The Hydrosphere:
Lecture 4: Hydrometeorology

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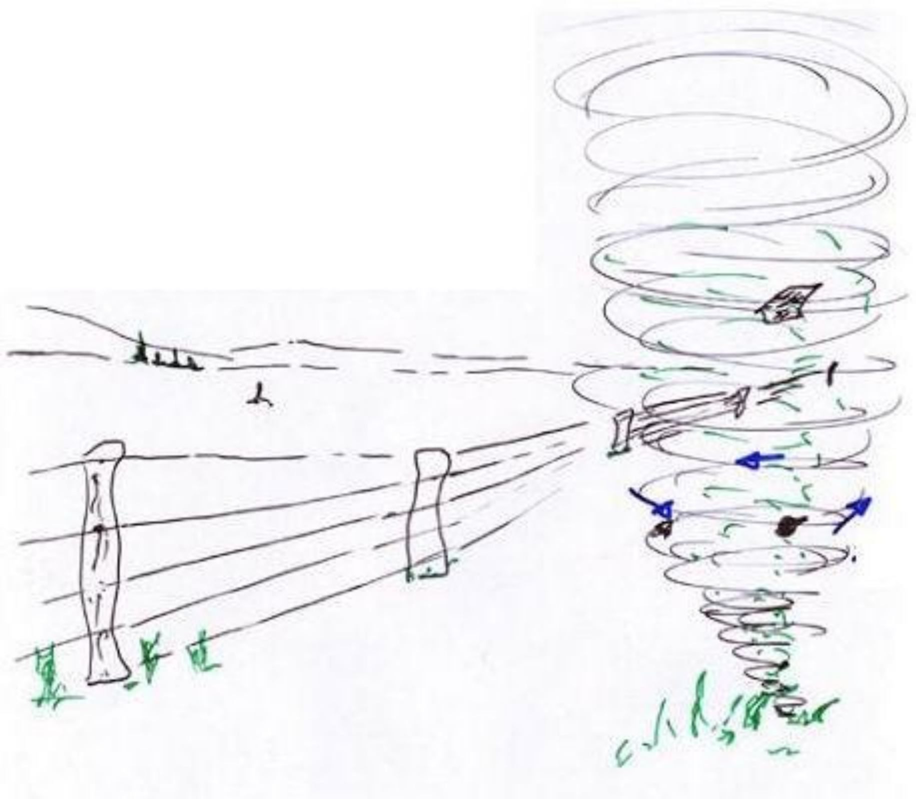
Hydrometeorology Defined

Hydrometeorology: a branch of meteorology and hydrology that studies the transfer of water and energy between the land surface and the lower atmosphere.

- Hydrometeorology is the science that is used to predict Climate Change.
- The branch of meteorology that deals with the occurrence, motion, and changes of state of atmospheric water
- That branch of meteorology which relates to, or treats of, water in the atmosphere, or its phenomena, as rain, clouds, snow, hail, storms, etc.



Micrometeorology: the branch of meteorology that deals with the small-scale processes, physical conditions, and interactions of the lowest part of the atmosphere, esp. in the first few hundred feet above the earth's surface.

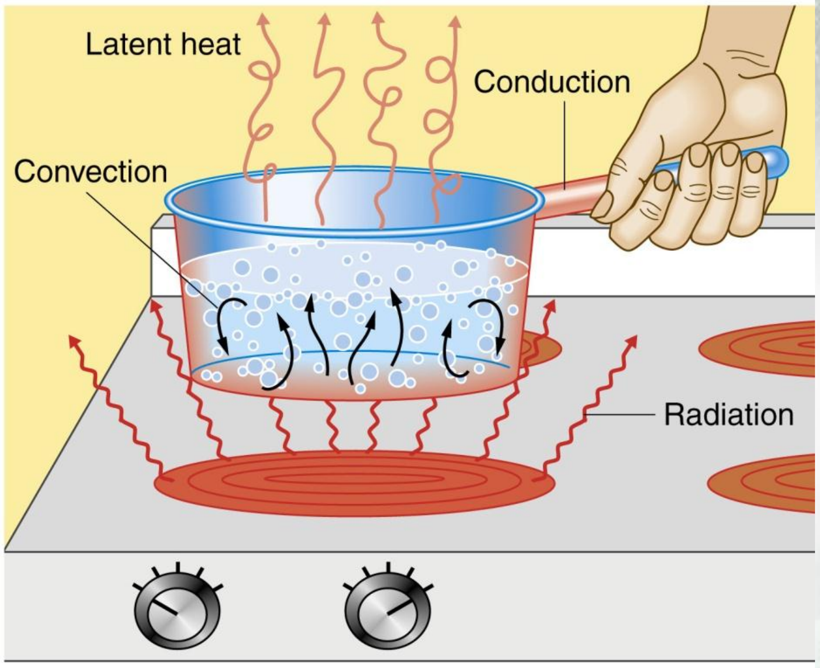
- The study of weather conditions on a small scale, as in the area immediately around a building, smokestack, or mountain.
- Scientific study of the turbulent atmospheric layer adjacent to ground. In order to understand processes such as (eg.) the location of maximum air pollutant concentration downstream from an industrial stack, or the pattern of pollination from a tree, (etc.), we require firstly to quantify the (fluctuating) wind field, and its interaction with such factors as the temperature stratification and terrain complexity.

Surface Energy Balance: Energy basics

- Energy: the ability to do work
- Many forms: electrical, mechanical, thermal, chemical, nuclear, …
- *Joule* (J): standard unit of energy (1 J= 0.239 calories)
- *Watt* (W): rate of energy flow (W = 1 J/s)

Methods of Energy Transfer

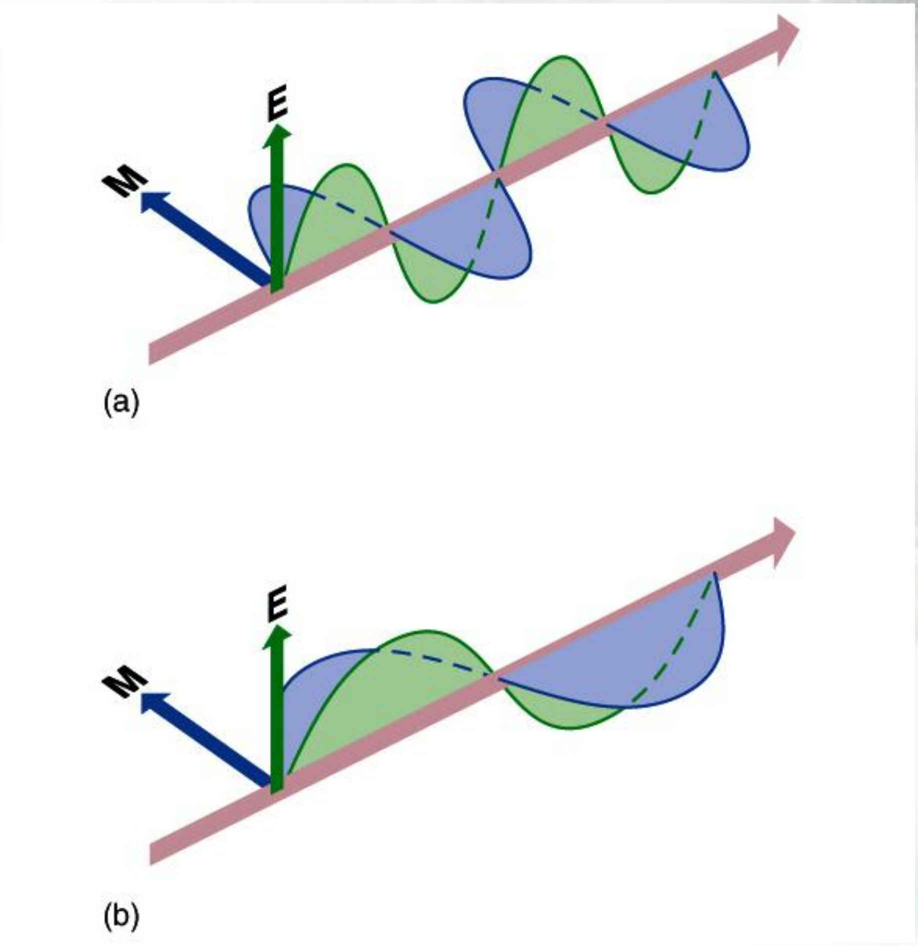
- **Conduction**
  - Molecule to molecule transfer
  - Heat flow: warm to cold
  - e.g. leather seats in a car



- **Convection**
  - transferred by vertical movement
  - physical mixing
  - e.g. boiling water
- **Radiation**
  - propagated without medium (i.e. vacuum)
  - solar radiation provides nearly all energy
  - The rest of this chapter deals with radiation

Radiation

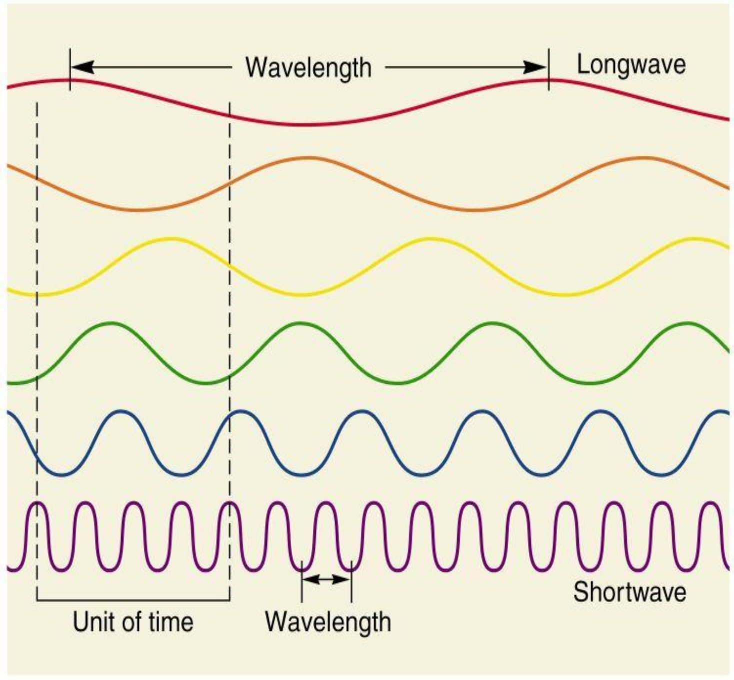
- Everything continually emits radiation
- Transfers energy in waves
- Waves are both electrical and magnetic, hence *electromagnetic radiation*



Radiation Quantity and Quality

•***Quantity:*** how much? → wave height **(amplitude).** Hotter bodies emit more energy than colder bodies

• ***Quality:*** what kind? → **wavelength:** distance btw. crest and crest (or trough and trough). generally reported in **μm** (microns)- one millionth of a meter. Hotter objects radiate at shorter wavelengths



• Travels at the speed of light (300,000 km/s). It takes 8 minutes for light from the Sun to reach Earth, and 4.3 years for light from the next nearest star, Proxima Centauri to reach us.

Wavelength of Sun and Earth Radiation

Sun =

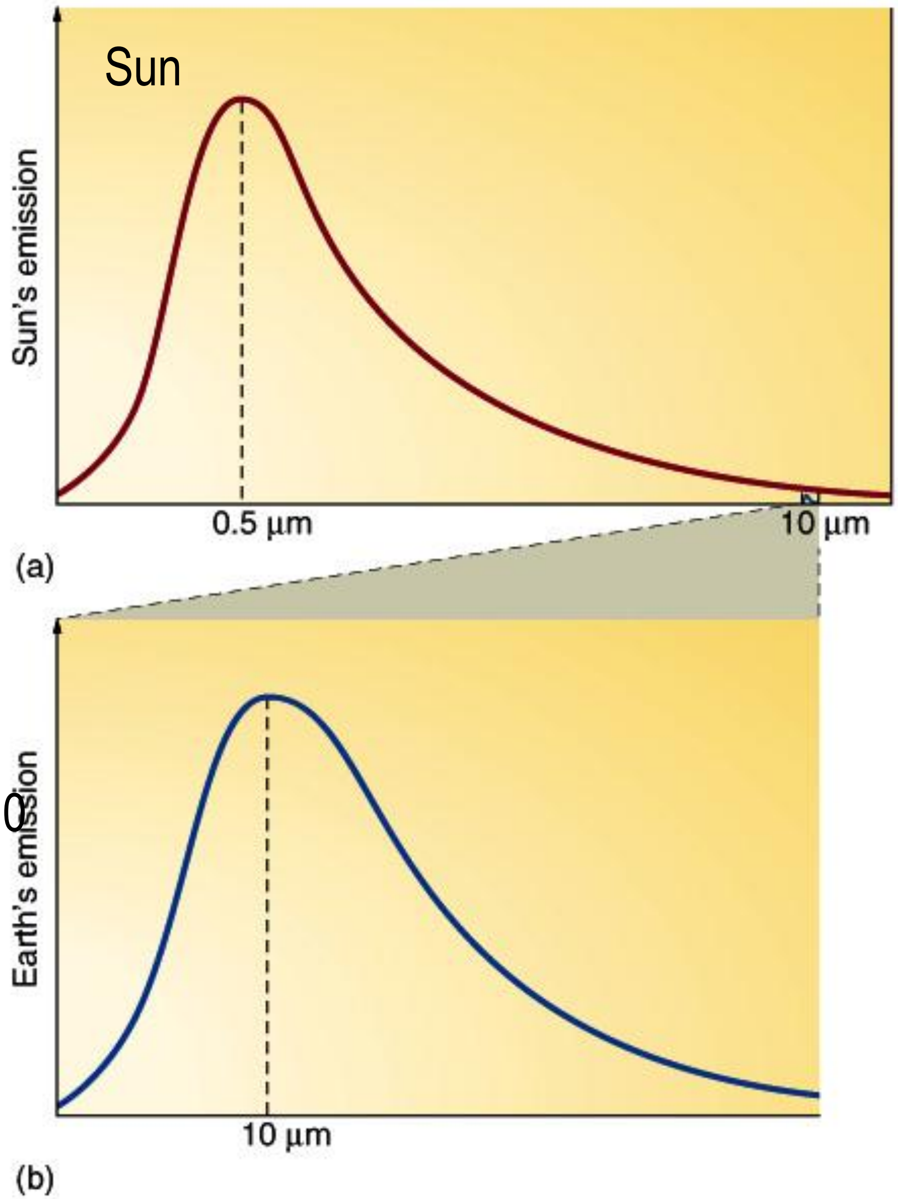
"shortwave"

(0.4-0.7 µm)
Peak 0.5 µm
(green)

Earth =

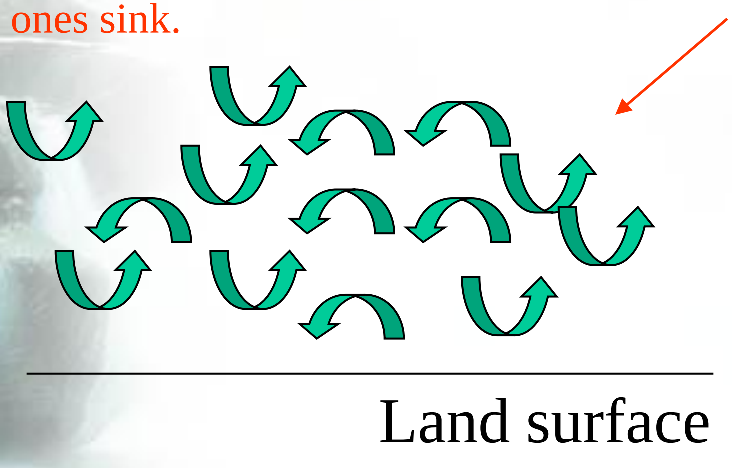
"longwave"

(4-100 µm) Peak 10 µm (infrared)



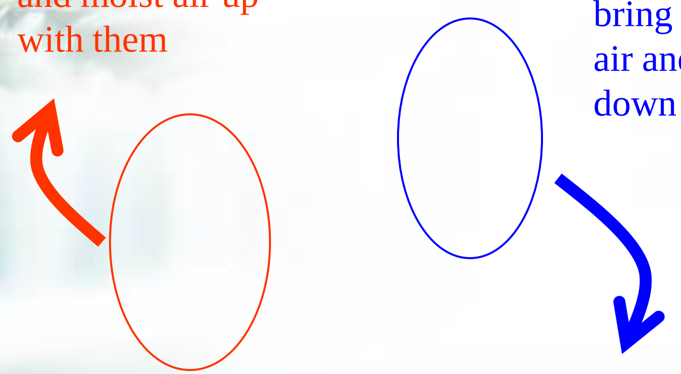
Sensible Heat flux

The air is full of eddies. Buoyancy tends to make warm pockets of air rise and cool ones sink.



