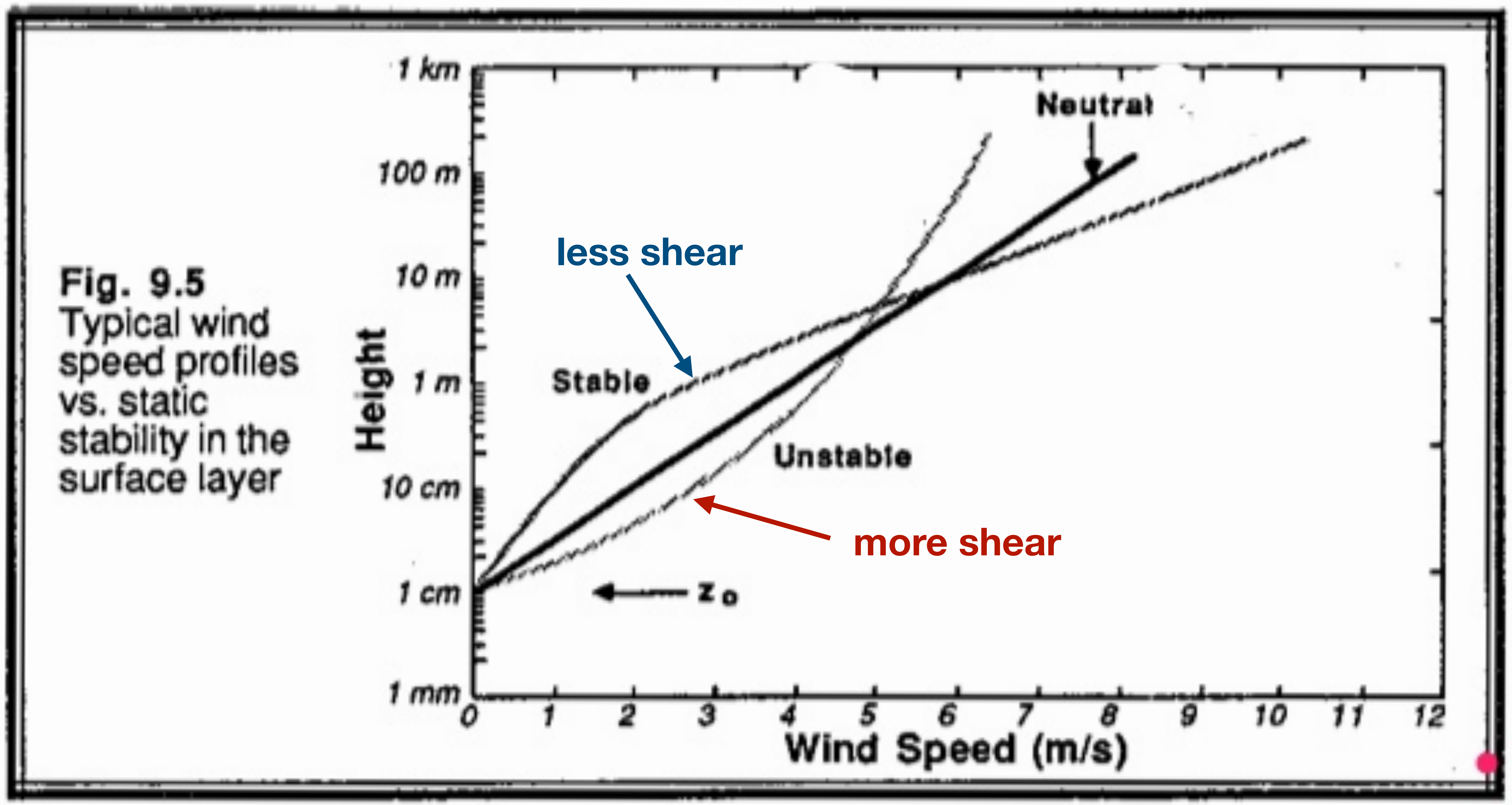



Fig. 5.21 Typical ranges of Obukhov length (L) evolution over a diurnal cycle.

Stull 1988

# Diabatic Wind Profiles




Stull 1988

# Stability Parameters

## Gradient Ri

$$Ri = \frac{(g/\overline{\theta_v})\, \partial\overline{\theta_v}/\partial z}{(\partial\overline{U}/\partial z)^2}$$

$$Ri \approx \frac{\text{buoyancy forcing}}{\text{shear forcing}}$$

- predicts turbulent/laminar flow
- applies any height in PBL

## Obukhov Length

$$L = -\frac{\theta_v\, u_*^3}{g\kappa\, (\overline{w'\theta_v'})_s}$$

$$L \approx -\frac{\text{shear forcing}}{\text{buoyancy forcing}}$$

- requires turbulence
- applies near-surface layer

# Law of the Wall (Logarithmic Wind Profile)

Neutral

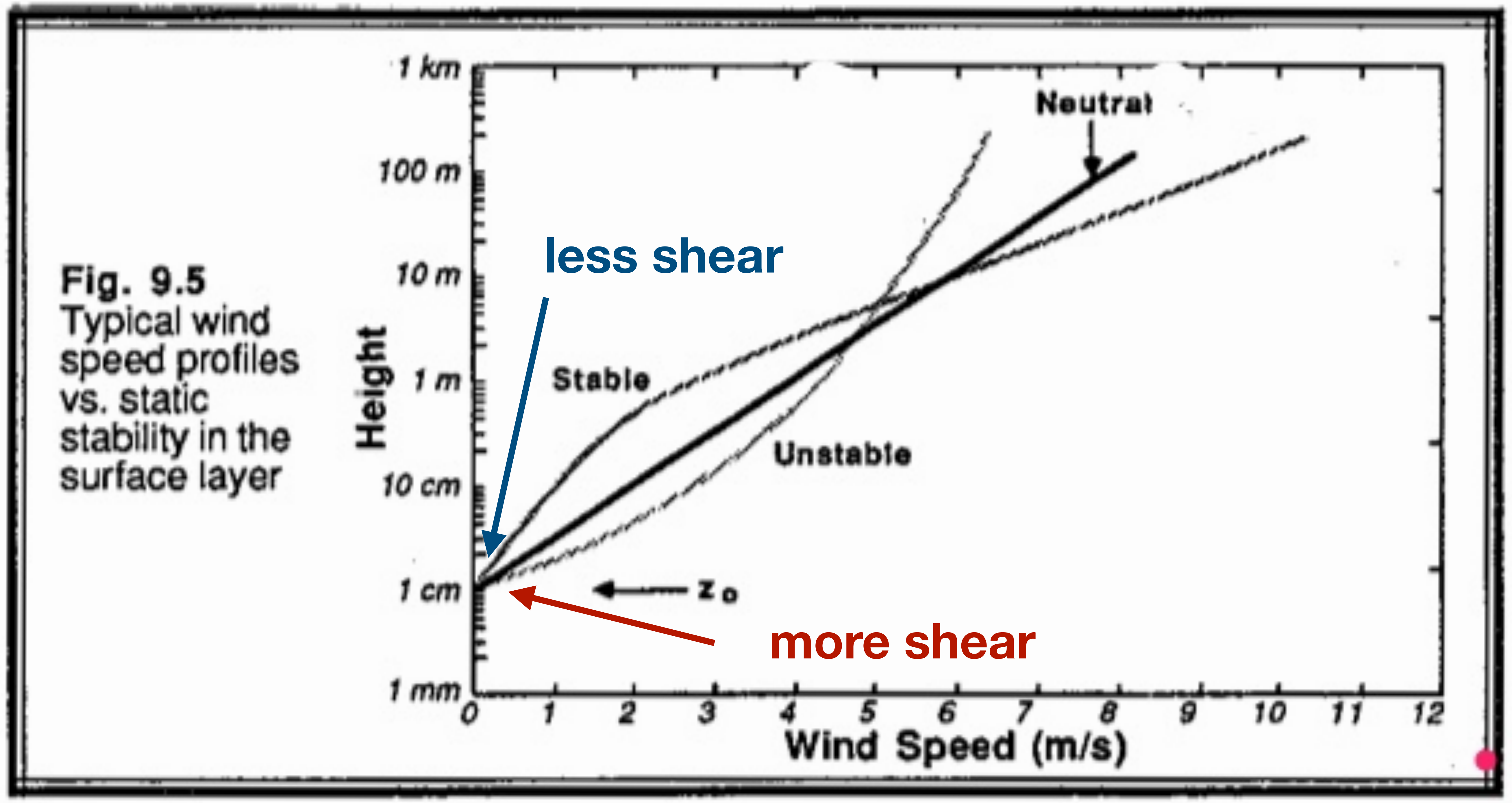
$$\frac{k}{u_*}\frac{d\overline{U}}{d\ln z} = 1$$

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

Diabatic

$$\frac{k}{u_*}\frac{d\overline{U}}{d\ln z} = \phi_m(\frac{z}{L})$$

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




Stull 1988

# Drag and Roughness, $z_0$

$$\frac{\tau}{\rho} = u_*^2 = C_D U^2$$

$$C_D = \frac{u_*^2}{U^2}$$

Intuitively, the more rough the surface ➠ more drag

Recall, for neutral conditions: $\bar{U}(z) = \frac{u_*}{k} ln \frac{z}{z_0}$ (↑$z_0$, ↓U)

Rewrite as $\frac{u_*}{U(z)} = k[ln \frac{z}{z_0}]^{-1}$

$C_D = \frac{u_*^2}{U(z)^2} = k^2 [ln \frac{z}{z_0}]^{-2}$ (↑$z_0$, ↑$C_D$)