Rising warm pockets bring both warm air and moist air up with them

Descending cool pockets bring both cool air and dry air down with them



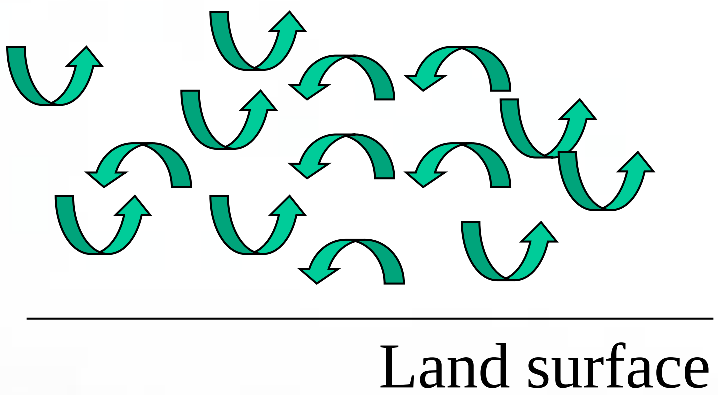
- Heat energy which is readily detected
- Magnitude is related to an object's specific heat
  - The amount of energy needed to change the temperature of an object a particular amount in J/kg/K
- Related to mass
  - Higher mass requires more energy for heating
- Sensible heat transfer occurs from warmer to cooler areas (i.e., from ground upward)

Sensible heat flux

- $Q_h = \rho\, C_d\, C_p\, V\, (T_{surface} - T_{air})$

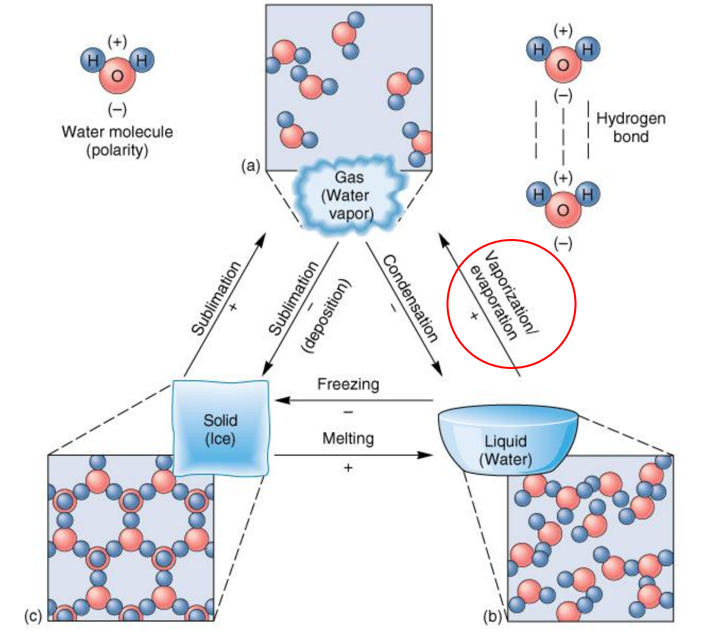
Where ρ is the air density, C_d is flux transfer coefficient, C_p is specific heat of air, V is surface wind speed, $T_{surface}$ is surface temperature, T_{air} is air temperature

- Magnitude is related surface wind speed
  - Stronger winds cause larger flux
- Sensible heat transfer occurs from warmer to cooler areas (i.e., from ground upward)
- C_d needs to be measured from complicated eddy flux instrument



Latent Heat flux

- Energy required to induce changes of state in a substance
- In atmospheric processes, invariably involves water
- When water is present, latent heat of evaporation redirects some energy which would be used for sensible heat
  - Wet environments are cooler relative to their insolation amounts
- Latent heat of evaporation is stored in water vapor
  - Released as latent heat of condensation when that change of state is induced
- Latent heat transfer occurs from regions of wetter-to-drier



Latent heat flux

- $Q_e = \rho\ C_d\ L\ V\ (q_{surface} - q_{air})$

Where ρ is the air density, C_d is flux transfer coefficient, L is latent heat of water vapor, V is surface wind speed, $q_{surface}$ is surface specific humidity, q_{air} is surface air specific humidity

- Magnitude is related surface wind speed
  - Stronger winds cause larger flux
- Latent heat transfer occurs from wetter to drier areas (i.e., from ground upward)
- C_d needs to be measured from complicated eddy flux instrument

Bowen ratio

- The ratio of sensible heat flux to latent heat flux

$$B = \frac{Q_h}{Q_e}$$

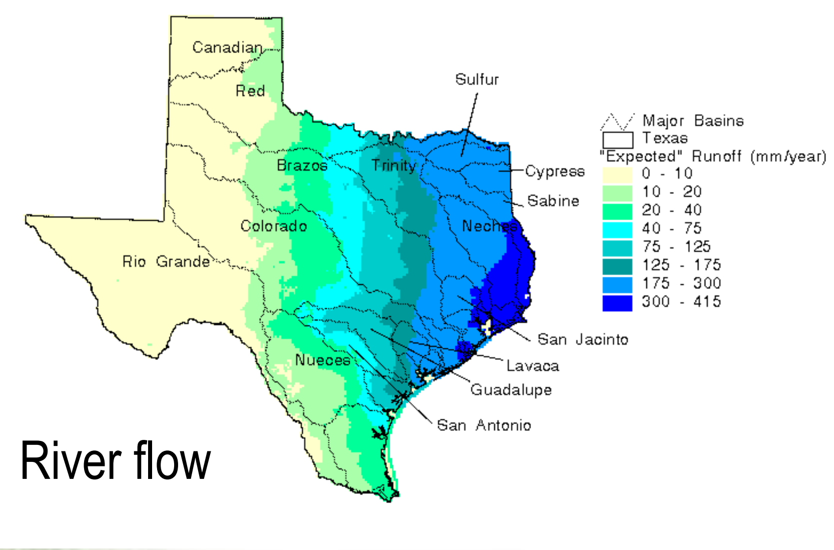
Where Q_h is sensible heat flux, Q_e is latent heat flux

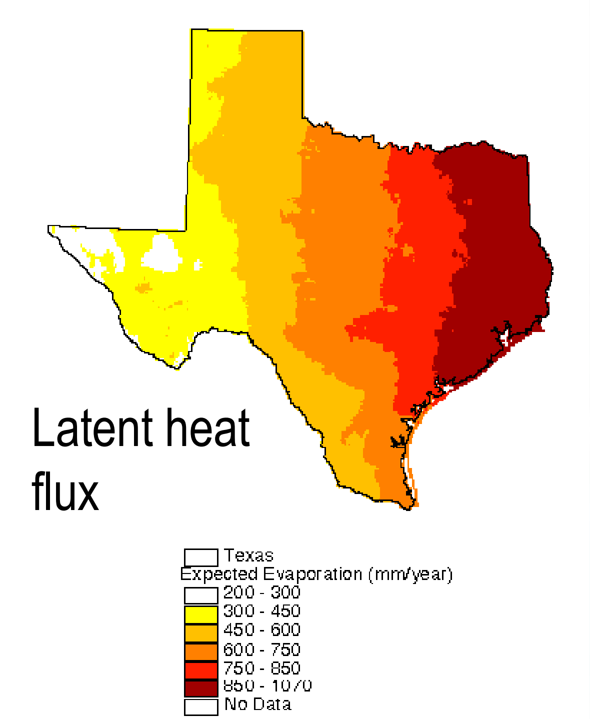
- $B = C_p(T_{surface} - T_{air}) / L(q_{surface} - q_{air})$ can be measured using simple weather station. Together with radiation measurements (easier than eddy flux), we can get an estimate of Q_h and Q_e
- When surface is wetter, B is smaller and less energy is released from the surface in the form of sensible heat flux, while more energy is released as latent heat flux
- Typical values:

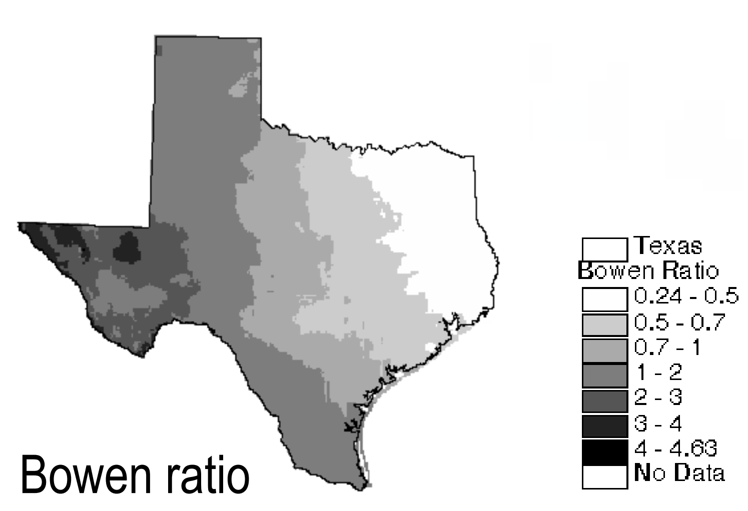
  Semiarid regions: 5
  Grasslands and forests: 0.5
  Irrigated orchards and grass: 0.2
  Sea: 0.1
  Some advective situations (e.g. oasis): negative

Map of Bowen ratio for Texas

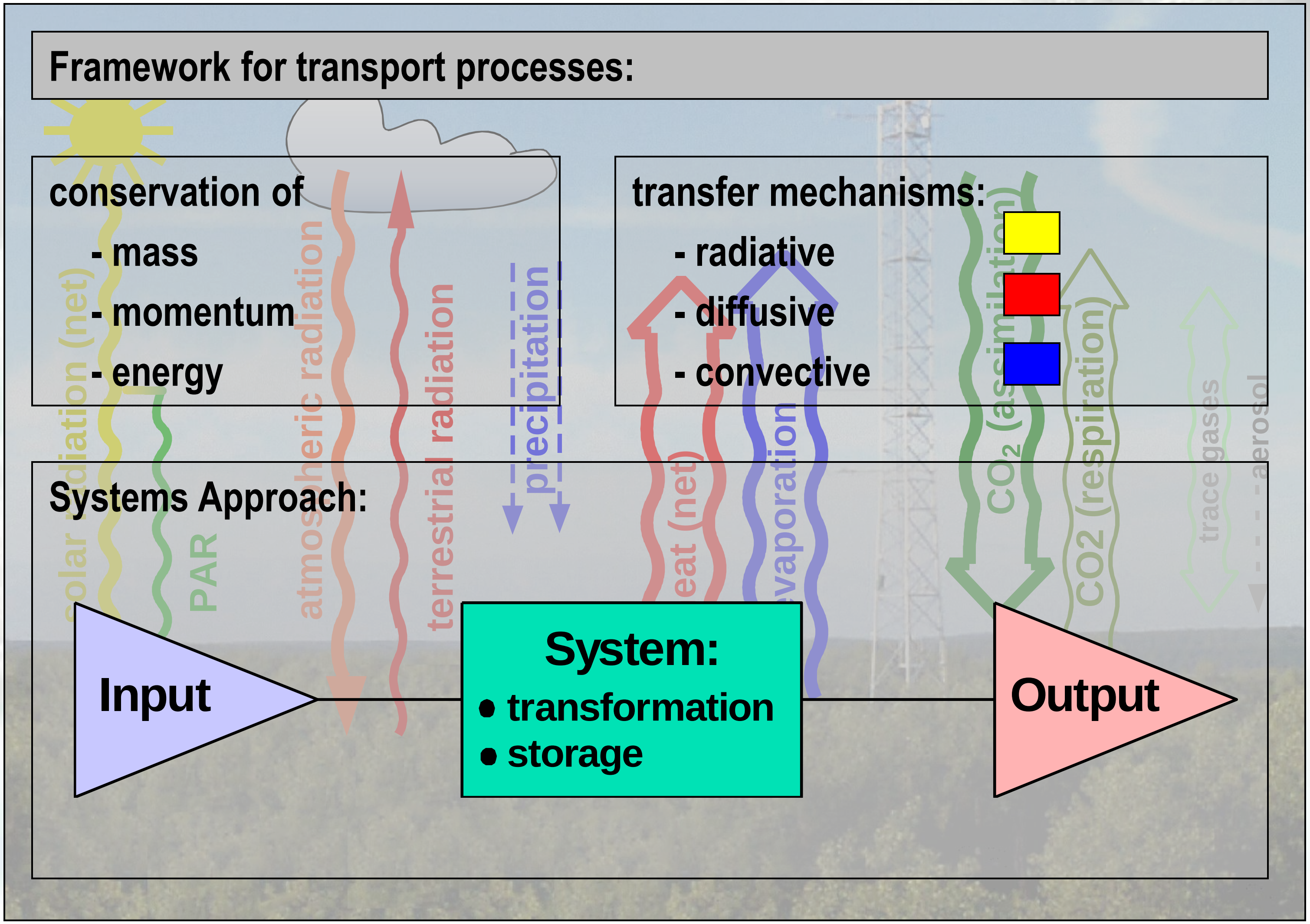
(By Prof. Maidment, U of Texas)





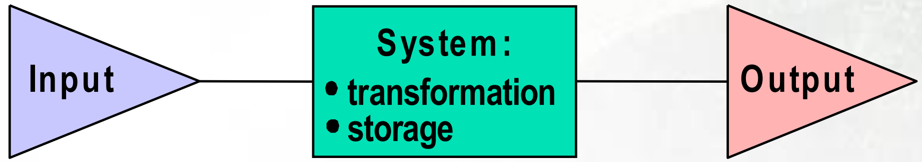


Transport Processes in Atmosphere – Biosphere Exchange



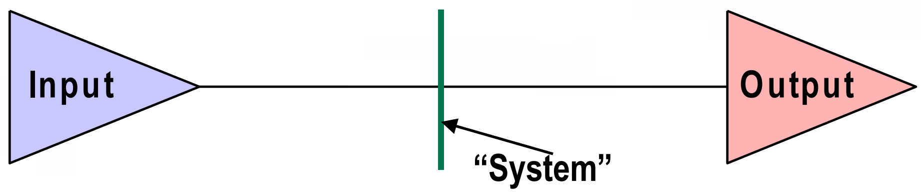
Systems Approach: 2 Possibilities

- "control volume" approach:



  - "system" is finite volume (e.g., a *box*)
  - "system" has storage capacity (mass, energy, momentum)
  - if system in *balance*: Input = Output ± Storage Change

- "control surface" approach:

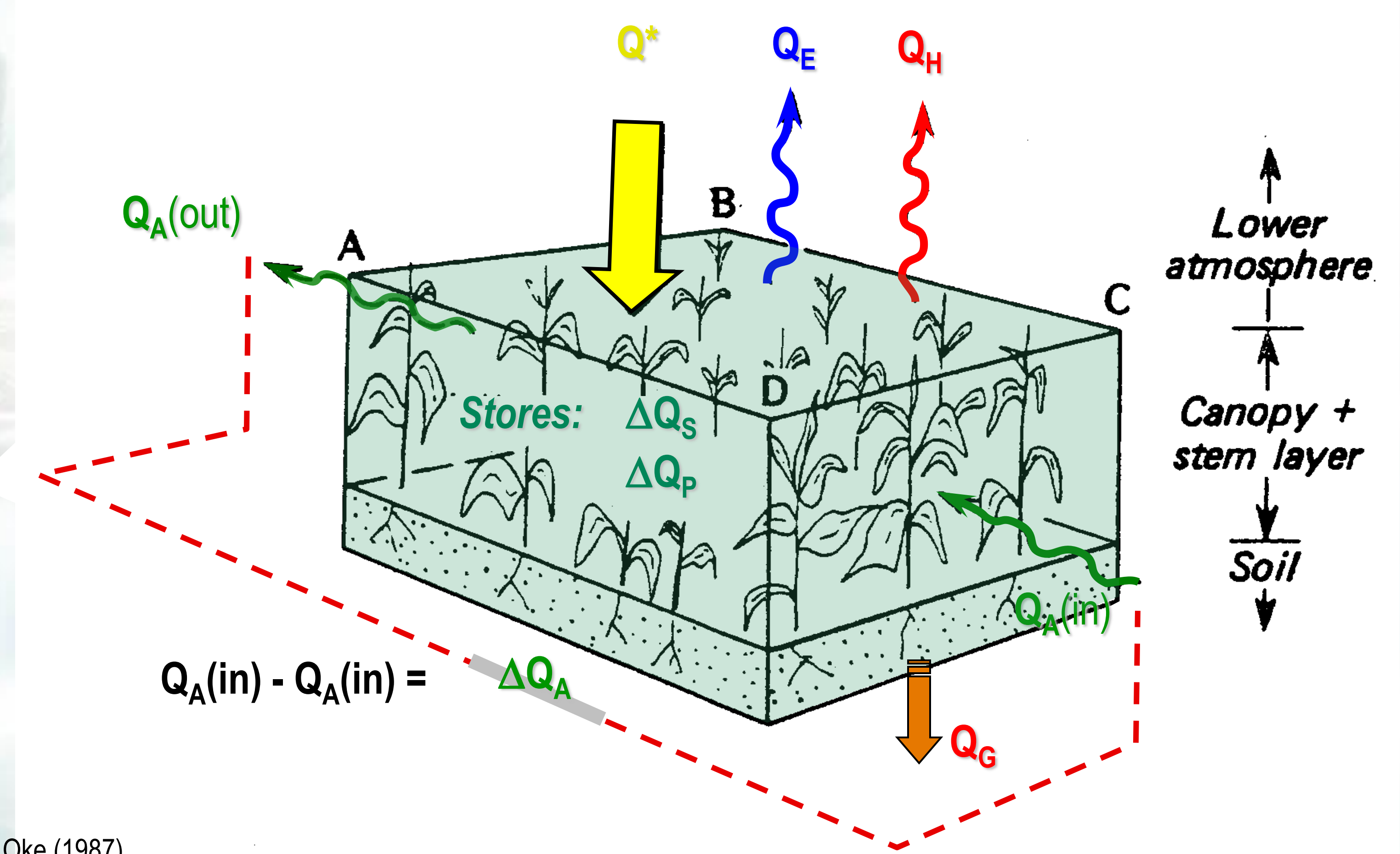


  - "system" is infinitesimally thin "volume" (i.e., an *interface*)
  - "system" has no storage capacity (mass, energy, momentum)
  - if system in *balance*: Input = Output

- **Plant-soil-atmosphere interface is geometrically complex: "control volume" is more convenient!**

Conservation of (thermal) Energy: Heat

Input:	Storage Change:	Output:
Q^* : net radiation	ΔQ_S : heat storage ΔQ_P : net photosynthesis ΔQ_A : advection	Q_E : latent heat flux Q_H : sensible heat flux (Q_G : ground heat flux)



modified, from Oke (1987)

Plant-Environment Interaction: Heat

- solar radiation for photosynthesis (PAR)
- net radiation (long and short wave) for temperature control (not too much, not too little)
- (sensible) heat to keep cool (not too much)
- latent heat (transpiration) to keep cool
- heat storage as buffer

Functional budget equation:

alternate notation!

$$\underbrace{K\downarrow - K\uparrow + L\downarrow - L\uparrow = Q^*}_{\text{radiation balance}} \quad \underbrace{R_n = H + LE + \Delta S + G + \Delta A + M_P}_{\text{energy balance}}$$

What are the biophysical controls of these transfer processes?

Radiation & Energy Budget: Controls

Process:		*Type:*	*Controls:*
K↓		rad	time of day, date, latitude, cloud cover, turbidity
K↑		rad	**K↓,** albedo → plant geometry, density, condition, species → ground cover
L↓		rad	temperature and humidity structure of the atmosphere, cloud cover, synoptic conditions
L↑		rad	temperature of vegetation, soil surface (material → ε)
Q_H	**H**	conv	temperature difference between vegetation and air, turbulence, leaf area
Q_E	**LE**	conv	vapor pressure deficit, turbulence, H_2O availability, energy availability (for evaporation), leaf area, plant growth state
ΔQ_S	**ΔS**		temperature difference between biomass and air/soil, thermal properties of biomass/soil/air
Q_G	**G**	diff	temperature difference between biomass and air/soil, thermal properties of biomass/soil
ΔQ_A	**ΔA**	conv	temperature difference between biomass and upwind areas, wind velocity
Q_P	**M_P**		PAR availability, leaf area, CO_2 availability, temperature, water availability (in cells), nutrient availability, health of plant, age of leaves.

Conservation of Water

Input:

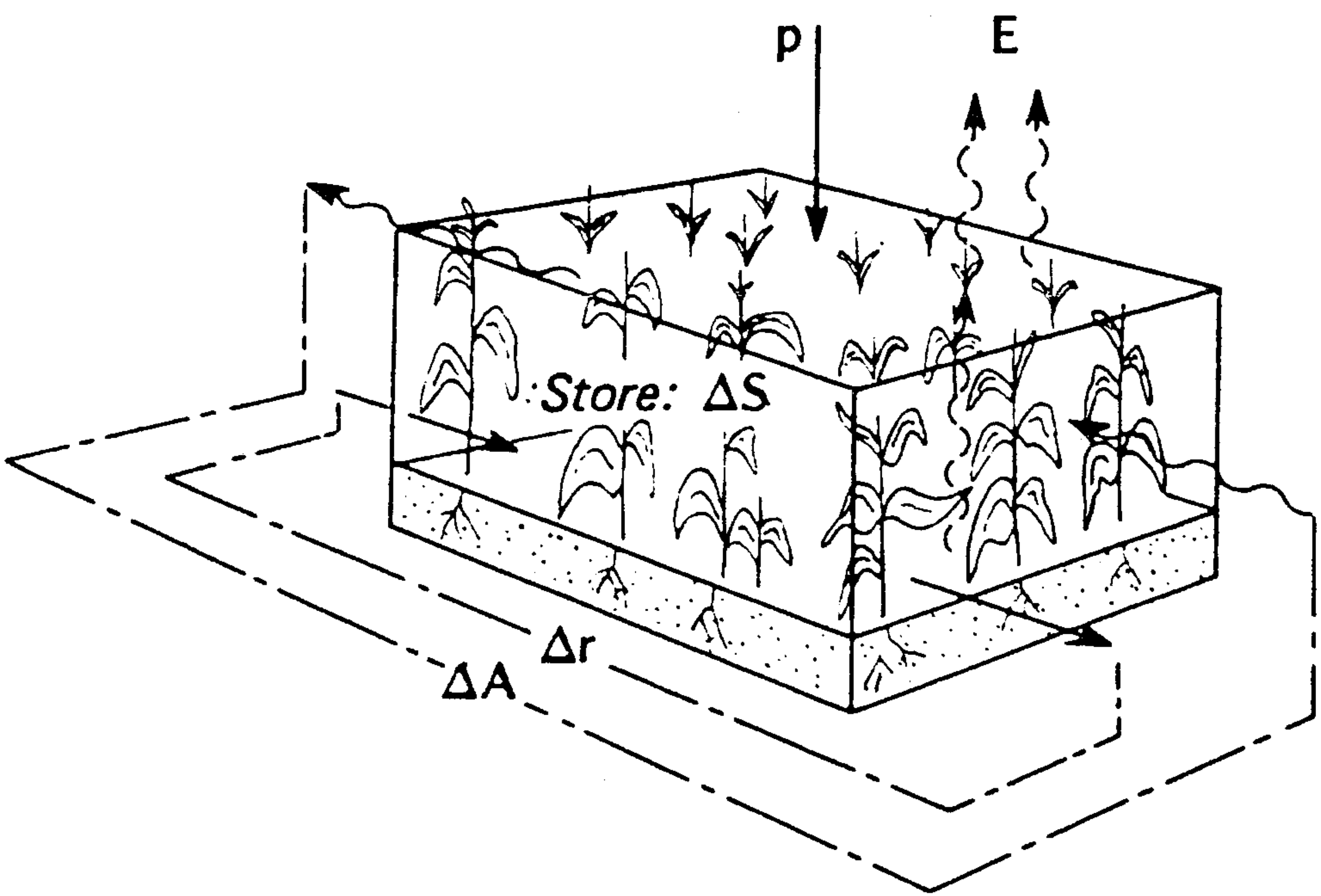
P : precipitation

Storage Change:

ΔS: surface water change
Δr : net runoff
ΔA : vapor advection

Output:

E : evapotranspiration
= *evaporation* + *transpiration*



from Oke (1987)

Plant-Environment Interaction: Water

- evaporation/transpiration to keep cool
- dew/frost to keep warm
- precipitation to re-stock soil water for root uptake (but not so much to "drown" soil microbes)
- storage in biomass to maintain turgor, and as solvent and means of transport for nutrients and metabolic products

Functional budget equation:

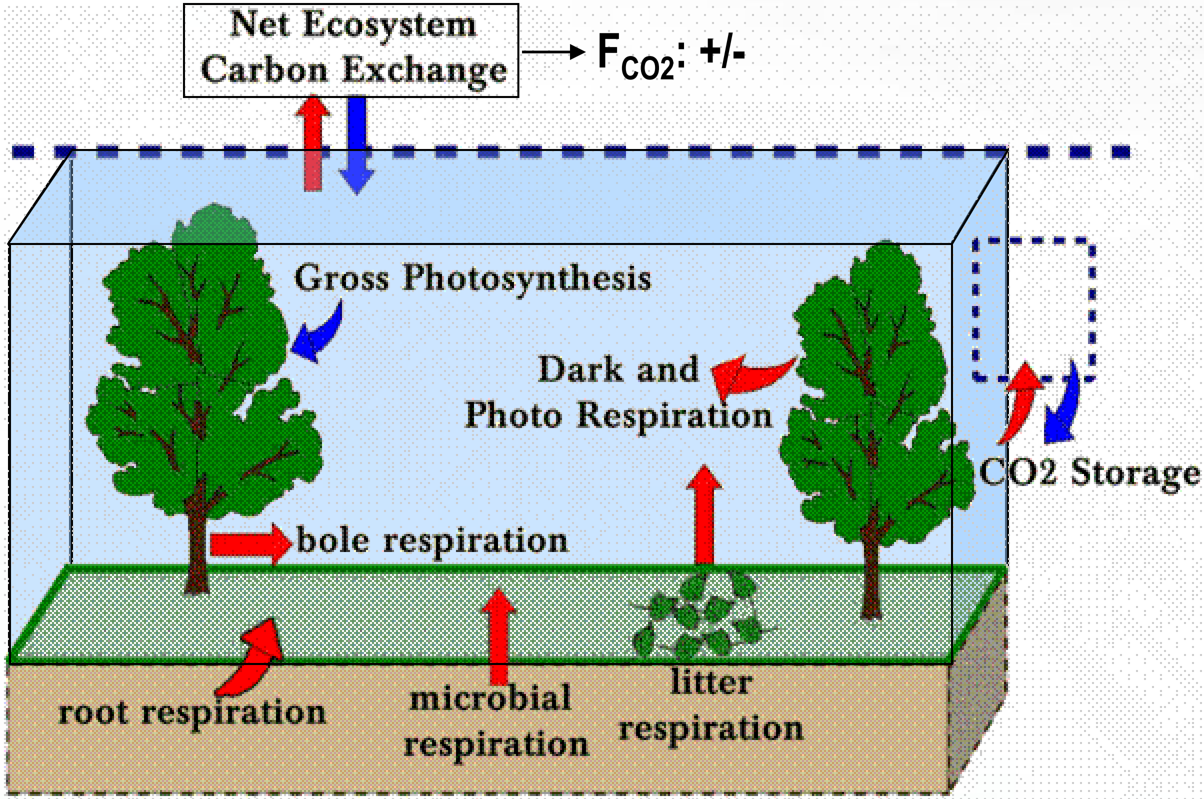
$$Q_E/L = E = P - R - \Delta S - I$$

Process:	*Type:*	*Controls:*
E	conv	(evaporation + transpiration), same controls as $\mathbf{Q_E}$
P		synoptic conditions
R		topography, soil saturation (hydrologic history), P-intensity, ground water hydraulics
ΔS		same control as R – reverse effect
I		leaf area, leaf geometry and texture, canopy architecture, P-intensity and -history

Conservation of Carbon (CO_2)

Control volume: ***air volume in and around canopy***

Input:	**Storage Change:**	**Output:**
$F_{CO2} < 0$: net sink R_A : autotr. respiration R_H: heterotr. respiration	ΔC_S: CO_2 conc. change ΔC_A : CO_2 advection	$F_{CO2} > 0$: net source A_C : CO_2 assimilation



Plant-Environment Interaction: CO_2

- Photosynthesis of CO_2 (assimilation)
- Autotrophic respiration
  - photorespiration (cost of assimilation)
  - dark respiration (cost of plant growth and maintenance)
- Heterotrophic respiration (leaf litter, woody debris, soil microbes and fungi)
- Net CO2 flux (net ecosystem exchange)

Functional budget equation:

$$F_{CO2} = -V_{Cn} + R_P + R_G + R_M + R_F + R_{Sr} + R_{Sm} - \Delta C_S - \Delta C_A$$
$$= -A_C + R_E - \Delta C$$

- $V_n + R_P = A_C$
- $R_D + R_P + R_{Sr} = R_A$
- $R_G + R_M = R_D$
- $R_F + R_{Sm} = R_H$
- $R_G + R_M + R_F + R_{Sr} + R_{Sm} = R_E$
- $\Delta C_S - \Delta C_A = \Delta C$

Plant-Environment Interaction: CO_2

Notation: all in [μmol m^{-2} s^{-1}]