# Bulk Flux Parameterizations

$$\text{vertical flux} = \text{Coeff} \cdot [\text{horizontal advective flux}]$$

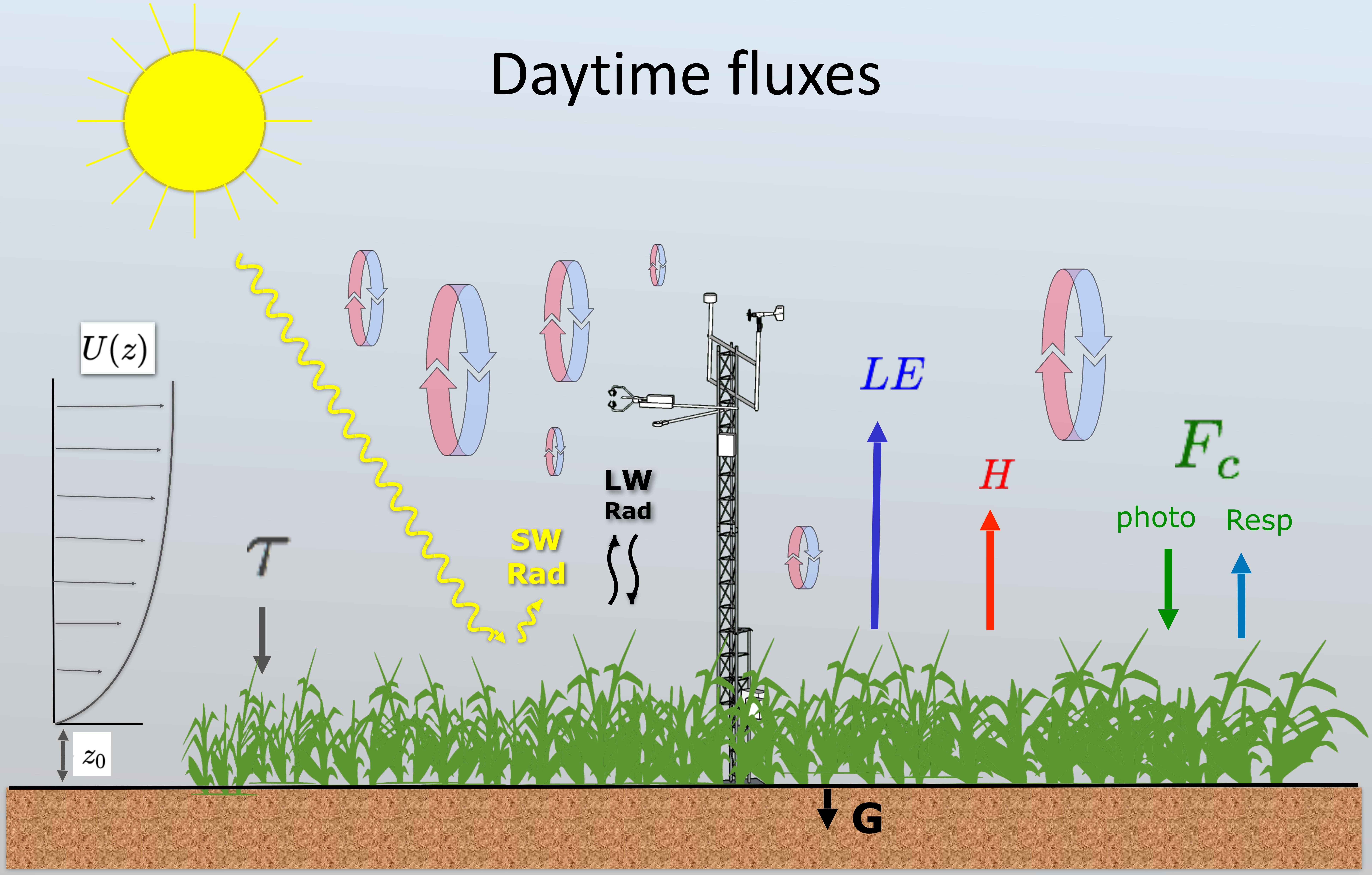
**momentum:** $$\tau = \mathrm{C_{D10}} \cdot \left[ U_{10} \cdot \rho_a U_{10} \right]$$

**heat:** $$H = \mathrm{C_{H10}} \cdot \left[ U_{10} \cdot \rho_a \mathrm{c_p} (\theta_s - \theta_{10}) \right]$$

**moisture:** $$H_L = \mathrm{C_{E10}} \cdot \left[ U_{10} \cdot \rho_a \mathrm{L_v} (q_s - q_{10}) \right]$$

$\mathrm{C_D},\ \mathrm{C_H},\ \mathrm{C_E}$ depend on $z_0$, $z/L$, …

# Daytime fluxes




**Energy Balance:** $R_n - G = H + LE$

# Surface Energy Balance

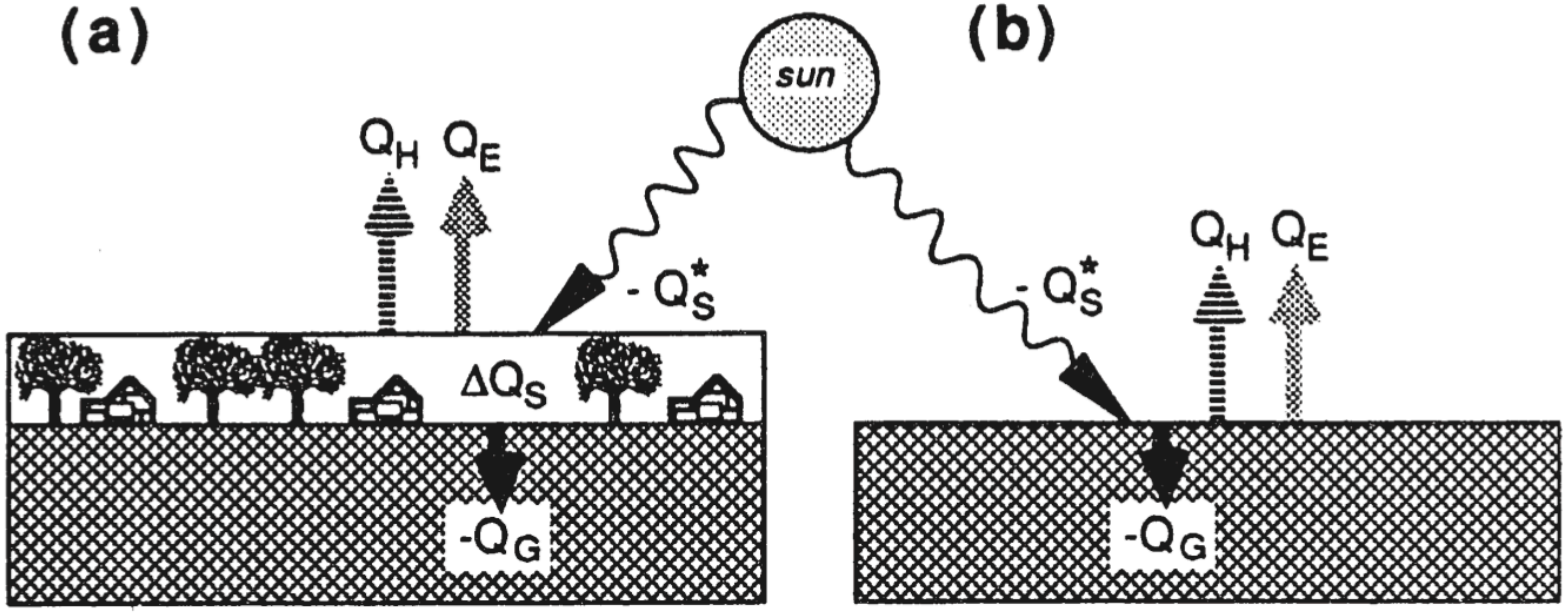



**Fig. 7.2** Contributions to the surface energy balance (a) for a finite thickness box and (b) for an infinitesimally thin layer. $-Q_S^*$ is the net radiative contribution, $Q_H$ is turbulent sensible heat flux, $Q_E$ is turbulent latent heat flux, $-Q_G$ is molecular flux into the ground, and $\Delta Q_S$ is storage.

**Stull 1988**

$$-Q_s^* = Q_H + Q_E - Q_G + \Delta Q_S \qquad (7.2b)$$

$Q^*_s$ = net upward radiation at the surface

$Q_H$ = represents the upward sensible heat flux out of the top

$Q_E$ = represents the upward latent heat flux out of the top

$Q_G$ = represents the upward molecular heat flux into the bottom

$\Delta Q_S$ = denotes the storage or intake of internal energy (positive for warming and for chemical storage by photosynthesis).