F_{CO2}	net carbon flux
V_{Cn}	net carbonation rate (carbonation - oxygenation)
R_P	photorespiration
A_C	carbon assimilation rate
R_G	growth respiration
R_M	maintenance respiration
R_D	dark respiration
R_{Sr}	root respiration
R_A	autotrophic respiration
R_F	forest floor respiration (leaf litter, debris)
R_{Sm}	soil microbial respiration
R_H	heterotrophic respiration
R_E	ecosystem respiration

complex system with many component processes

CO_2 Exchange: Controls

Process:	***Type***	***Controls***:
F_{CO_2}	conv	CO_2 concentration in air, turbulence, all other terms
A_C	diff	PAR availability, temperature, nutrient (N) availability, water availability, leaf area, plant type (C_3, C_4, CAM), growth stage, plant health
R_P	diff	plant type (C_3, C_4, CAM), assimilation rate
R_G	diff	assimilation rate
R_M	diff	plant size (biomass dry weight), temperature
R_H	diff	soil temperature, soil moisture, soil nitrogen
ΔC		ventilation, vertical mixing, stability, horizontal homogeneity

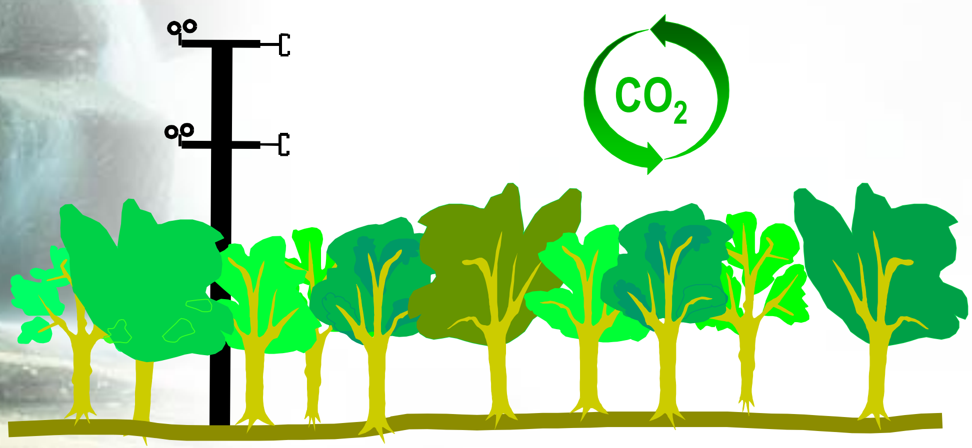
Other Significant Vegetation-Atmosphere Exchanges:

- nitrogen compounds (nutrients)
- aerosol (sulphur, nitrogen)
- pollutants (e.g., ozone)
- biogenic VOCs (e.g., isoprene, terpenes)

Plant-Environment Interaction: CO_2

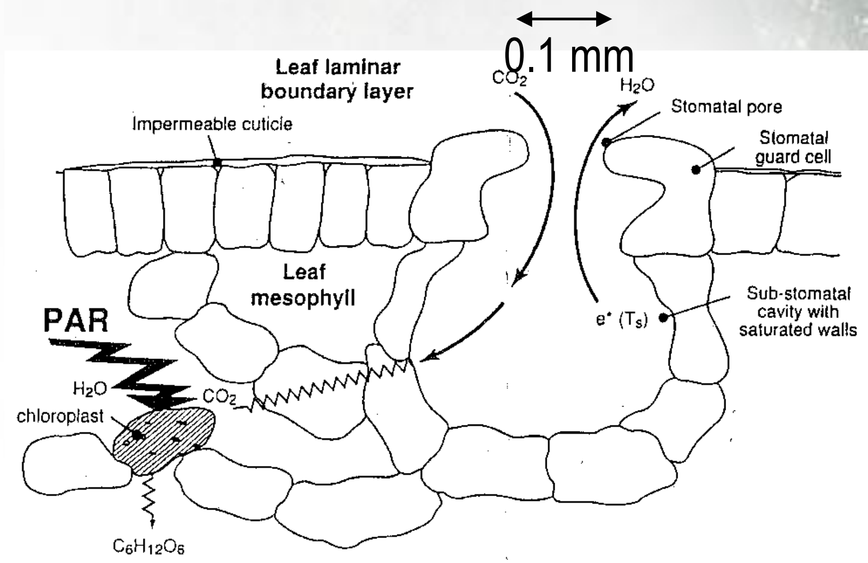
Scale of Approach

Macroscopic Approach



- ecosystem exchange
- transport
- 10^2 - 10^3 m
- hourly – multi-year

Microscopic Approach



- intercellular exchange
- transformation, chemical pathways
- 10^{-5} – 10^{-2} m
- seconds – hourly

everything in between

Energy Budget at land surface

Energy budget of a surface layer without vegetation:

$$\frac{\partial E_s}{\partial t} = G = R_n - LE - H - \Delta F_{out}$$

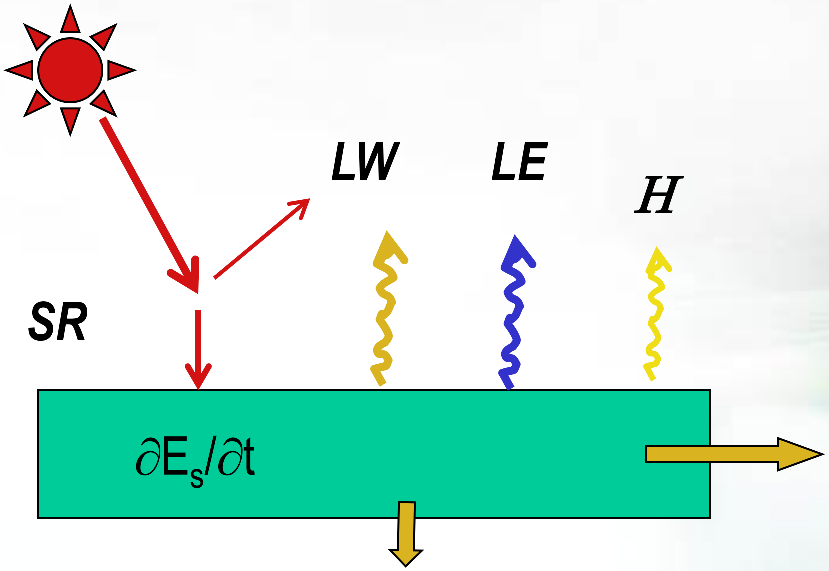
where G : heat storage of the surface layer

$\mathrm{R_n}$: surface net radiation, $\mathrm{R_n} = SR - LW$

LE : surface latent flux

H : surface sensible flux

$\Delta \mathrm{F_{out}}$: *divergence* of energy below the surface



The surface layer

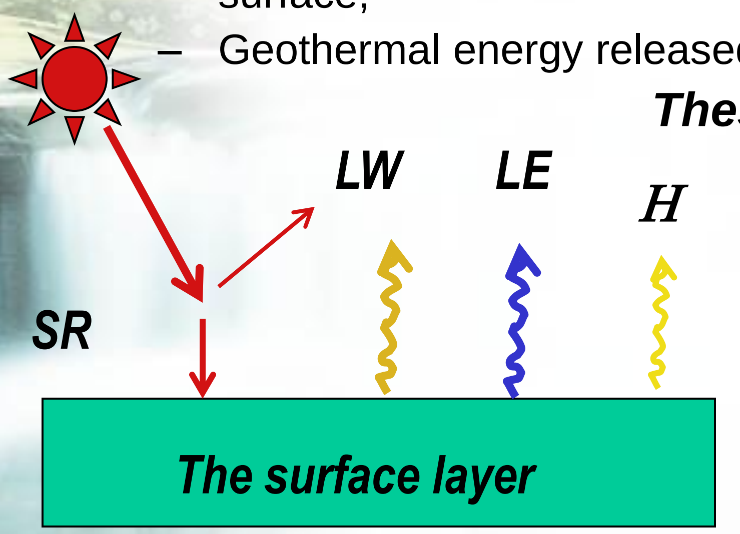
Energy Budget at land surface

- Under steady-state and when the energy storage is small, the energy budget of a land surface layer can be simplified to

  R_n=LE-H because usually $\Delta F_{out} << R_n$, LE, H

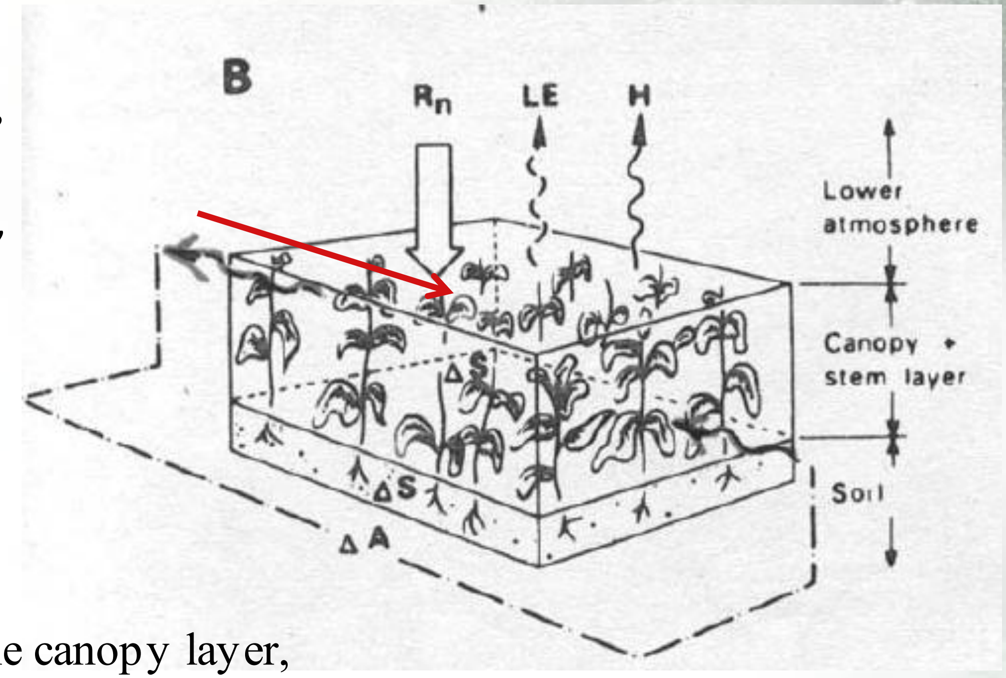
- Factors left out but can be important for some locations and periods:
  - Fraction of solar energy that is stored in the chemical bones formed during photosynethesis, can reach 5% during growing period (?);
  - Heat released by oxidation of biological substance as in biological decade or biomass burning;
  - Heat released by fossil fuel burning or nuclear power generation (e.g., big urbane areas);
  - Convection of the kinetic energy of winds into thermal energy;
  - Heat transferred by precipitation, especially when precipitation is much cooler than the surface;
  - Geothermal energy released in hot springs, earthquakes and volcanoes.

These factors are generally unimportant globally.



Surface with vegetation:

- ***Vegetation will contributes to or dominate the surface albedo. The top of the surface layer is at the canopy level. Energy divergence and energy storage in the canopy layer can be significant, especially on scale shorter than daily.***



$$\frac{\partial E_c}{\partial t} = R_n - LE - H - G_o + \mathrm{D_h}$$

where $\mathrm{E_c}$, R_n, $L\mathrm{E}$, H, $\mathrm{G_o}$ are energy of the canopy layer, net radiation, latent and sensible flux at the canopy layer, the heat flux from the canopy - stem layer to ground, respectively.

$\mathrm{D_h}$ is a horizontal energy advection

Where $\frac{\partial \mathrm{E_c}}{\partial t} = S_T + S_q + S_{veg}$ ← Sensible, latent heat of the air column & change of heat content of vegetation

$$(S_T, S_q, S_{veg}) = \int \left[\partial(\rho C_p T, \rho C_p q, \rho_{veg} C_{veg} T_{veg}) / \partial T\right] dz$$

Physical Properties of Soil

	Specific heat (c_p) (J kg^{-1} K^{-1})	Density (ρ) (kg m^{-3})	ρc_p (J m^{-3} K^{-1})
Soil inorganic material	733	2600	1.9×10^6
Soil organic material	1921	1300	2.5×10^6
Water	4182	1000	4.2×10^6
Air	1004	1.2	1.2×10^3

minerals organics water ice air

$$C_s \ (Jm^{-3}K^{-1}) = \rho_s c_s f_s + \rho_c c_c f_c + \rho_w c_w f_w + \rho_i c_i f_i + \rho_a c_a f_a$$

heat capacity of soil

density

heat capacity

volume fraction

Soil is a mixture of several materials, each with quite different physical properties

- Soil temperature and heat flux:
- Heat flux

If we neglect the latent flux in the soil layer

$$G = -K_s \frac{\partial T_s}{\partial z'}$$

where $K_s = \rho_s c_s \kappa_s$, κ_s is the thermal diffusivity

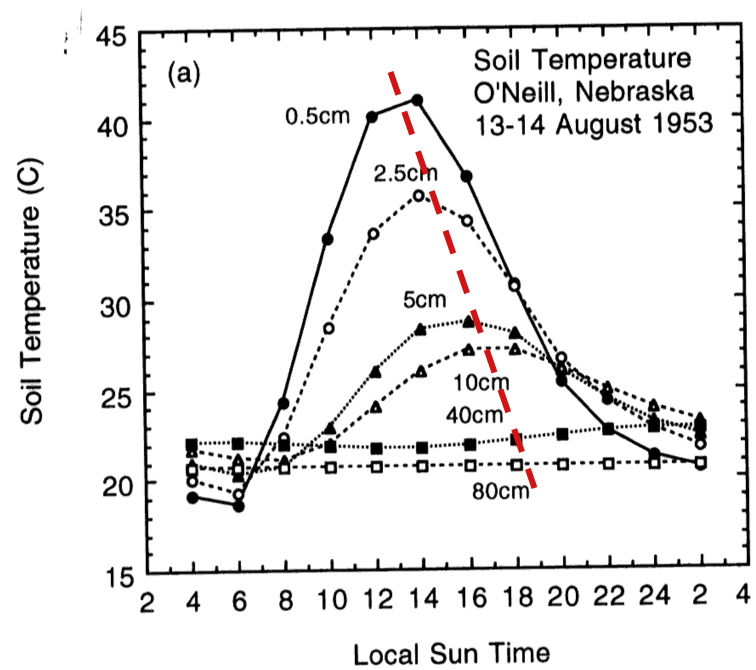
For a periodic forcing of period τ, e.g., diurnal, seasonal, the response of $T(z)$ is also periodic, but damped and delayed with depth relative to surface forcing.

- Soil temperature:

The change of soil temperature :

$$\frac{\partial T_s}{\partial t} = -\frac{1}{\rho_s c_s}\frac{\partial G}{\partial z'} = \kappa_s \frac{\partial^2 T_s}{\partial z'^2}$$

where z': depth of the surface soil layer



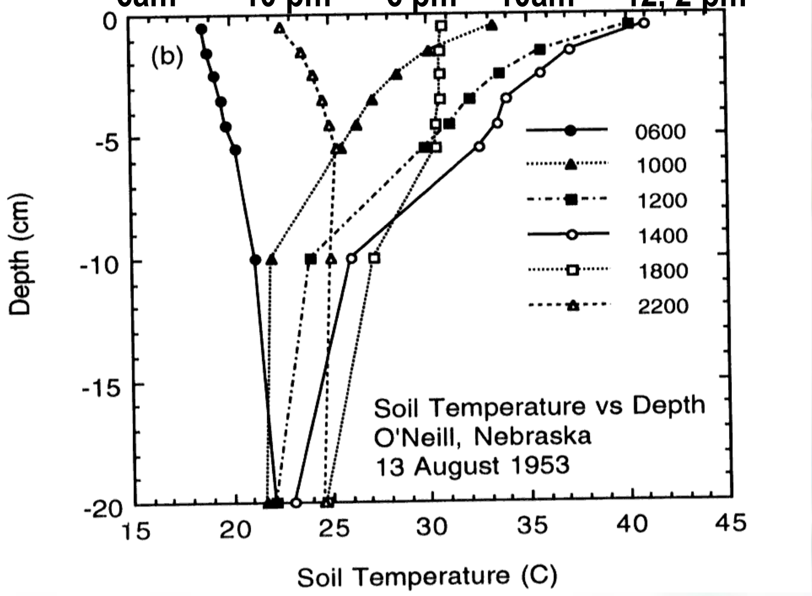
– For $2\kappa_s$~ 5 x 10^{-7} m^2 s^{-1},
  - diurnal cycle, $\tau = 1$ day, D~0.1 m
  - annual cycle, $\tau \sim 1$ year, D~1.5 m
  - glacial-interglacial, $\tau = 10{,}000$ years, D~150 m

- In reality, changes are not pure sinusoidal, use

$$\frac{\partial T_s}{\partial t} = \kappa_s \frac{\partial^2 T_s}{\partial z'^2}$$

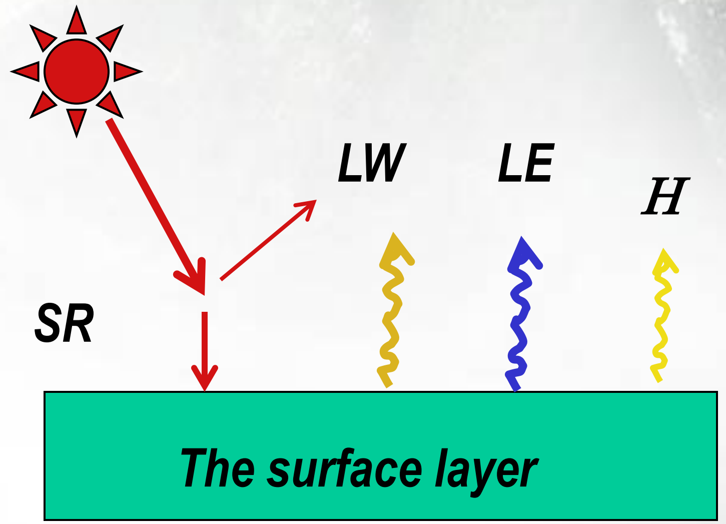
- 25 K diurnal cycle at 0.5 cm, Max T around 2 PM
- Only 6 K diurnal range at 10 cm', Max T about 6 PM: Damped and delayed with depth
- Negligible diurnal cycle at 50 cm

6am 10 pm 6 pm 10am 12, 2 pm



The surface fluxes:

The net radiative flux (R_n)
The latent flux (LE)
The sensitive flux (H)



R_n: The Radiative fluxes

- $R_{n,o}$ = Solar flux↑ - longwave flux↓

  $= SR(1-\alpha_s) + (1-\varepsilon_s)LW_{DN,o} - \varepsilon_s \sigma T^4$

- α_s: surface albedo
- $\varepsilon_{s:}$ surface emissivity.

- The surface albedo, emissivity changes with surface type, wave frequency.
  - Surface Albedo: $\alpha_{s,\lambda}$, the ratio of reflected vs. incoming solar flux (integrated over all direction of the hemisphere).
  - Total surface solar albedo:

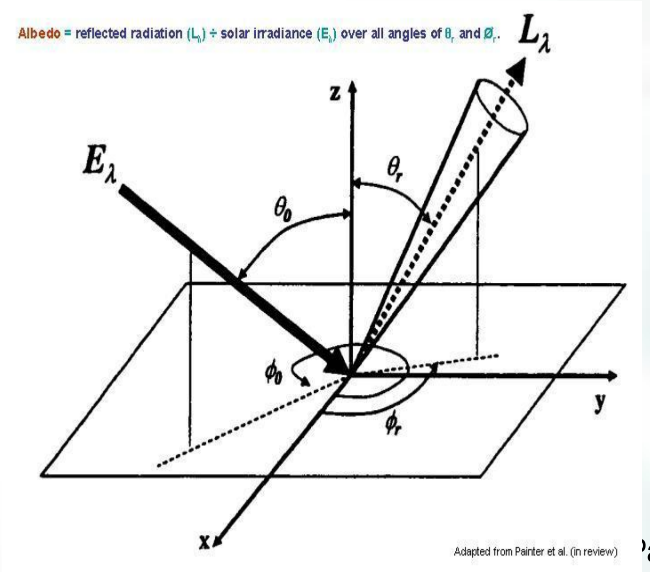
α_s integrates for all solar spectrum

$$\alpha_s = \int \alpha_{s,\lambda} d\lambda$$

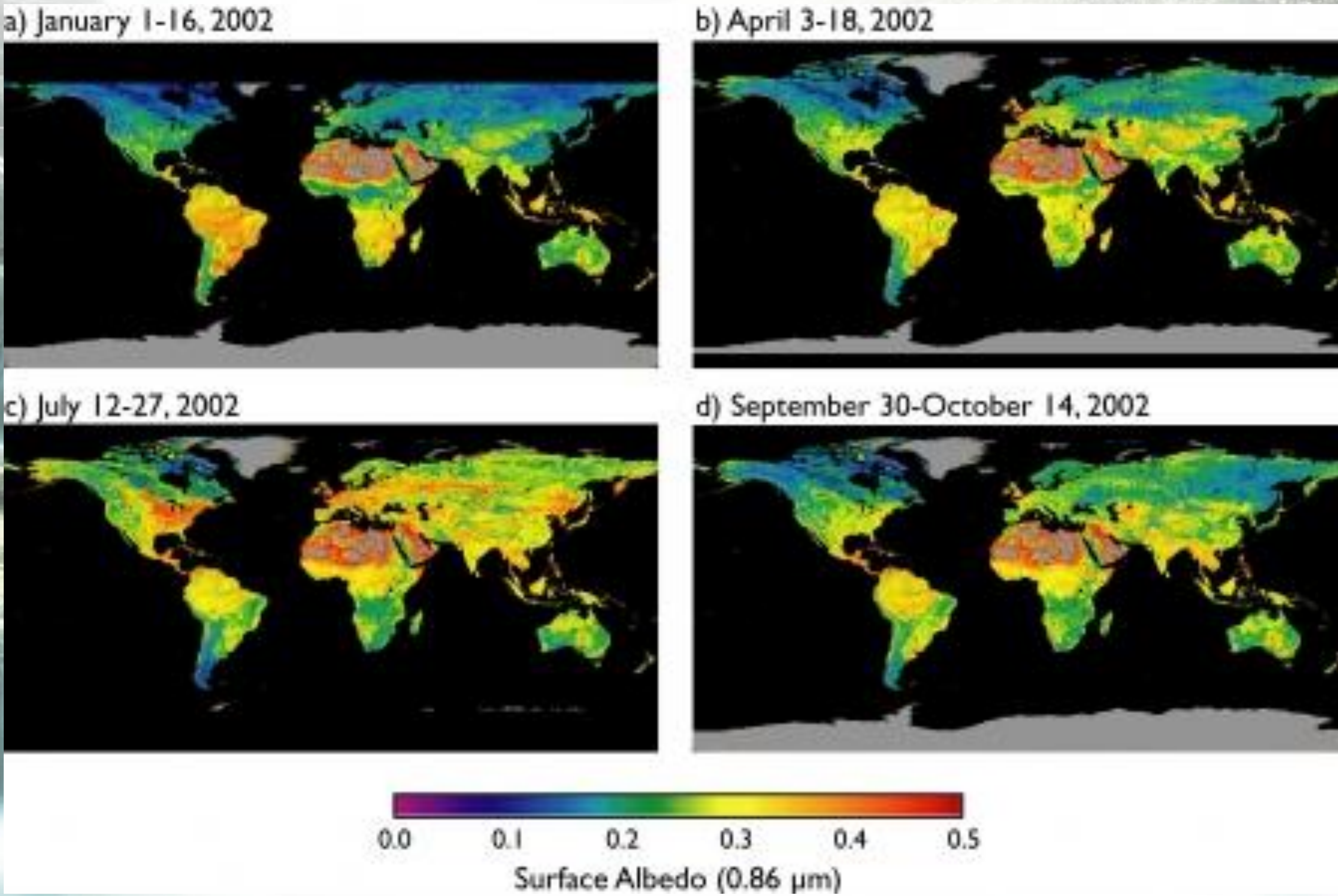
$$\alpha_{s,\lambda} = \iint \frac{L_\lambda}{E_\lambda} d\theta d\phi$$

θ : *local* zenith angle,

ϕ : azimuthal angles



Global land surface albedo:



Surface Albedo (percent)

Surface type	Range	Typical value
Water		
Deep water: low wind, low altitude	5–10	7
Deep water: high wind, high altitude	10–20	12
Bare surfaces		
Moist dark soil, high humus	5–15	10
Moist gray soil	10–20	15
Dry soil, desert	20–35	30
Wet sand	20–30	25
Dry light sand	30–40	35
Asphalt pavement	5–10	7
Concrete pavement	15–35	20
Vegetation		
Short green vegetation	10–20	17
Dry vegetation	20–30	25
Coniferous forest	10–15	12
Deciduous forest	15–25	17
Snow and ice		
Forest with surface snowcover	20–35	25
Sea ice, no snowcover	25–40	30
Old, melting snow	35–65	50
Dry, cold snow	60–75	70
Fresh, dry snow	70–90	80

- Snow and ice brightest
- Deserts, dry soil, and dry grass are very bright
- Forests are dark
- Coniferous (cone-bearing) needleleaf trees are darkest

Table 4.3

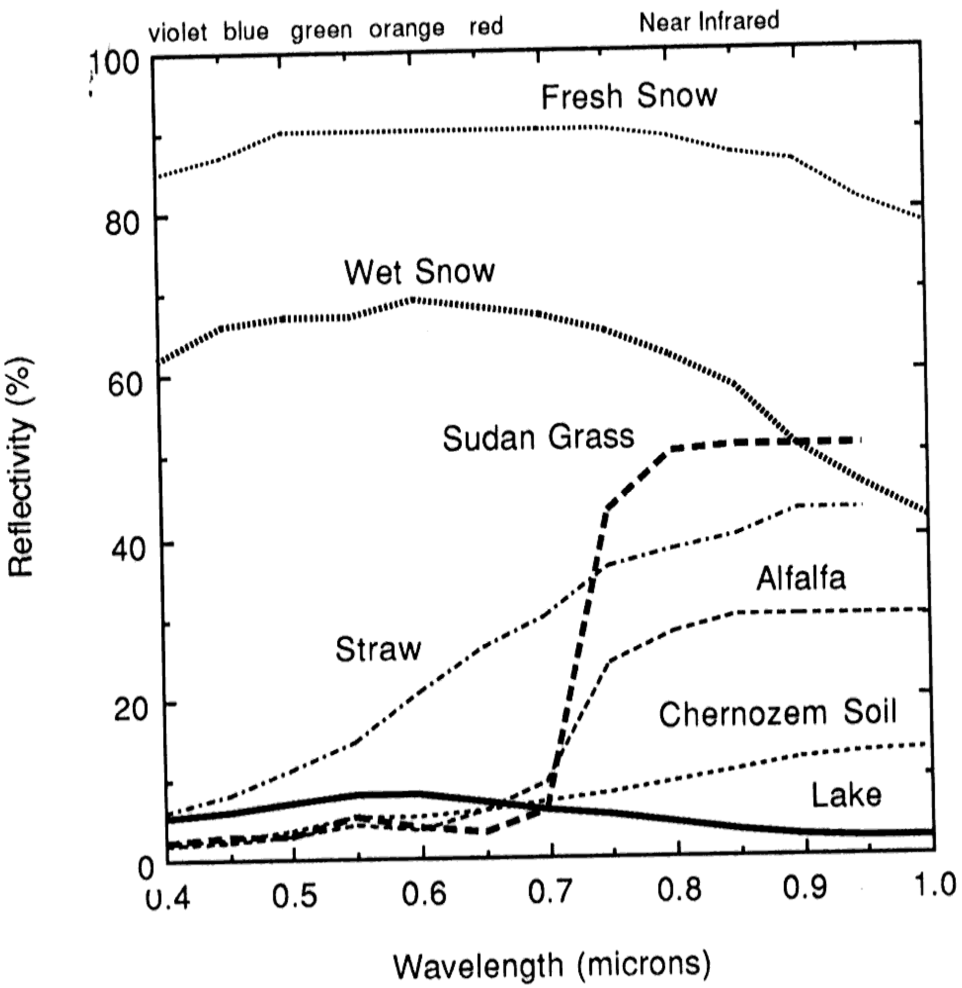
Albedos for Dry and Moist Soil Surfaces

	Even surface		Tilled surface	
	Dry	Moist	Dry	Moist
Chernozem of dark gray color	13	8	8	4
Light chestnut soil of gray color	18	10	14	6
Chestnut soil of grayish red color	20	12	15	7
Gray sandy soil	25	18	20	11
White sand	40	20	—	—
Dark blue clay	23	16	—	—

[From Mironova (1973).]

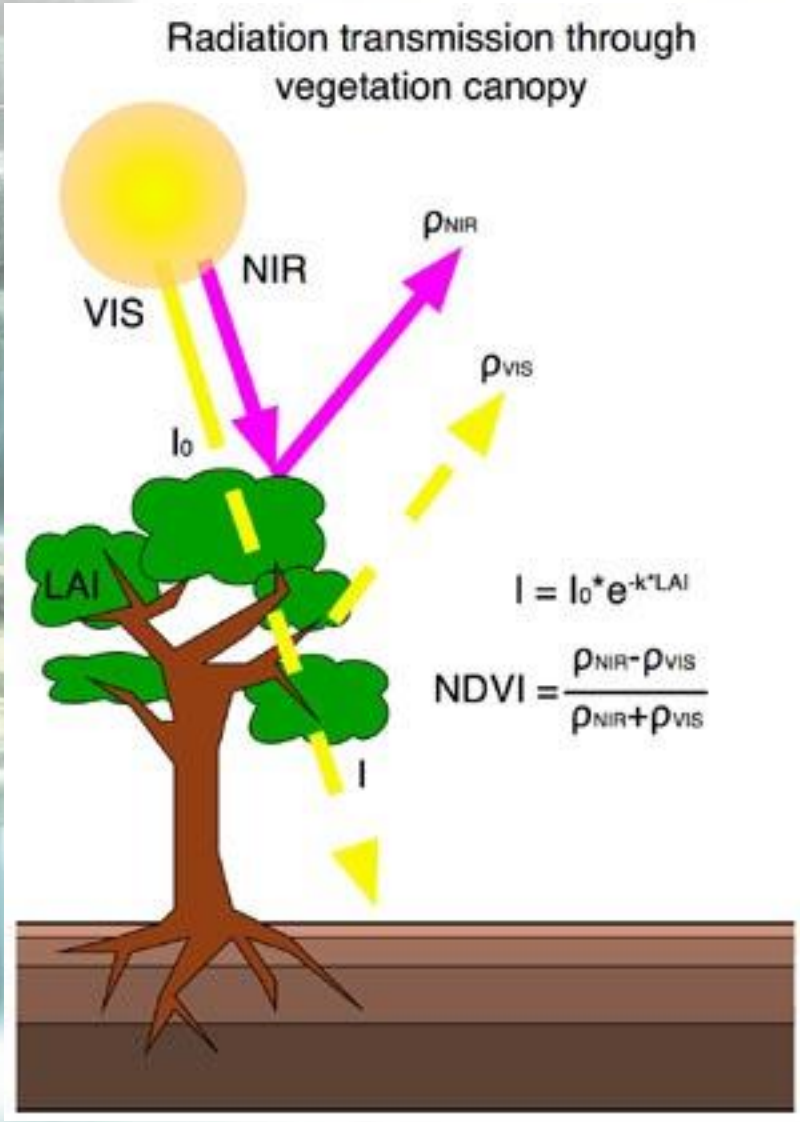
Because surface albedo is highly variable and has a strong effect on absorbed solar radiation, it can have a large effect on surface temperature. Surface albedo can also have a strong effect on the sensitivity of climate, if it changes systematically with climatic conditions. Feedback processes involving surface albedo are discussed in Chapter 9.