# Surface energy balance

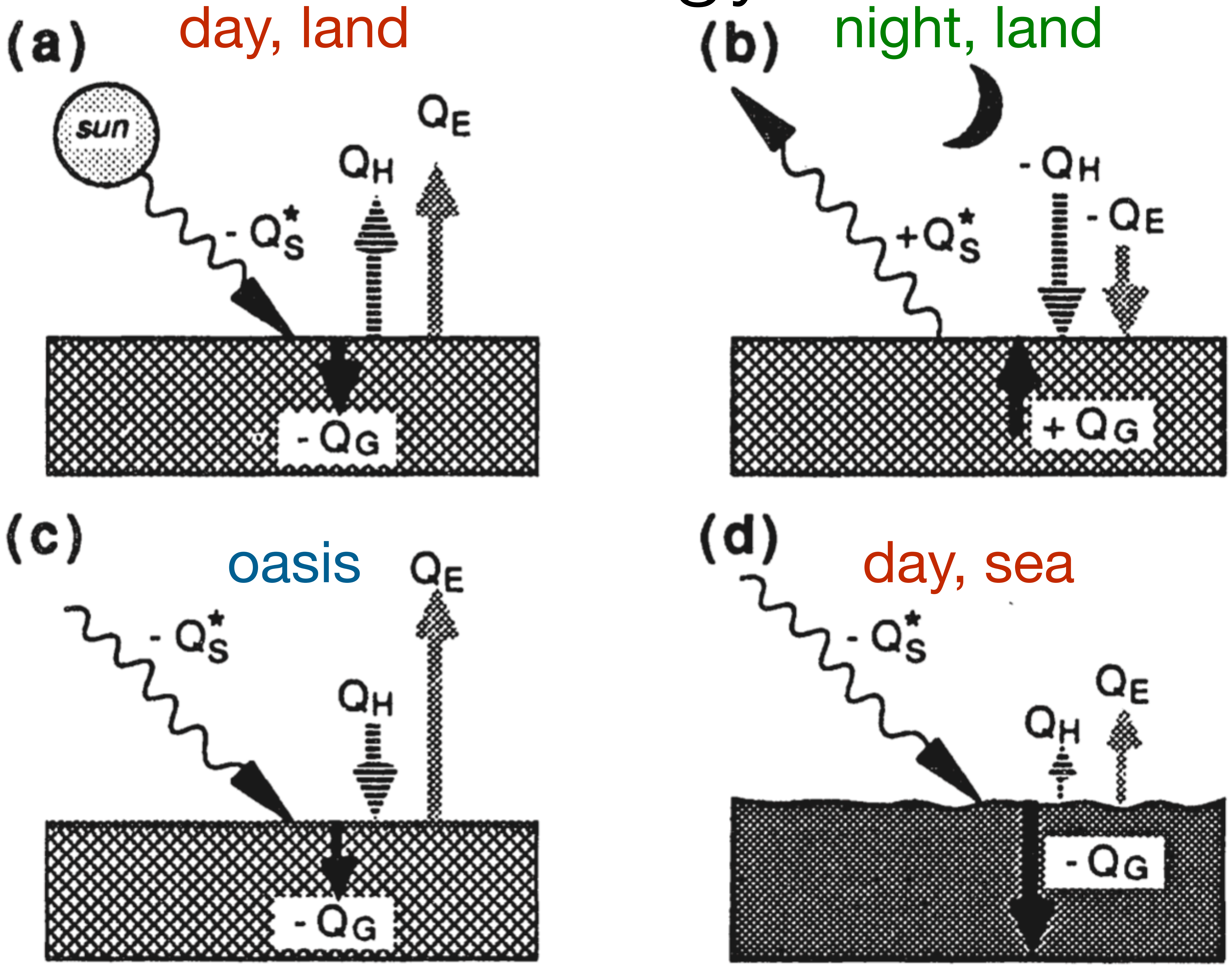


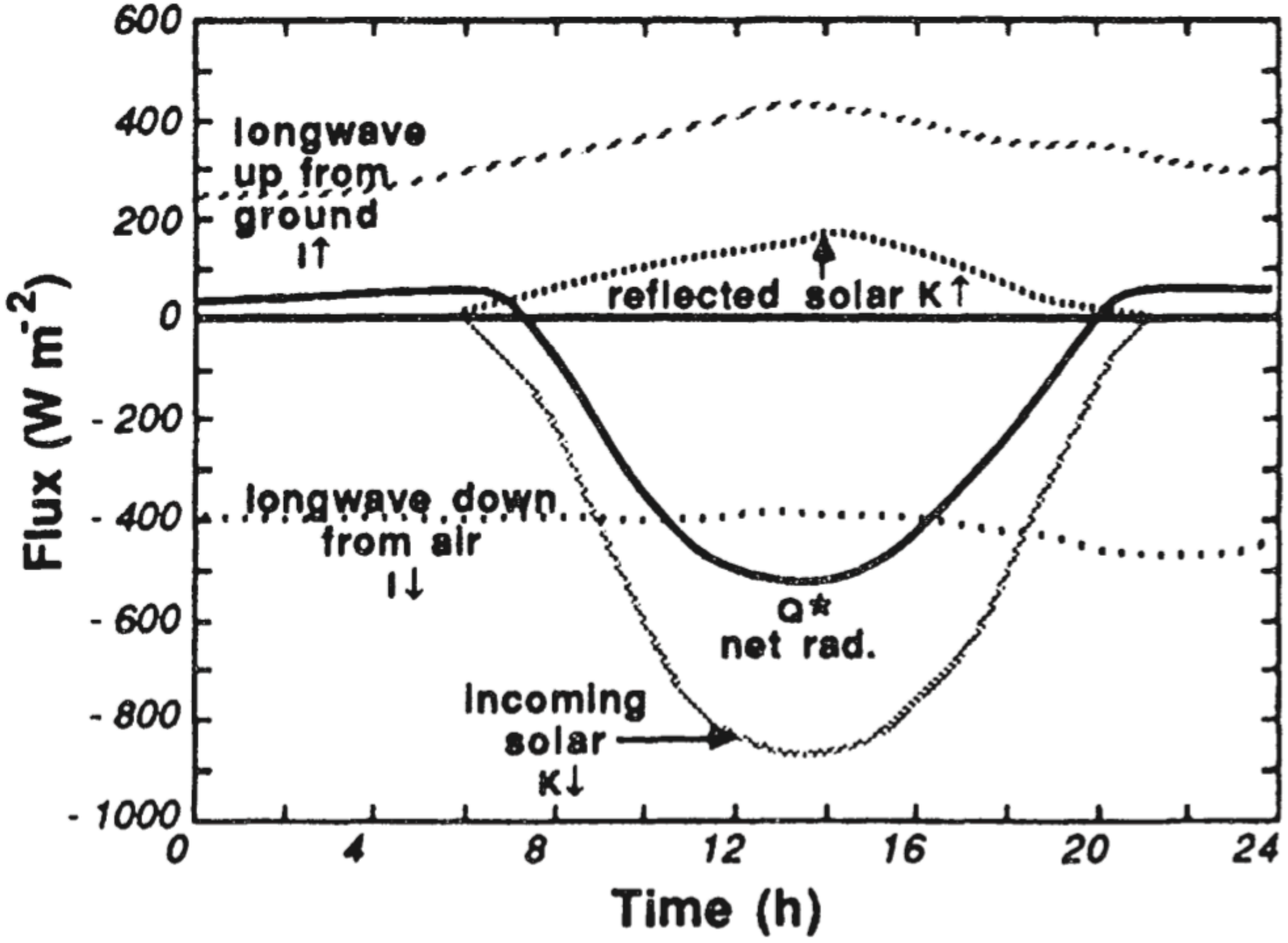

**Fig. 7.3** Typical variation of terms of the surface energy balance for (a) daytime over land; (b) nighttime over land; (c) oasis effect of warm dry air advection over a moist surface; and (d) daytime over the sea with no advection. Arrow size indicates relative magnitude.

**Stull 1988**

# Surface Energy Balance—Radiation Components

**Fig. 7.5** **Stull 1988**

Surface radiation budget components for 30 July 1971, at Matador, Saskatchewan (50°N) over a 0.2 m stand of native grass. Cloudless skies in the morning, increasing clouds in the late afternoon and evening (after Ripley and Redmann, 1976).

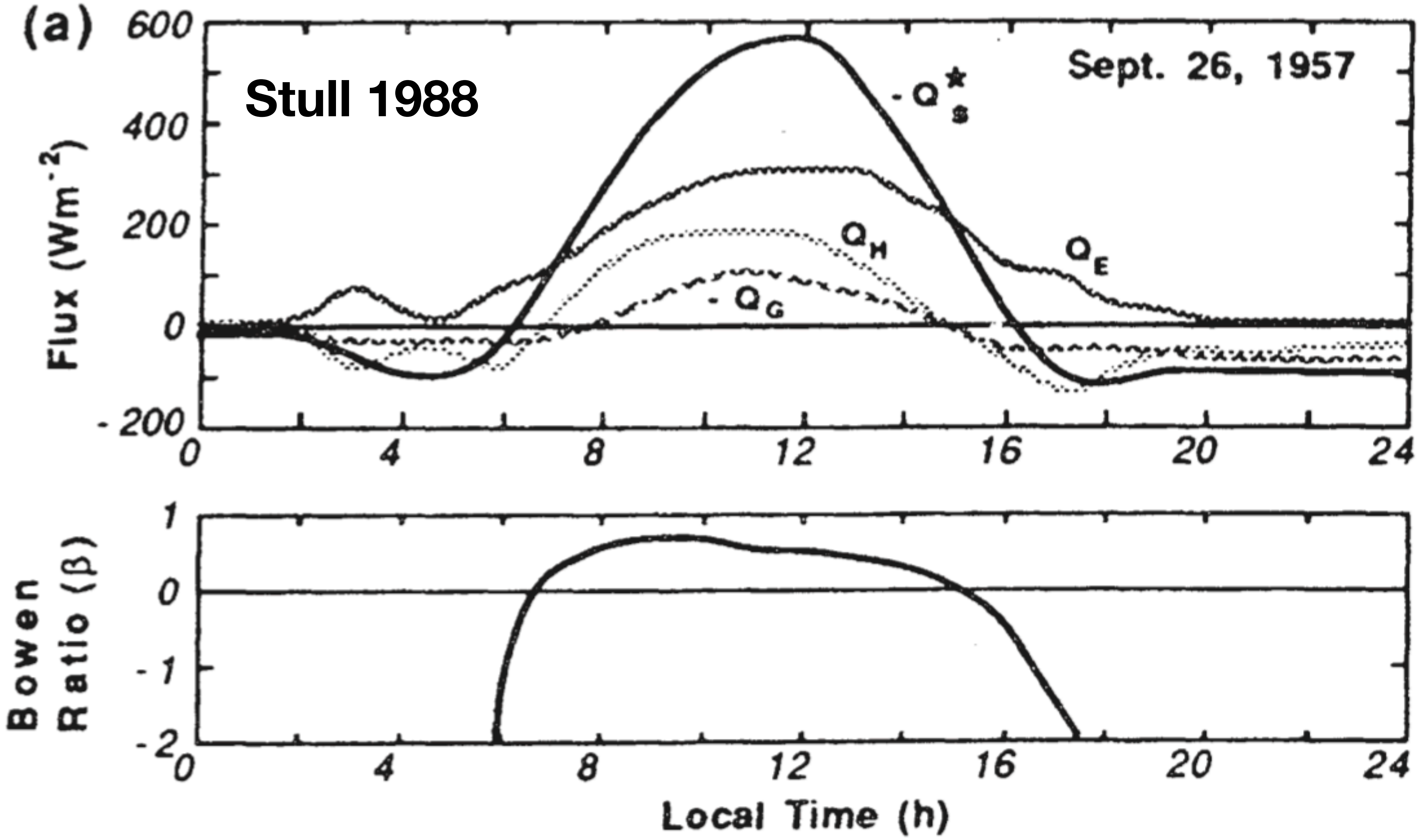


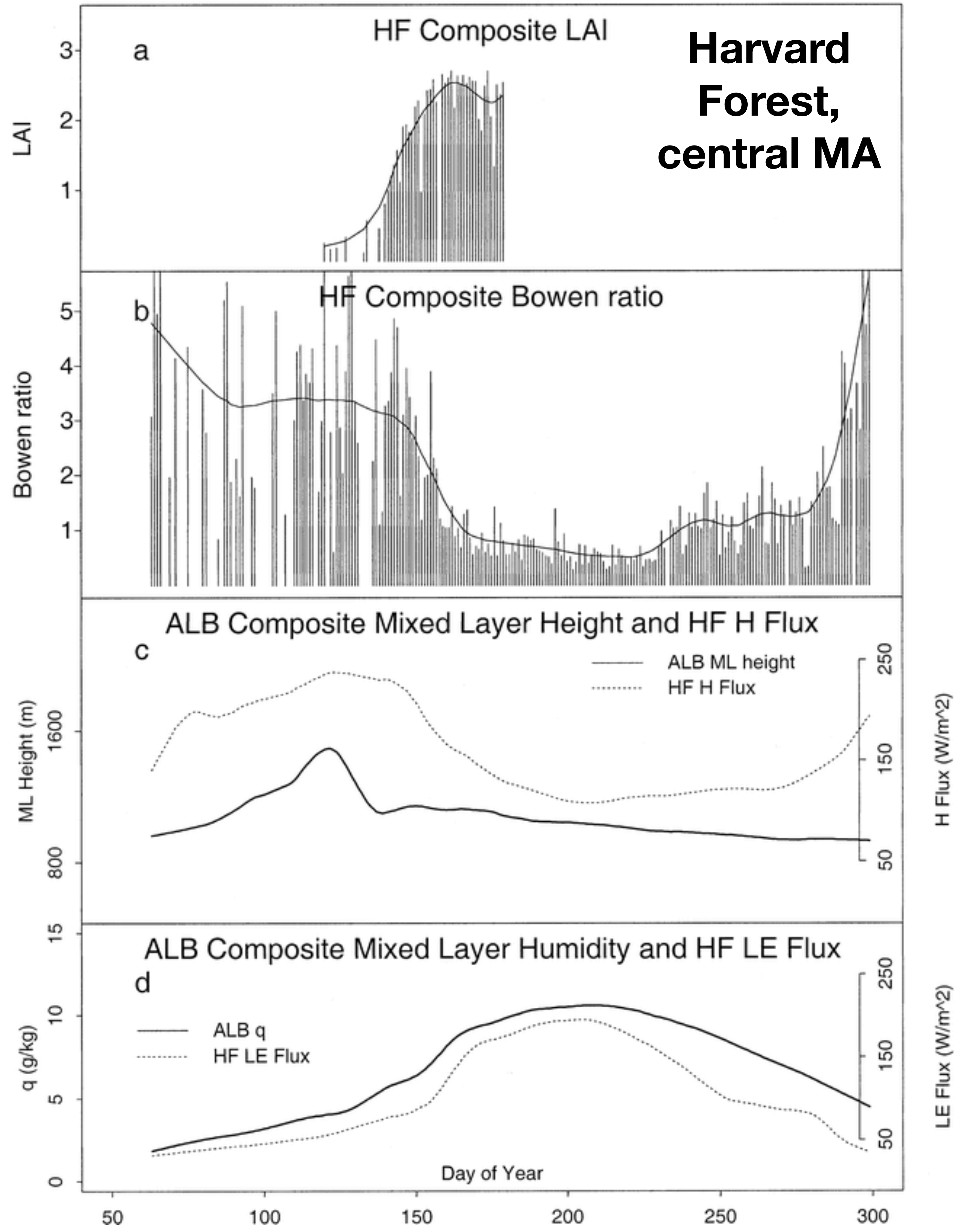

**126 NYSM standard sites measure incoming solar**
**18 NYSM flux sites measure all 4 components (radiation)**

# Surface Energy Balance—Fluxes, Bowen Ratio

$$\beta = \frac{Q_H}{Q_E} = \frac{\text{sensible heat}}{\text{latent heat}}$$

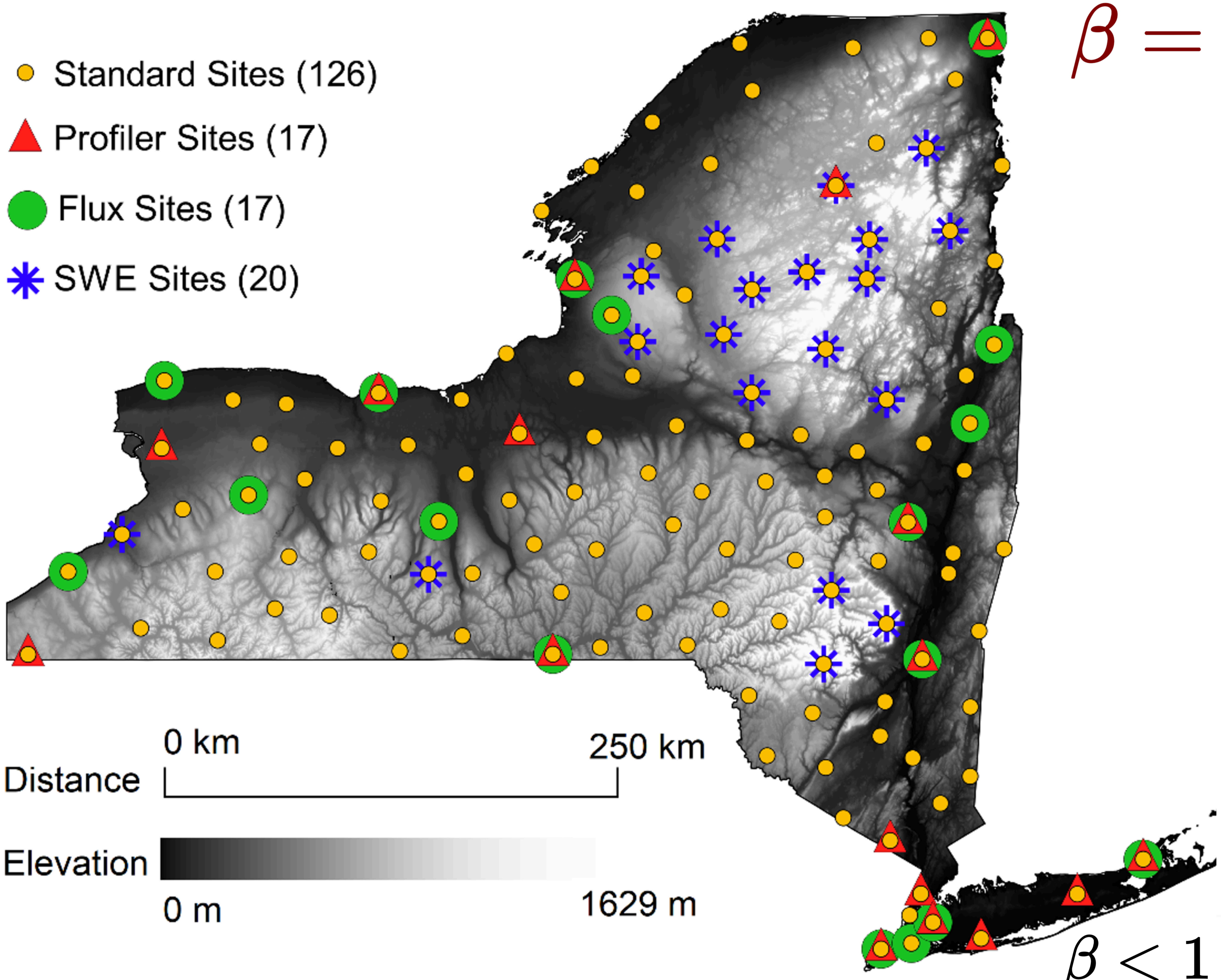
irrigated crop field




$\beta < 1$ for moist surface $\beta > 1$ for dry surface

# Surface Energy Balance—NYSM Leafout




$$\beta = \frac{Q_H}{Q_E} = \frac{\text{sensible heat}}{\text{latent heat}}$$

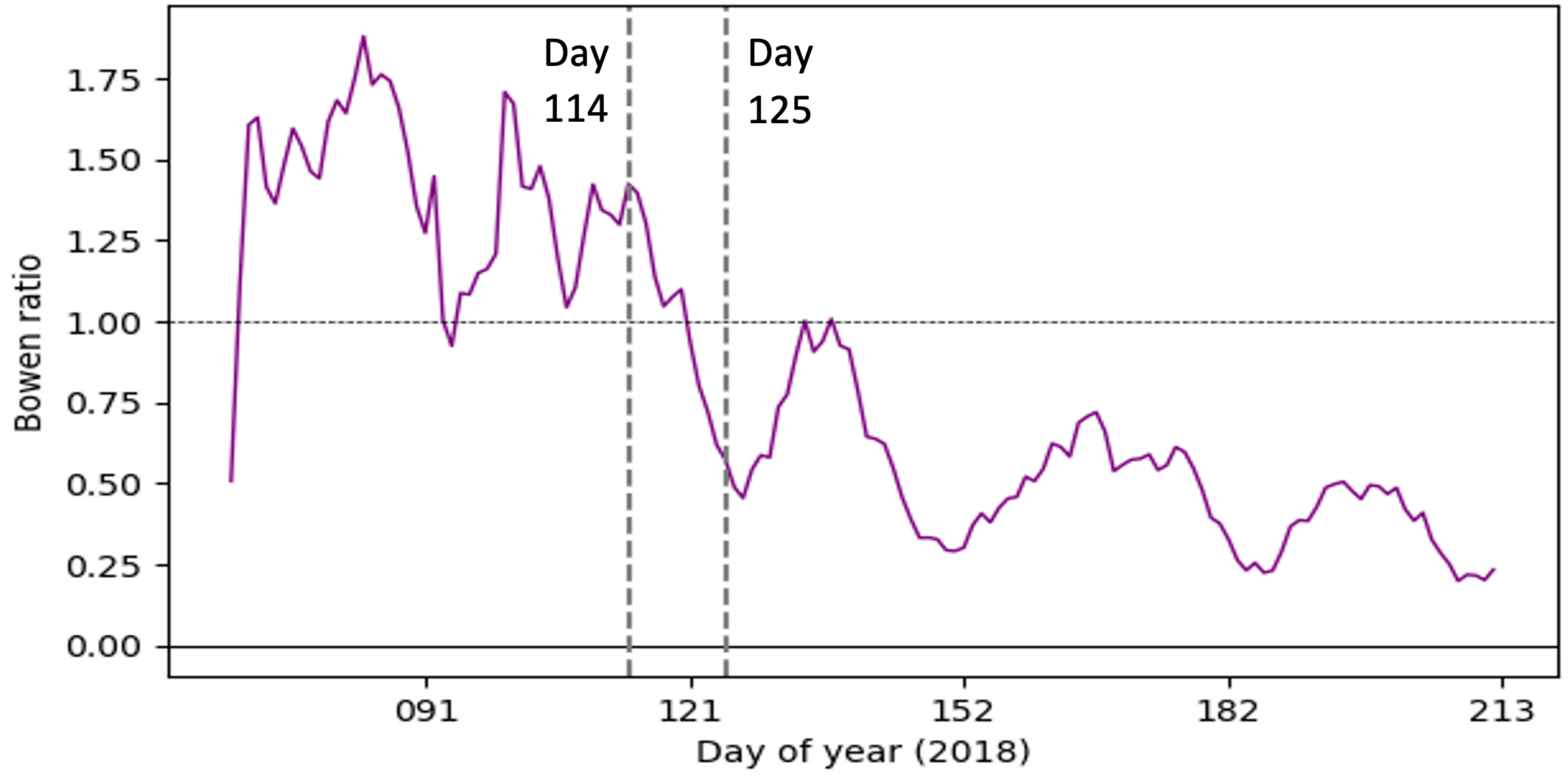
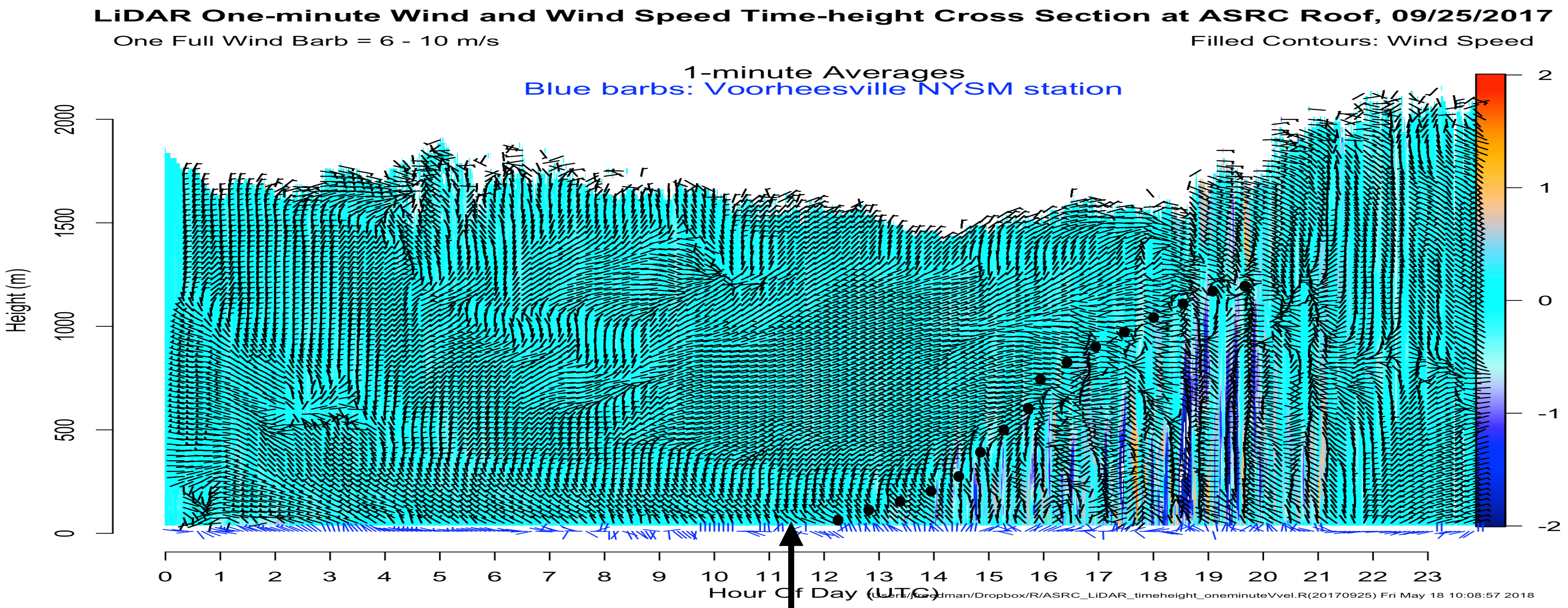
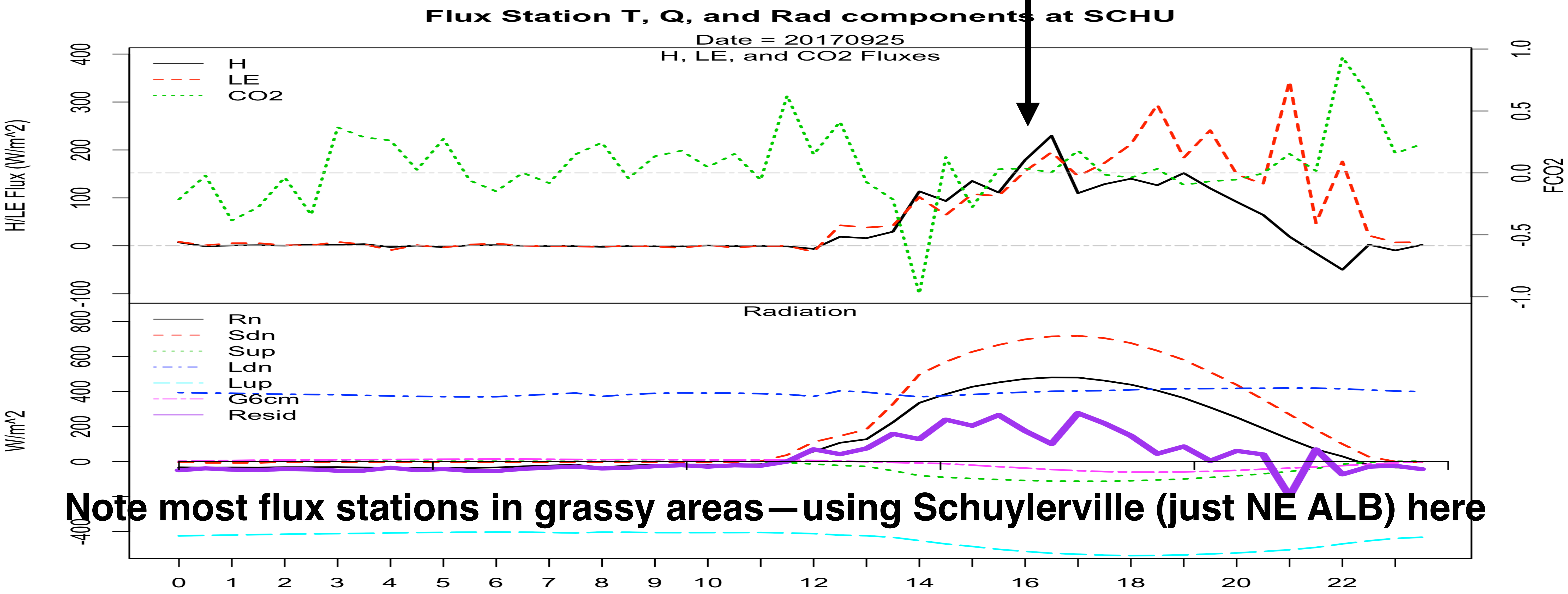