Spectrum dependent Land-Surface Albedo



- Strong wavelength dependence over vegetated land
- Plants absorbs strongly (upto 90%) of the solar radiation between 0.3-0.7 mm for photosynthesis (photosynethetically active radiation or PAR), reflect strongly in near-IR to avoid heating.
- Total albedo is a weighted mean over wavelengths

Solar radiation at the canopy:



- Visible solar (VIS) is less reflected and can penetrate deeper in canopy; whereas the near infrared solar (NIR) is more reflected by canopy;
- Thus, $\rho_{nir}/\rho_{vis} \downarrow$ with depth below top of the plants;
- Larger difference between ρ_{nir} and ρ_{vis}, i.e., the large NDVI (Normalized Differential Vegetation Index), indicates large biomass of the plants.
- Cavities and different leaf orientations of the canopy can lead to a lower canopy albedo compared to a single leaf.
- The absorption of PAR depends on the leaf area index (LAI).

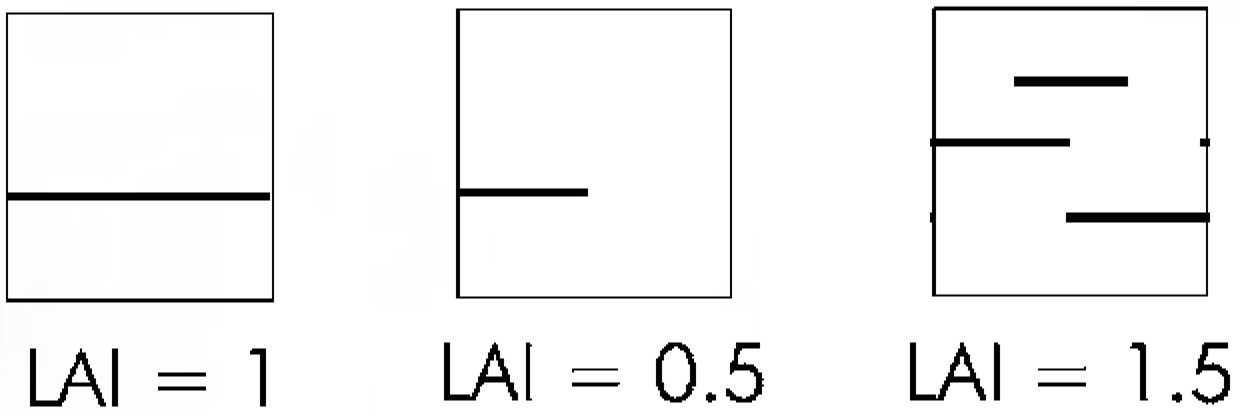
Leaf Area Index

<u>Leaf Area Index</u> *is the fraction of the surface area covered by leaf surfaces when viewed from above.*

No leaves at all: LAI = 0.

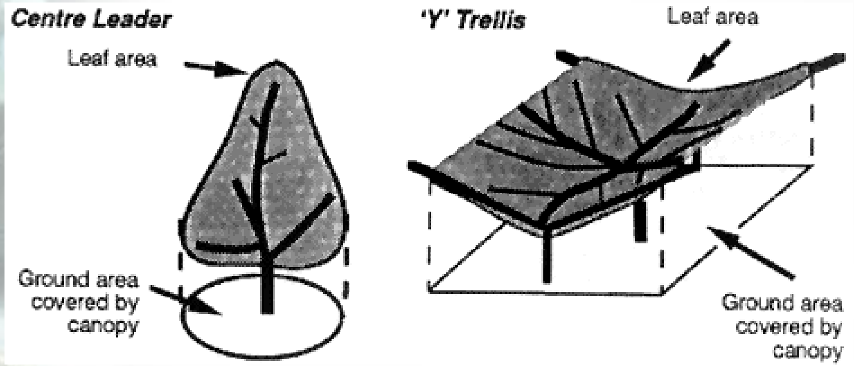
Imagine one giant leaf, covering everything like a blanket. The LAI would be 1.

An actual canopy has multiple leaves overlying any point of the surface. The LAI can exceed 1. In fact, for very dense canopies, it may be 5, 6, 7, or more. The LAI over an area is the average number of horizontal surfaces intercepted while traveling down through the canopy to the top to the soil:

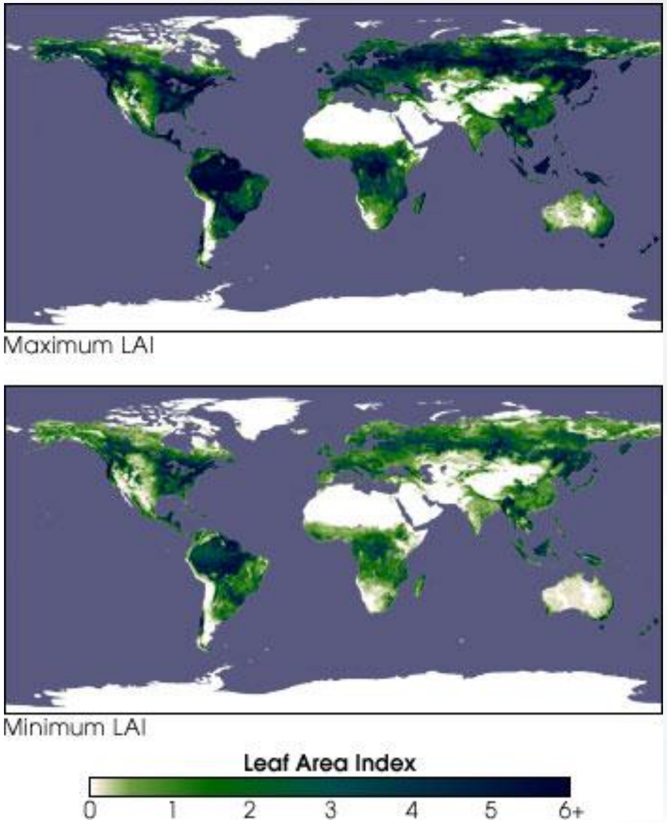


The leaf area index, LAI:

- LAI: The ratio of total canopy leaf surface area vs. horizontal ground area covered by canopy.



- Global annual minimum and maximum LAI.
- Tropical forests: > 4, 5
- Grassland:≤ 1

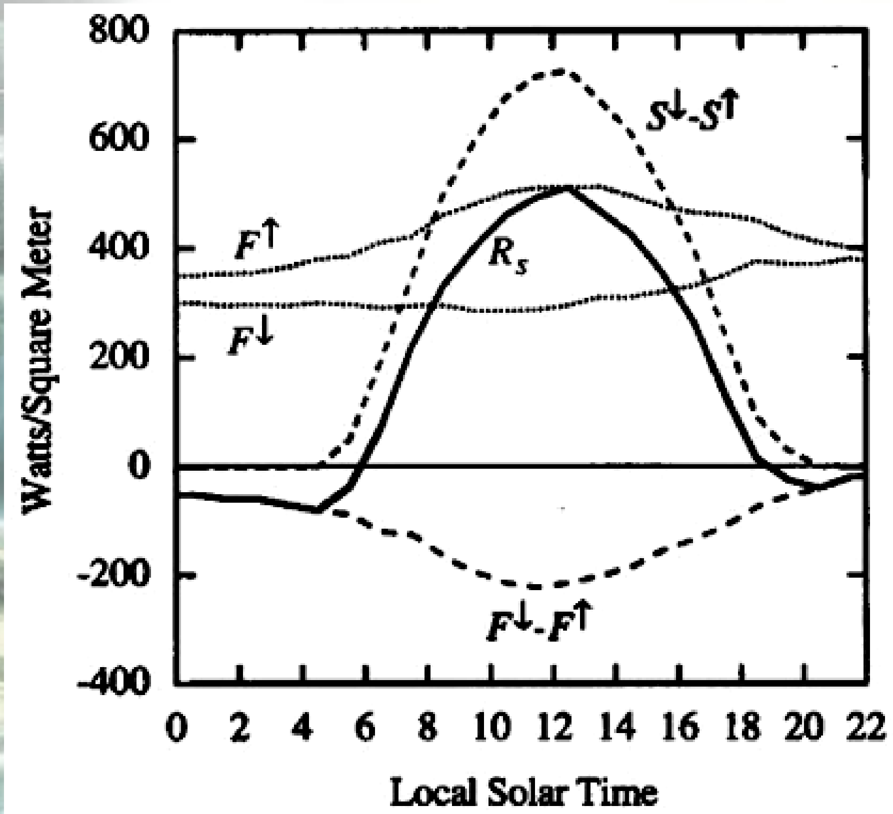


Asner et al. 2003: Global synthesis of LAI (Global Ecology & Biogeography, 12, 191-205)

Table 2 Statistical distribution of leaf area index by biome for the original data compilation and after removal of statistical outliers using Inter-Quartile Range (IQR) analysis (see Appendix 2 for key to acronyms)

	Original data					Data after IQR analysis				
Biome	Observations	Mean	Standard deviation	Min	Max	Outliers Removed	Mean	Standard deviation	Min	Max
All	931	5.2	4.1	0.01	47.0	53	4.5	2.5	0.01	18.0
Crops	88	4.2	3.3	0.2	20.3	5	3.6	2.1	0.2	8.7
Desert	6	1.3	0.9	0.6	2.8	0	1.3	0.9	0.6	2.8
Forest/BoDBL	58	2.6	1.0	0.3	6.0	5	2.6	0.7	0.6	4.0
Forest/BoENL	94	3.5	3.3	0.5	21.6	8	2.7	1.3	0.5	6.2
Forest/BoTeDNL	17	4.6	2.4	0.5	8.5	0	4.6	2.4	0.5	8.5
Forest/TeDBL	187	5.1	1.8	0.4	16.0	3	5.1	1.6	1.1	8.8
Forest/TeEBL	58	5.8	2.6	0.8	12.5	1	5.7	2.4	0.8	11.6
Forest/TeENL	215	6.7	6.0	0.01	47.0	16	5.5	3.4	0.01	15.0
Forest/TrDBL	18	3.9	2.5	0.6	8.9	0	3.9	2.5	0.6	8.9
Forest/TrEBL	61	4.9	2.0	1.5	12.3	1	4.8	1.7	1.5	8.0
Grasslands	28	2.5	3.0	0.3	15.4	3	1.7	1.2	0.3	5.0
Plantations	77	8.7	4.3	1.6	18.0	0	8.7	4.3	1.6	18.0
Shrublands	5	2.1	1.6	0.4	4.5	0	2.1	1.6	0.4	4.5
Tundra	13	2.7	2.4	0.2	7.2	2	1.9	1.5	0.2	5.3
Wetlands	6	6.3	2.3	2.5	8.4	0	6.3	2.3	2.5	8.4

Radiation Budget: Diurnal Cycle



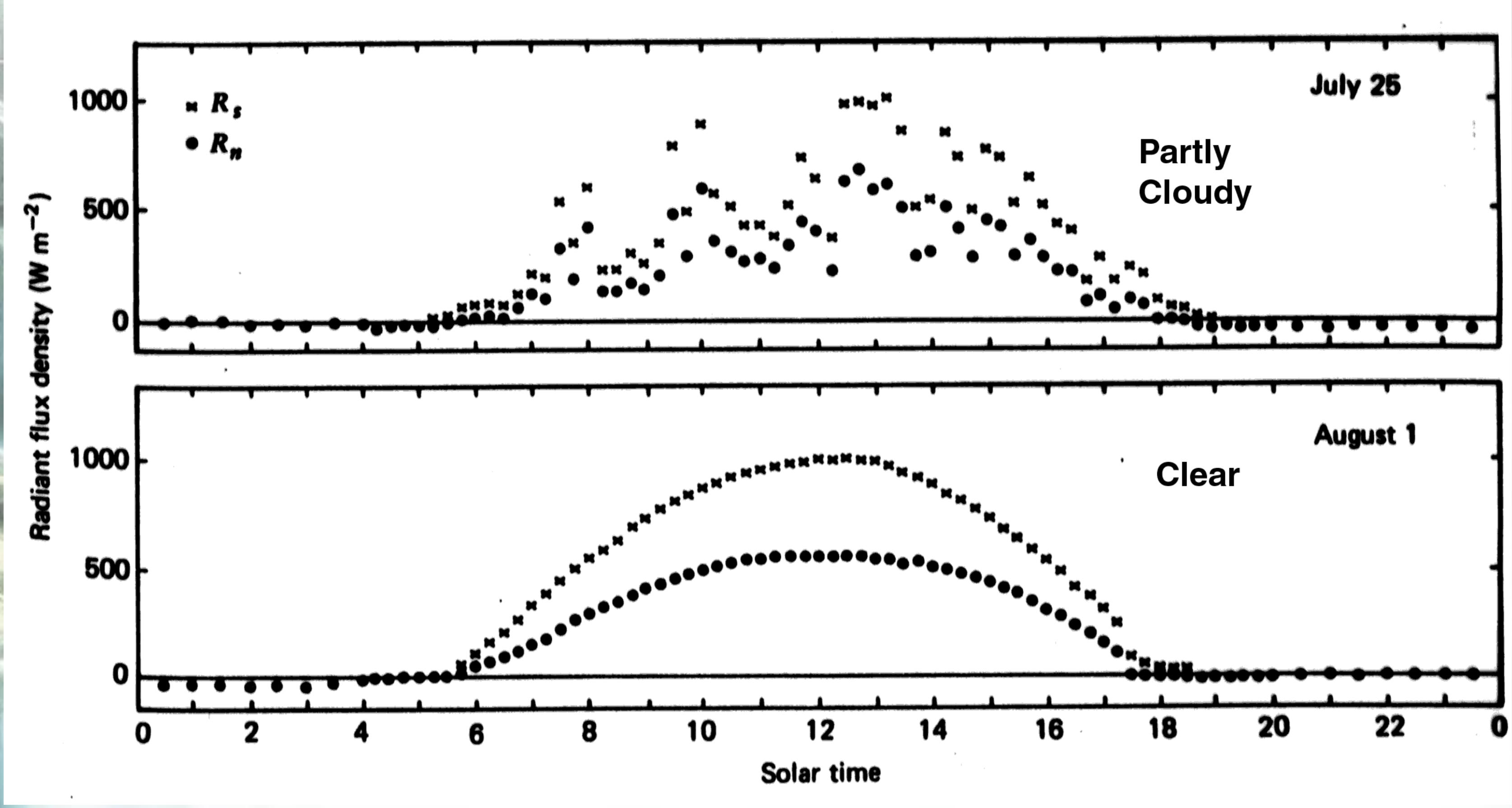
Canadian Prairie

Late July

(no clouds)

- Net solar follows cos q
- LW fluxes much less variable ($\varepsilon\sigma T^4$)
  - LW up follows surface T as it warms through day
  - LW down changes little
  - LW net opposes SW net
- R_s positive during day, negative at night

Effects of Clouds



Clouds add variance, and shift radiation load from direct to diffuse (nondirectional)

The Scope of Atmosphere – Biosphere Exchange

Atmosphere – Biosphere Exchange =

Applied Micrometeorology

The "realm" of micrometeorology is the …

Atmospheric Boundary Layer (ABL)

1st Definition: The lowest layer of the atmosphere that is influenced by the Earth's surface.