**Figure 8**. New York State Mesonet Flux Network average noontime (11–13 hrs) Bowen Ratio for all flux sites excluding New York City sites (BKLN, QUEE, STAT). The first vertical dotted line to the left indicates the state-wide average day of first visible grass growth (114), and the second vertical line represents the average day of first visible leaf emergence (125). Data with quality control grades 1-3 between Mar 10-Aug 1 2018 were used. Running mean …

$\beta < 1$ for moist surface $\quad \beta > 1$ for dry surface

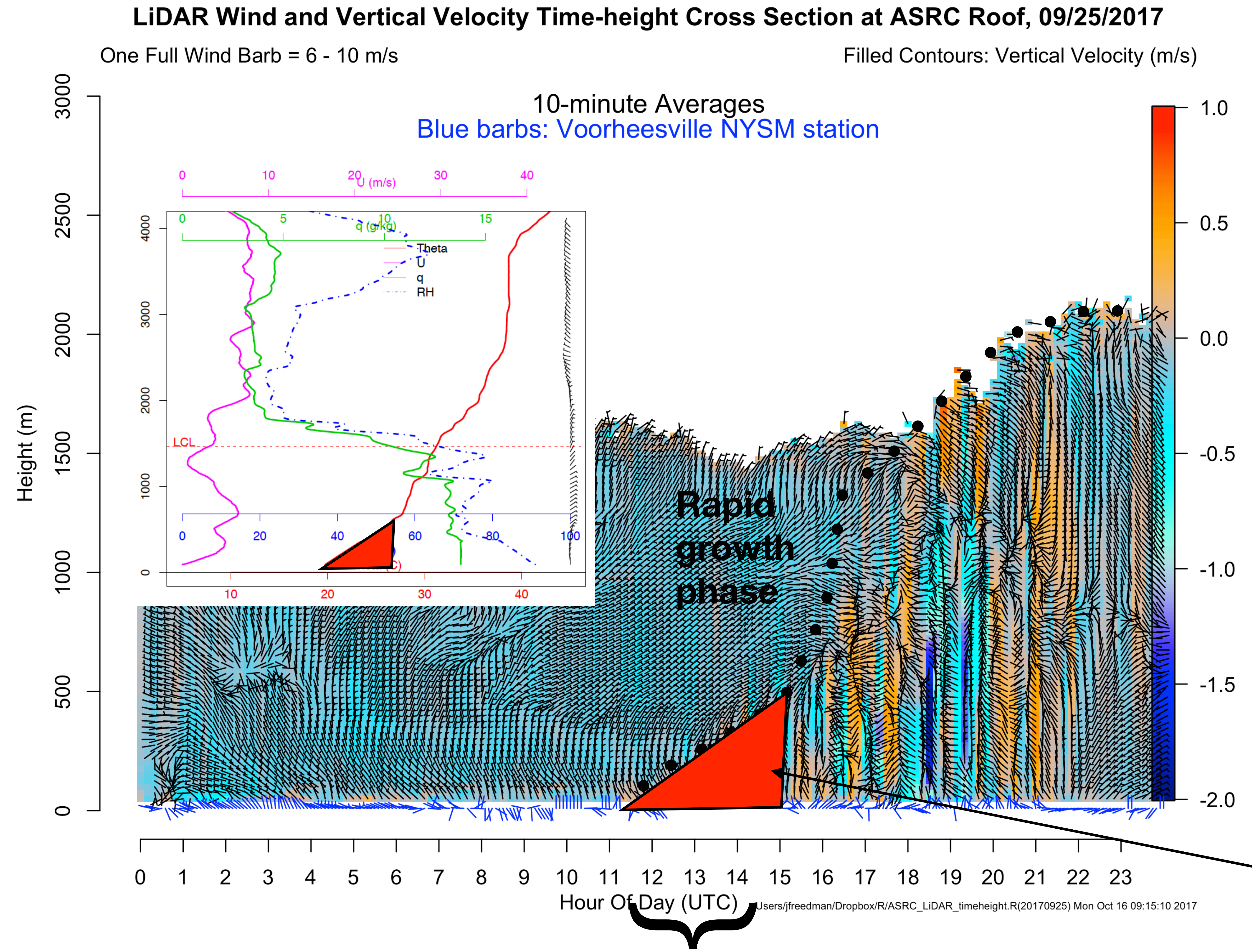



**Sunrise** **Noon**




**Note most flux stations in grassy areas—using Schuylerville (just NE ALB) here**

# Can you predict when surface inversion will erode? A homework problem!



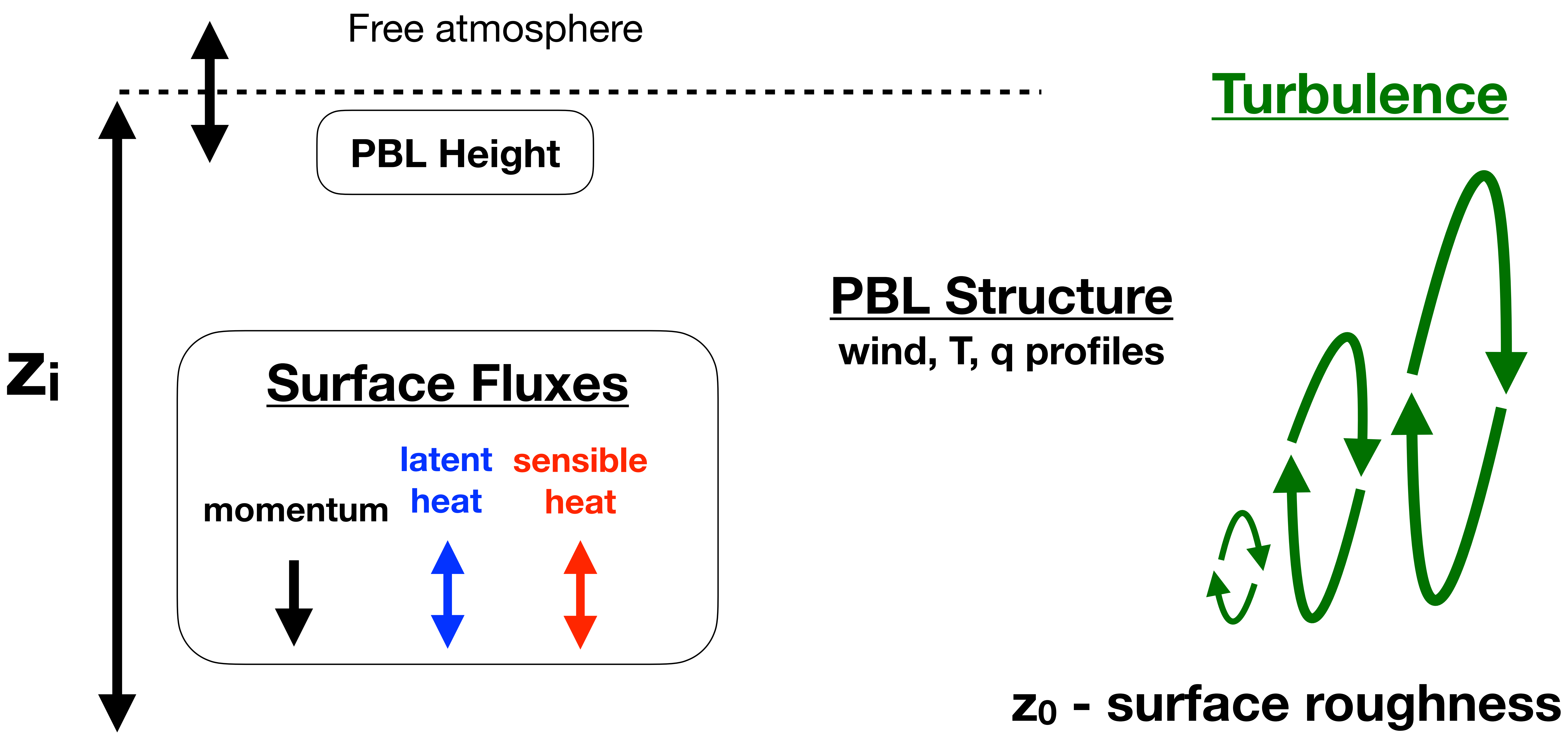

**Tennekes 1973**

$$t = \left[ \frac{2\Upsilon h_0 \Delta_0}{(w'\theta')_n} \right]^{1/2}$$

**Assume $\mathcal{T}$ = $10^4$ s, $h_0$ = thickness of the inversion, $(w'\theta')_n$ = midday surface flux (in kinematic units) and $\Delta_0$ = inversion strength, what is *t?***