However: Given enough time or space, the entire atmosphere will adjust to surface conditions (boundary conditions).

Atmospheric Boundary Layer (ABL)

Then again: Surface conditions usually change too quickly for the entire atmosphere to get into balance with them. The most dominant changes arise from the **diurnal cycle of energy exchange** at the surface (= diurnal thermal forcing of surface conditions).

2nd Definition: The lowest layer of the atmosphere that is in *direct contact* with Earth's surface. Conditions and processes within the ABL will *react to changes* at the surface within a period of *less than an hour* and within a distance of *less than 100 km*.

Atmospheric Boundary Layer (ABL)

- The ABL is usually characterized by **turbulence** and (mechanical or thermal) convection.
- In response to the surface forcing, the ABL forms a characteristic **vertical structure** with a characteristic **diurnal cycle** (evolution of the ABL)
- The ABL is the **interface** for all exchange between the surface and the higher atmosphere:

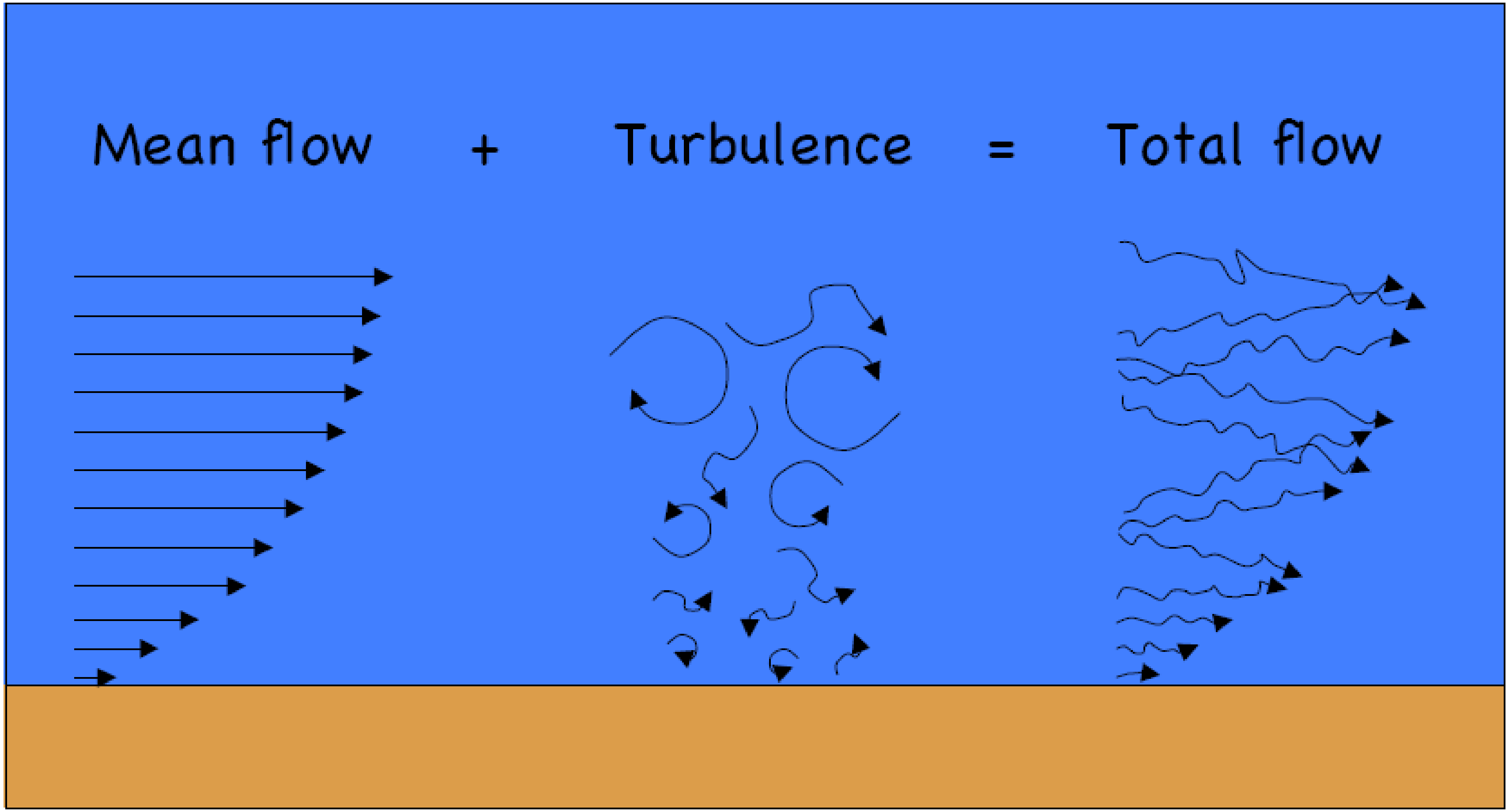
For example, any rain that falls must first…

… be evaporated and transported up,
through the boundary layer …

… by **micrometeorological processes**!

Turbulence

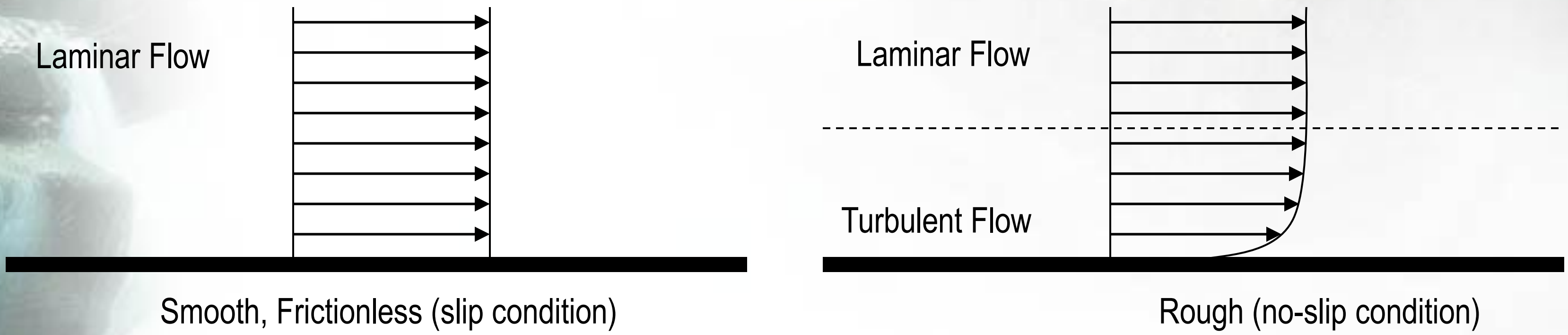
- Turbulence often defined as departure from mean flow:



- This separation between mean flow and turbulence is called a Reynolds Decomposition.

Friction

Earth's surface is <u>not</u> smooth, frictionless, or inert. Thus there exist vertical wind speed, temperature, and moisture gradients.



•Vertical gradients in wind are caused by the "no-slip" condition at the interface between the fluid (atmosphere) and solid surface:

$$\vec{V} = 0$$

•Vertical gradients in temperature are caused by the different thermal and radiative properties of the surface and fluid.

•Vertical gradients in moisture are caused by the different water holding properties of the surface and fluid.

Vertical Structure of the Atmosphere

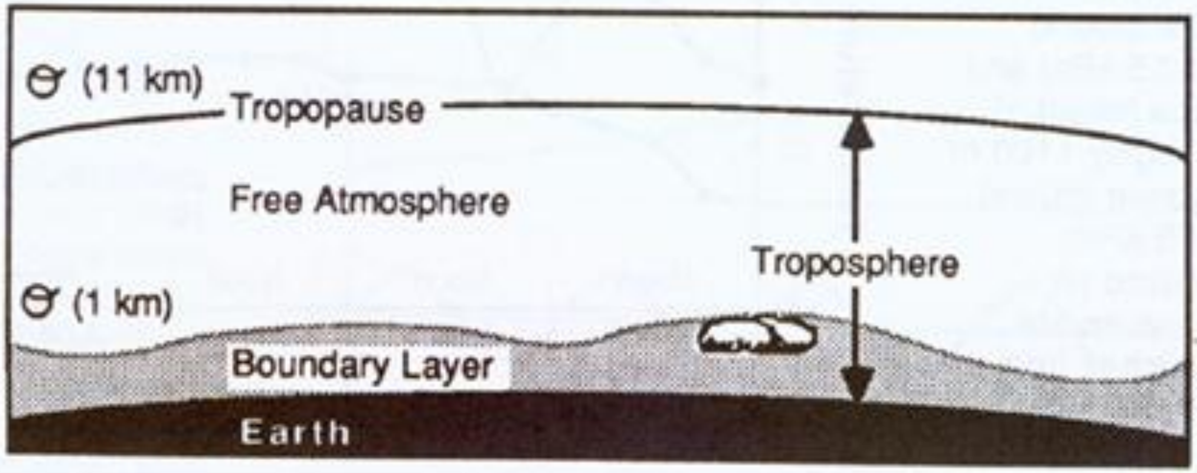


Fig. 1.1 The troposphere can be divided into two parts: a boundary layer (shaded) near the surface and the free atmosphere above it.

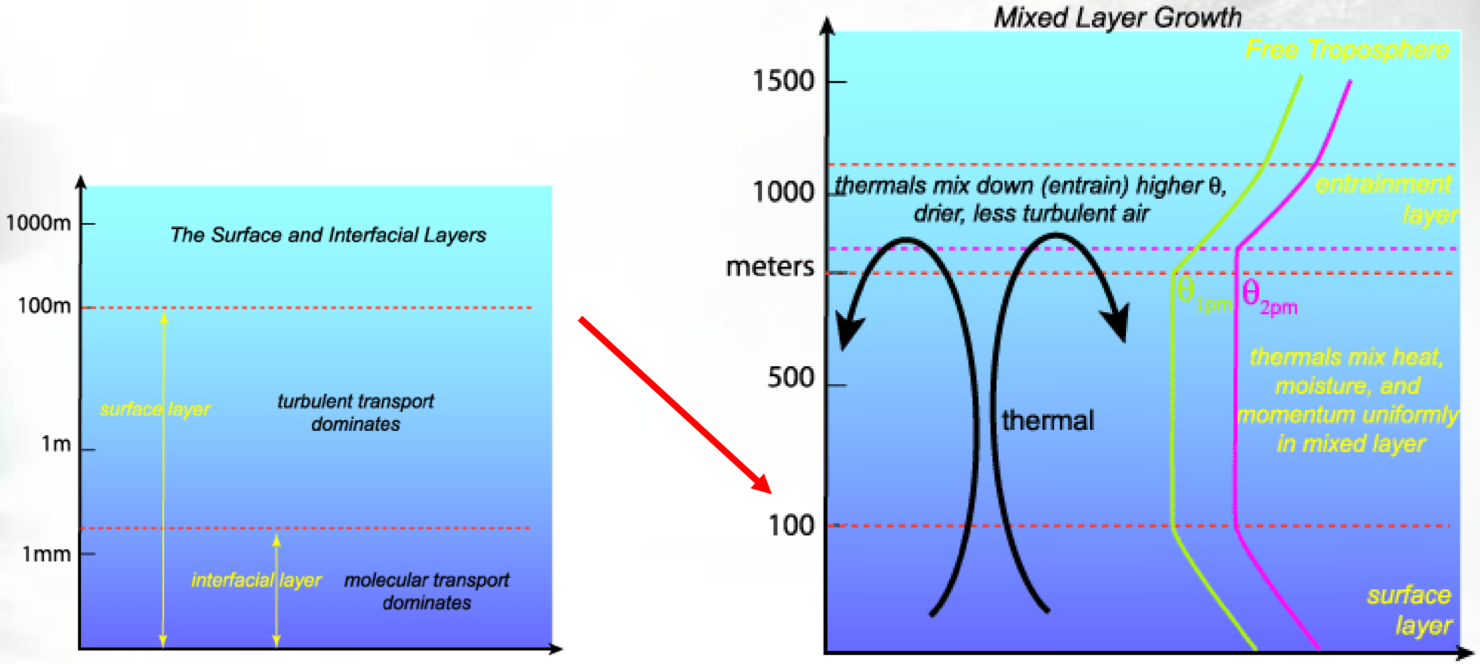
- Definition of the **boundary layer**: "that part of the troposphere that is directly influenced by the presence of the earth's surface and responds to <u>*surface forcings*</u> with a time scale of about an hour or less."
- Scale: variable, typically between 100 m - 3 km deep

Difference between boundary layer and free atmosphere

The boundary layer is:

- More turbulent
- With stronger friction
- With more rapid dispersion of pollutants
- With non-geostrophic winds while the free atmosphere is often with geostrophic winds

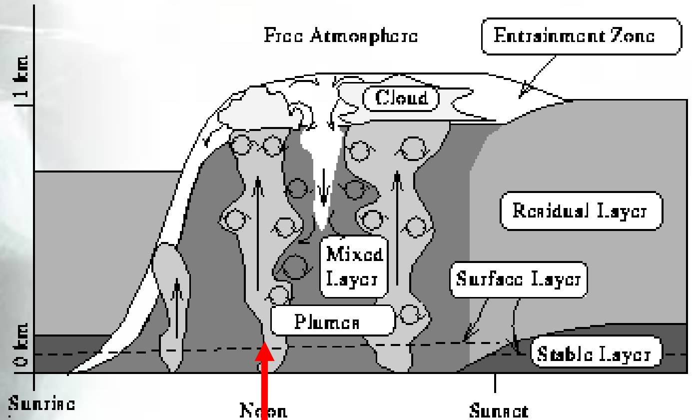
Vertical structure of the boundary layer