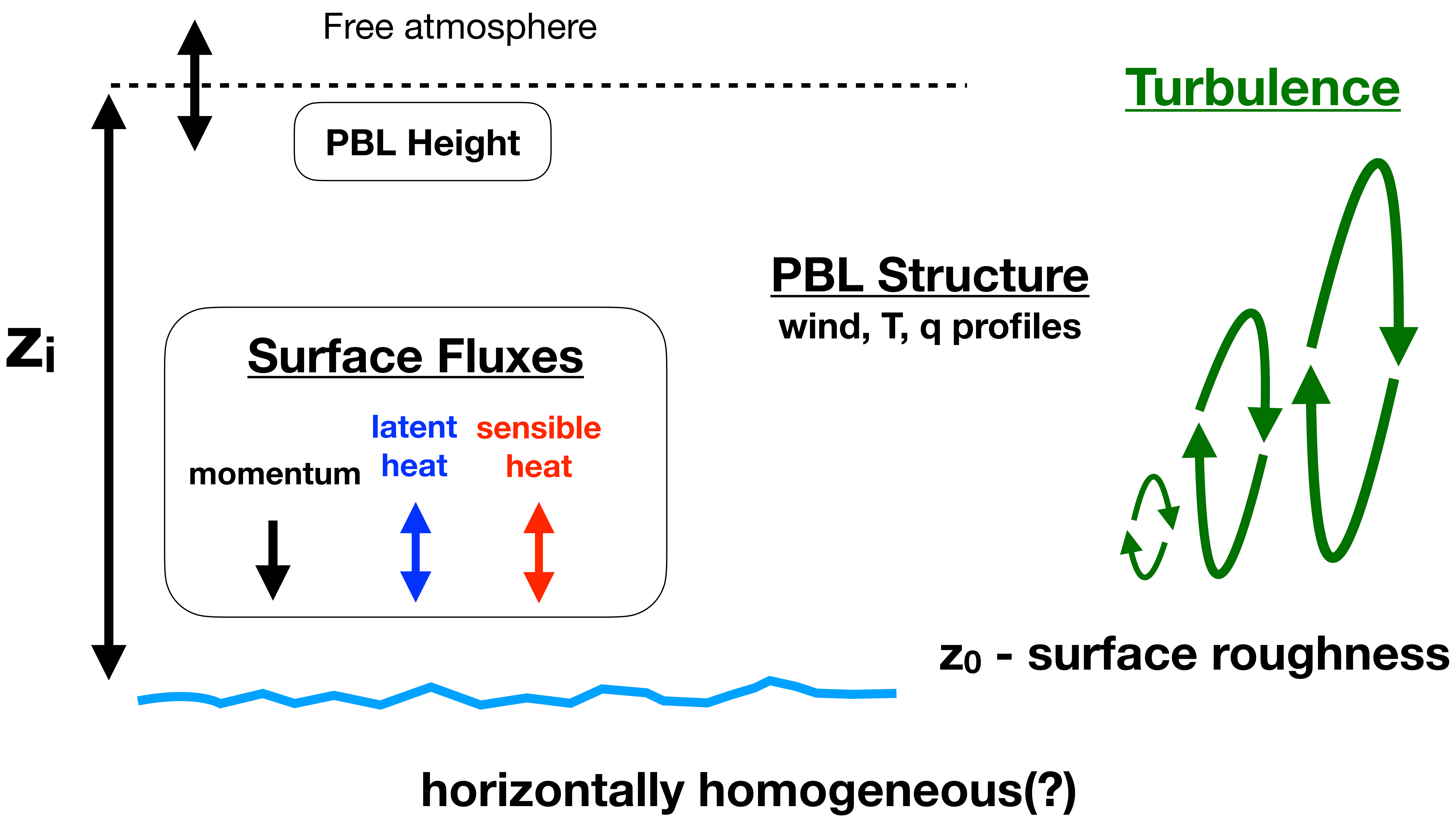
**Note: most flux stations in grassy areas—using Schuylerville here (previous slide)**

# Conceptual Model of the ABL
# But what about over the ocean?

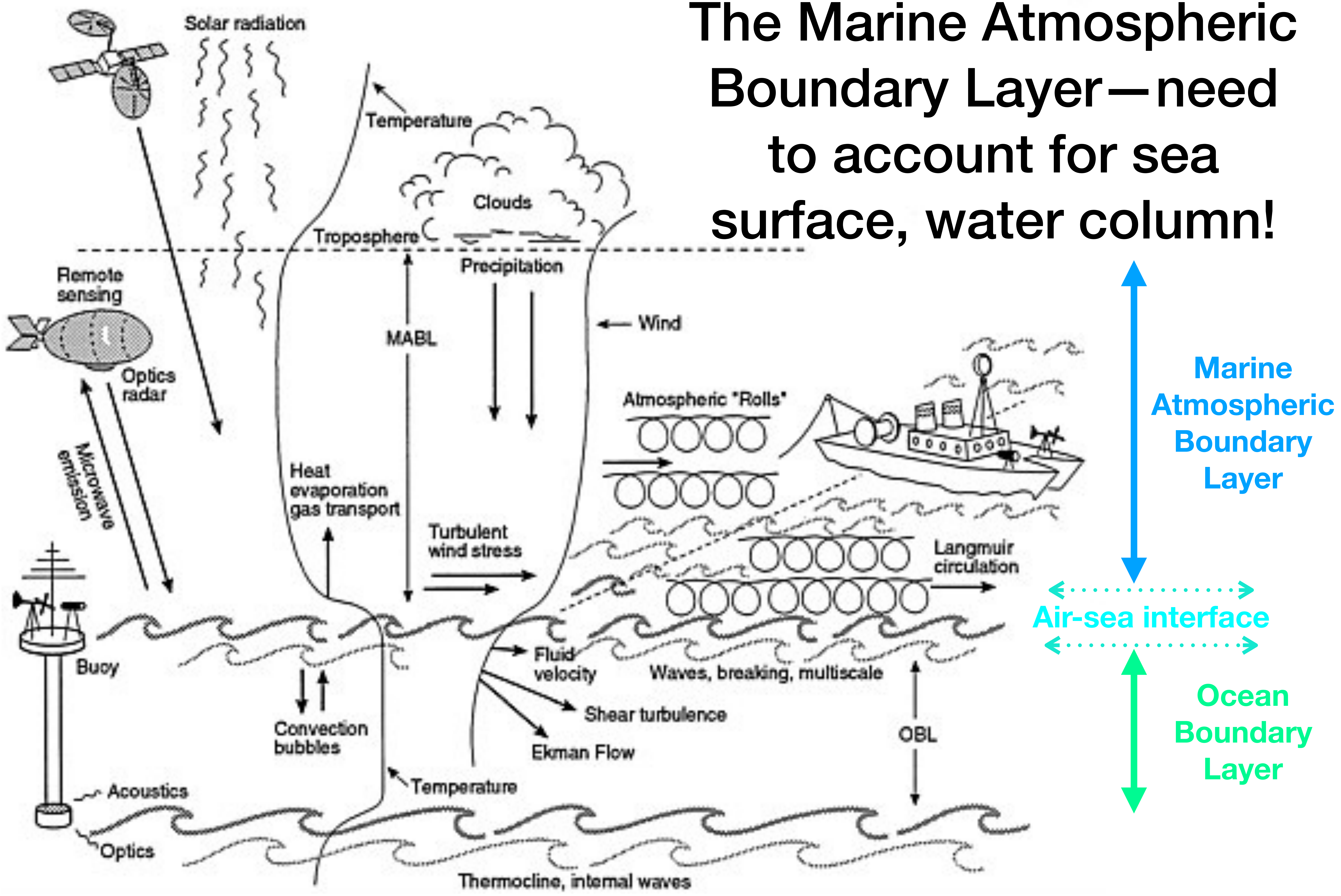



**horizontally homogeneous(?)**

# Conceptual Model of the ABL
# But what about over the ocean?



# The Marine Atmospheric Boundary Layer—need to account for sea surface, water column!



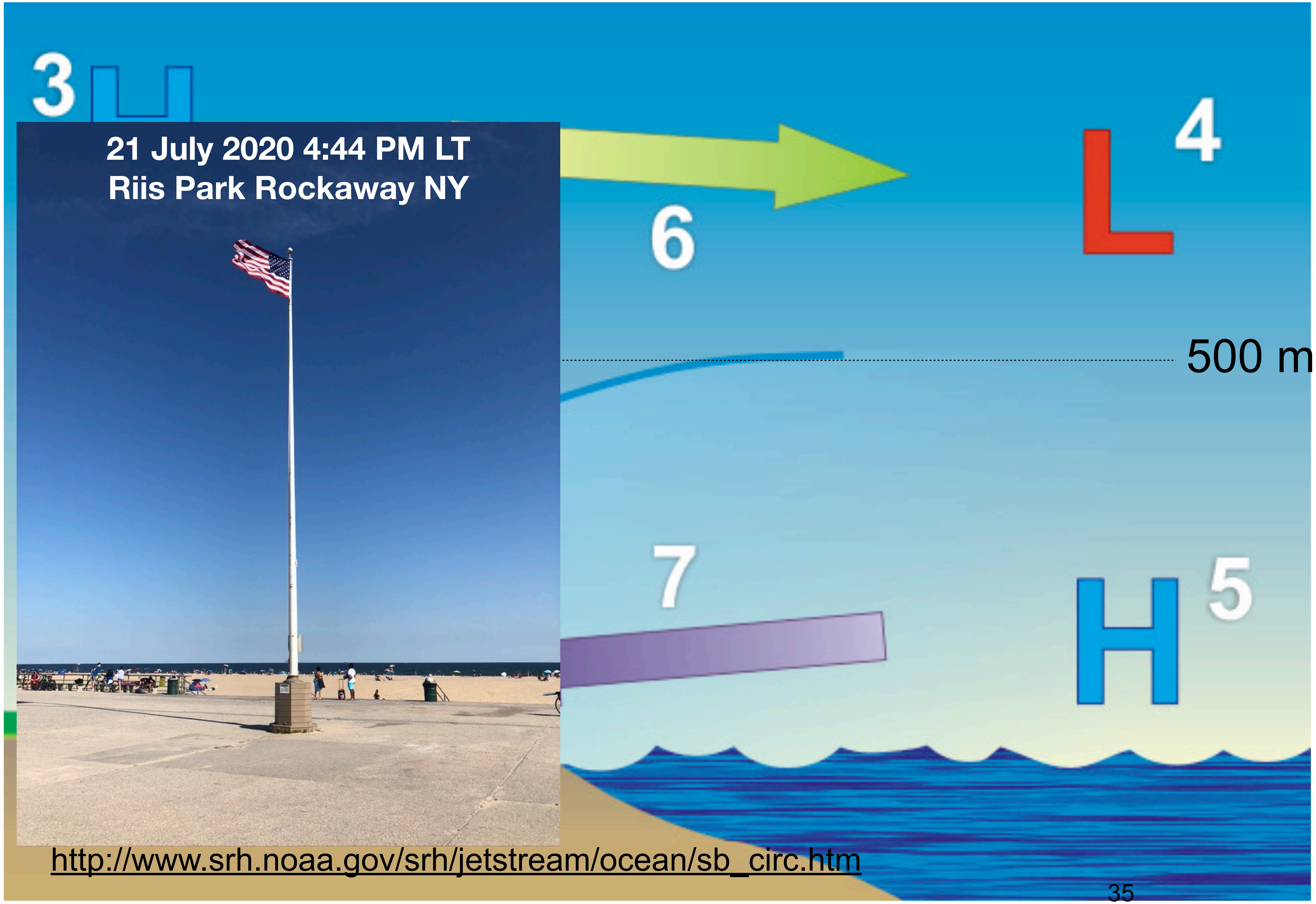
# Estimating H and LE over the ocean

$$Q_H = \rho c_p C_h u (T_s - (T_a + \gamma z))$$

$$Q_E = \rho L C_e u (q_s - q_a)$$

where ρ is the density of air; $c_p$, the specific heat capacity of air at constant pressure; $L$, the latent heat of vaporization; $C_h$ and $C_e$, the stability and height dependent transfer coefficients for sensible and latent heat respectively; $u$, the wind speed; $T_s$ , the sea surface temperature; $T_a$, the surface air temperature with a correction for the adiabatic lapse rate, $\gamma$; $z$, the height at which the air temperature was measured; $q_s$, 98% of the saturation specific humidity at the sea surface temperature to allow for the salinity of sea water, and $q_a$, the atmospheric specific humidity.

# Sea or lake breeze circulation: Daytime




http://www.srh.noaa.gov/srh/jetstream/ocean/sb_circ.htm

# Or if you prefer Stull.....

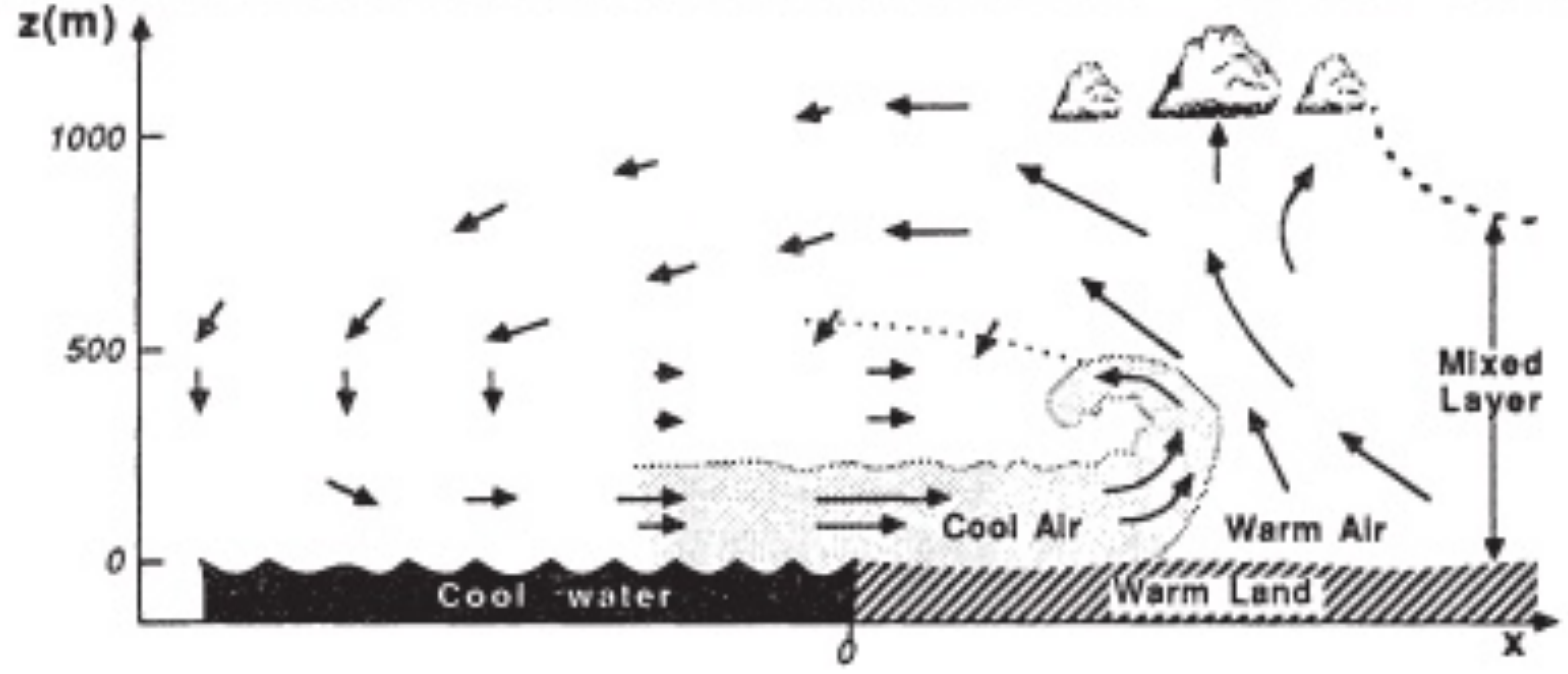
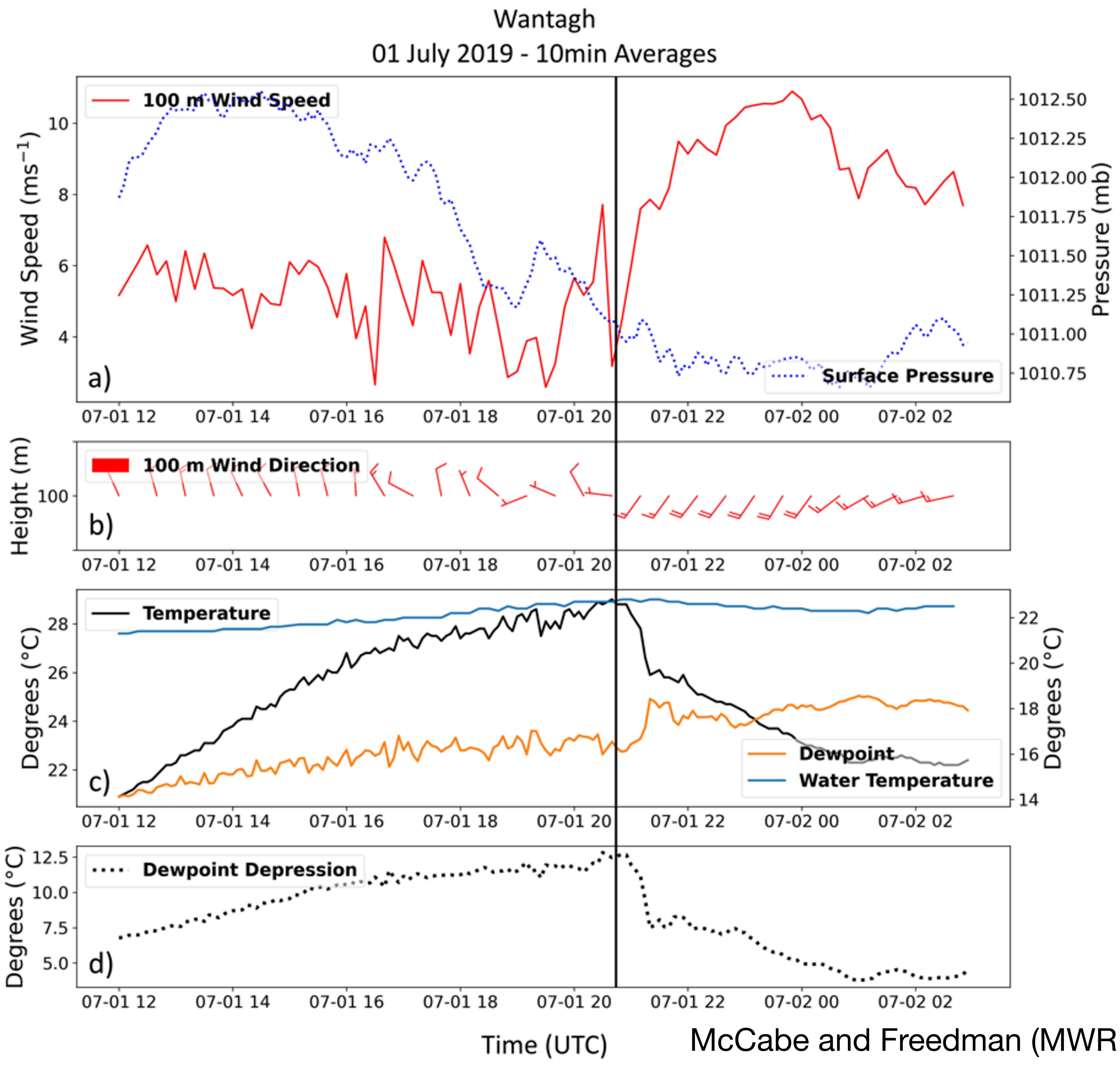
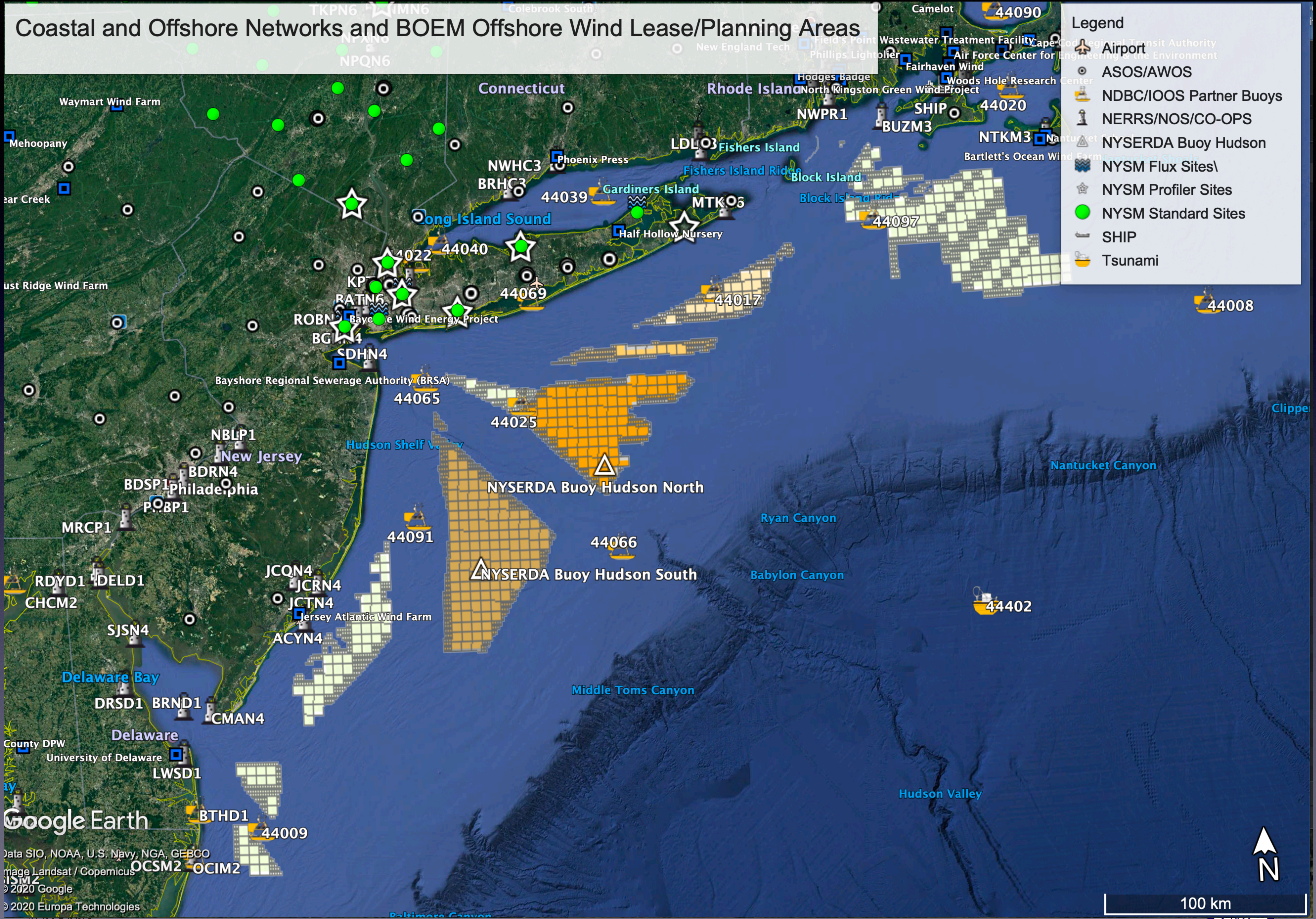



**Fig. 14.7** Idealized sea or lake-breeze circulation. (After Lyons (1975) and Ogawa, et al., 1986).



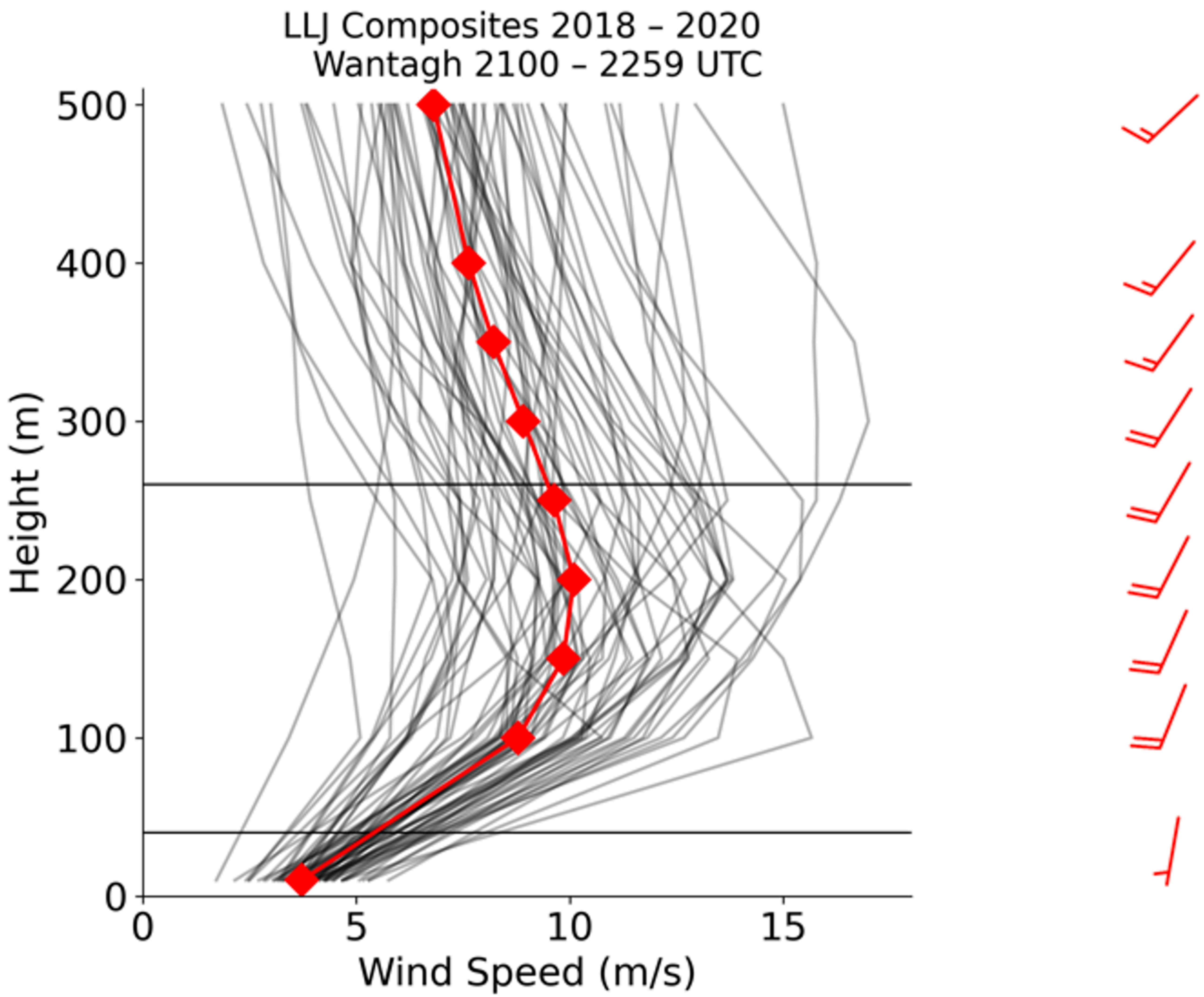
McCabe and Freedman (MWR 2023)

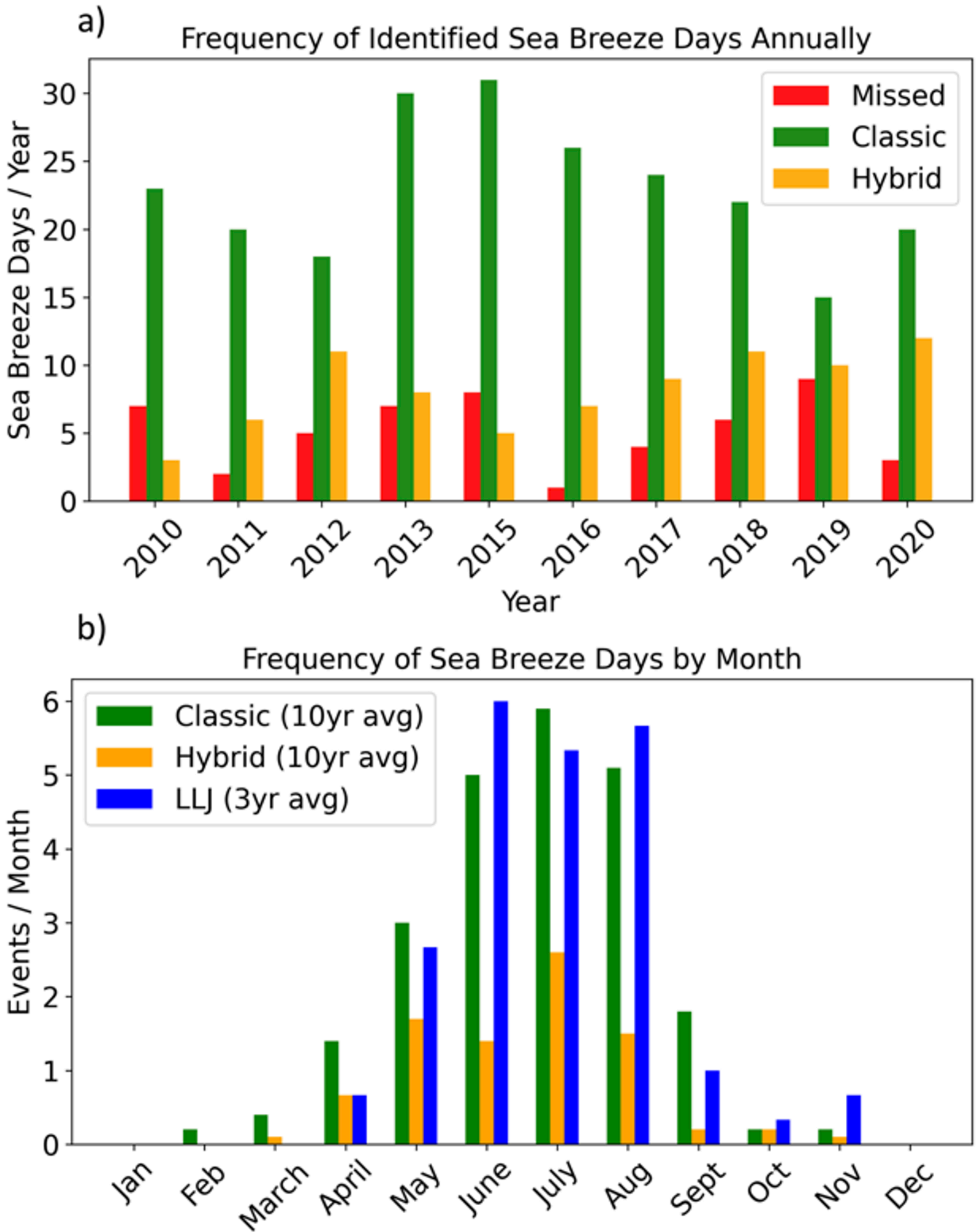
# Coastal and Offshore Networks and BOEM Offshore Wind Lease/Planning Areas



# NY Bight

## McCabe and Freedman (MWR 2023)






Sea breeze and LLJ Climatology (McCabe and Freedman 2023)

# Questions?

Horns Rev 12 February 2008 1010 UTC — Photo by Christian Steiness