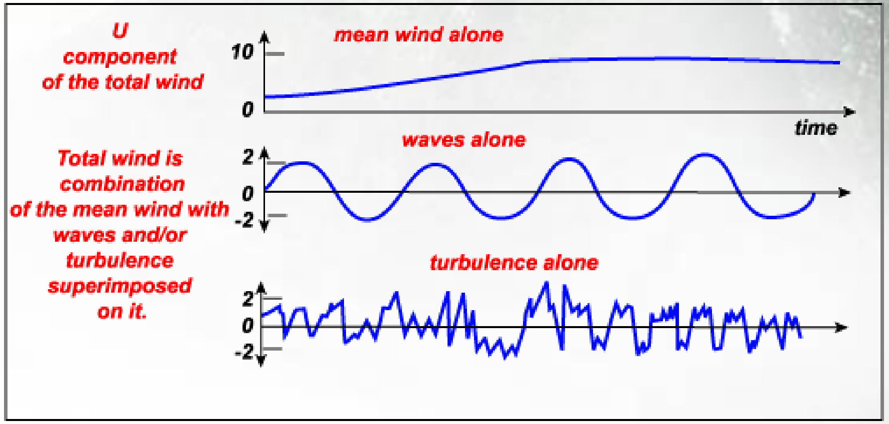


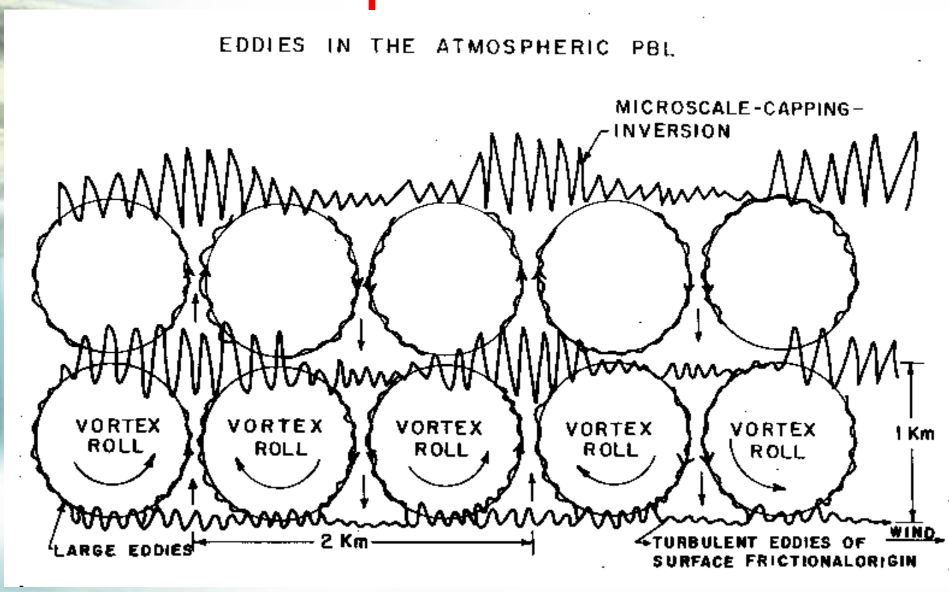
From bottom up:

- Interfacial layer (0-1 cm): molecular transport, no turbulence
- Surface layer (0-100 m): strong gradient, very vigorous turbulence
- Mixed layer (100 m - 1 km): well-mixed, vigorous turbulence
- Entrainment layer: inversion, intermittent turbulence

Turbulence inside the boundary layer







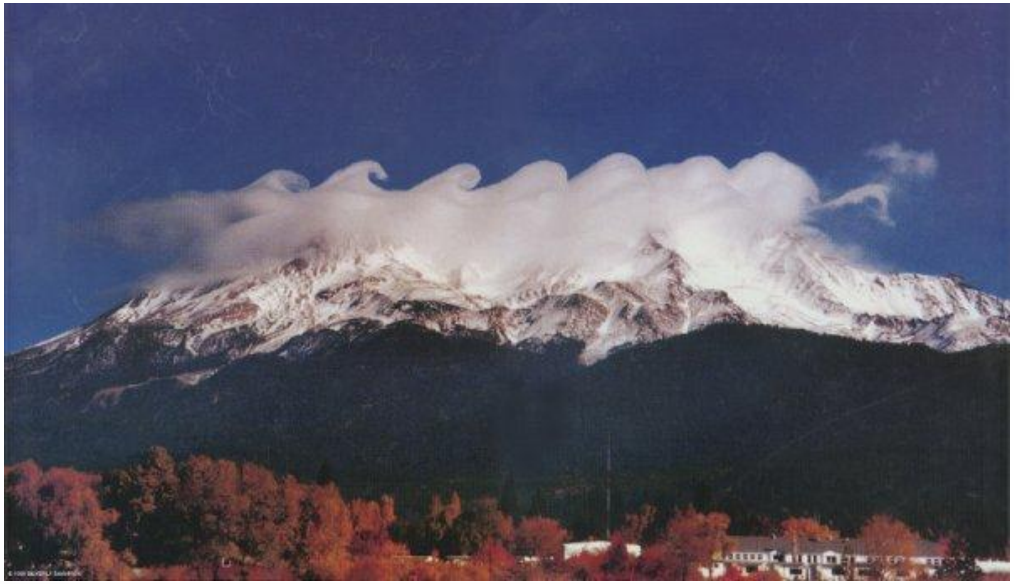
Definition of Turbulence:
The apparent chaotic nature of many flows, which is manifested in the form of irregular, almost random fluctuations in velocity, temperature and scalar concentrations around their mean values in time and space.

Example: Kelvin-Helmholtz instability

Shear instability within a fluid or between two fluids with different density

Lab experiment

Real world (K-H clouds)

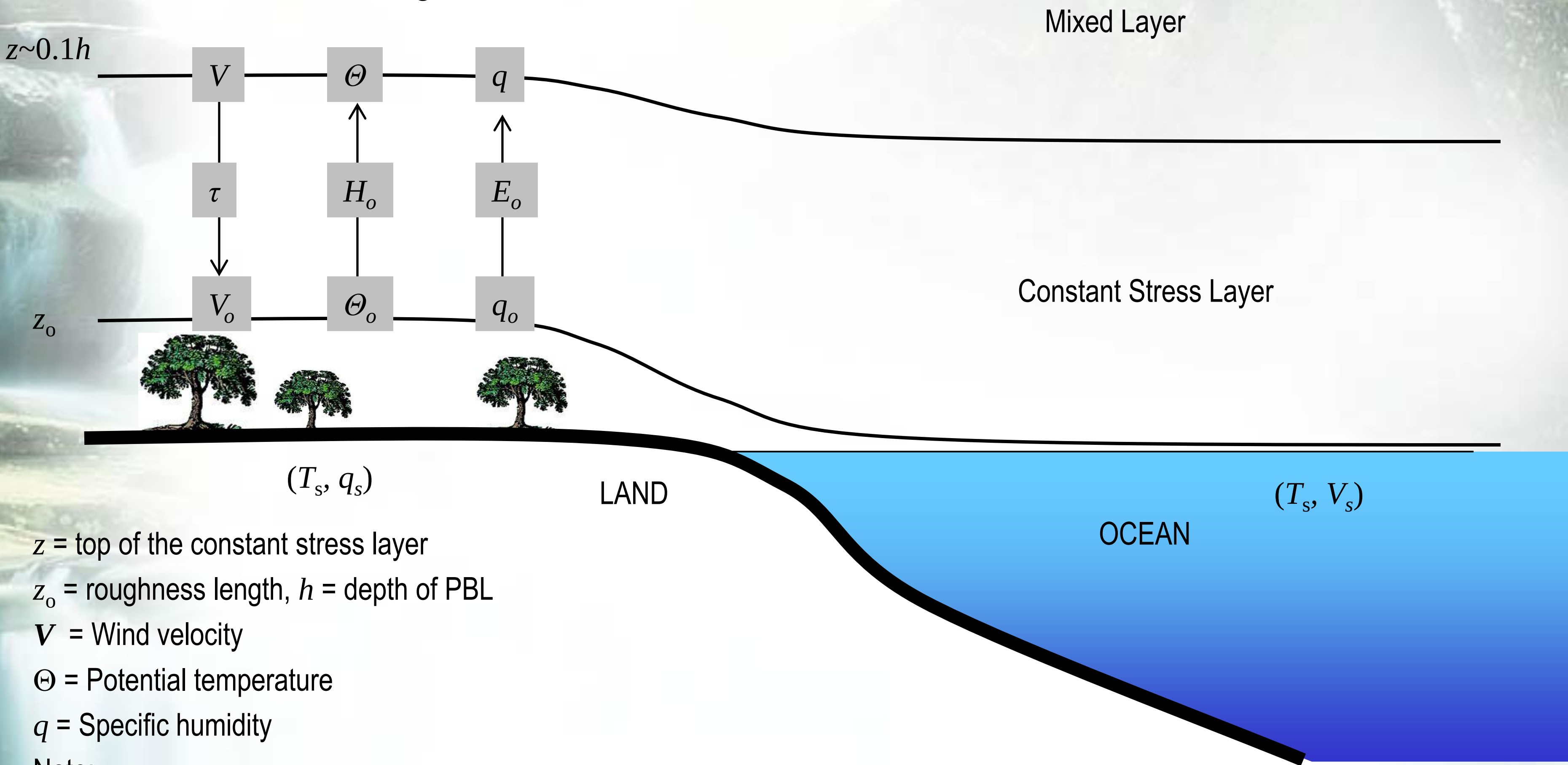


Boundary layer depth: Effects of ocean and land

- Over the oceans: varies more slowly in space and time because sea surface temperature varies slowly in space and time
- Over the land: varies more rapidly in space and time because surface conditions vary more rapidly in space (topography, land cover) and time (diurnal variation, seasonal variation)

Constant Stress Layer

This layer near the surface is of primary concern for land-atmosphere interactions; it lies between the soil/vegetation models and the AGCM.



z = top of the constant stress layer

z_o = roughness length, h = depth of PBL

$\boldsymbol{V}$ = Wind velocity

Θ = Potential temperature

q = Specific humidity

Note:

$\boldsymbol{V}_o = 0$ over land (assumed), $\boldsymbol{V}_o = \boldsymbol{V}_s$ over ocean

Boundary layer depth: Effect of highs and lows

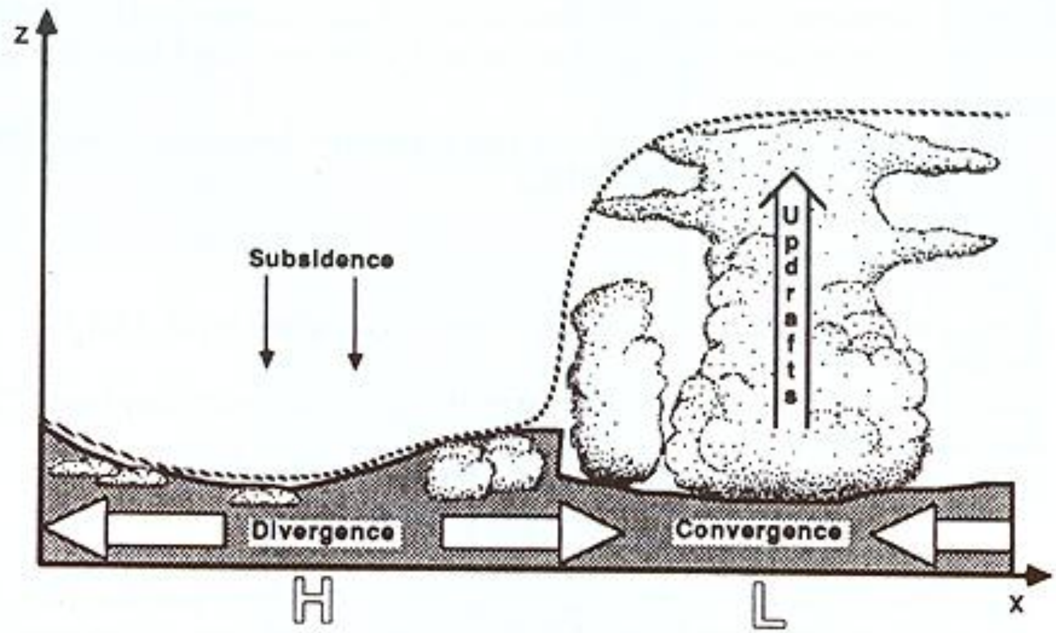


Fig. 1.6 Schematic of synoptic - scale variation of boundary layer depth between centers of surface high (H) and low (L) pressure. The dotted line shows the maximum height reached by surface modified air during a one-hour period. The solid line encloses the shaded region, which is most studied by boudary-layer meteorlogists.

from Stull 1988

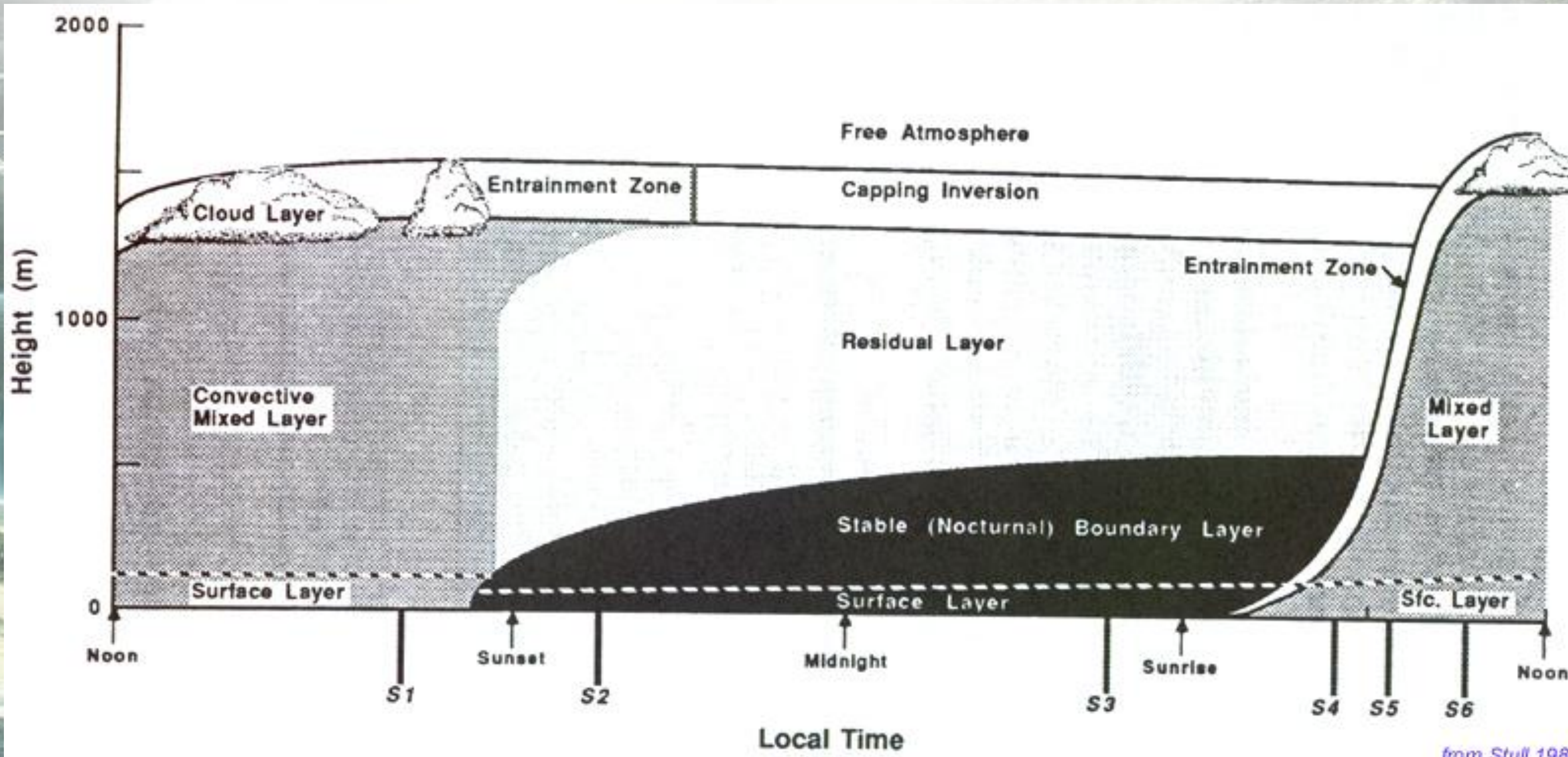
Near a region of high pressure:

- Over both land and oceans, the boundary layer tends to be shallower near the center of high pressure regions. This is due to the associated subsidence and divergence.
- Boundary layer depth increases on the periphery of the high where the subsidence is weaker.

Near a region of low pressure:

- The rising motion associated with the low transports boundary layer air up into the free troposphere.
- Hence, it is often difficult to find the top of the boundary layer in this region. Cloud base is often used at the top of the boundary layer.

Boundary Layer depth: Effects of diurnal forcing over land



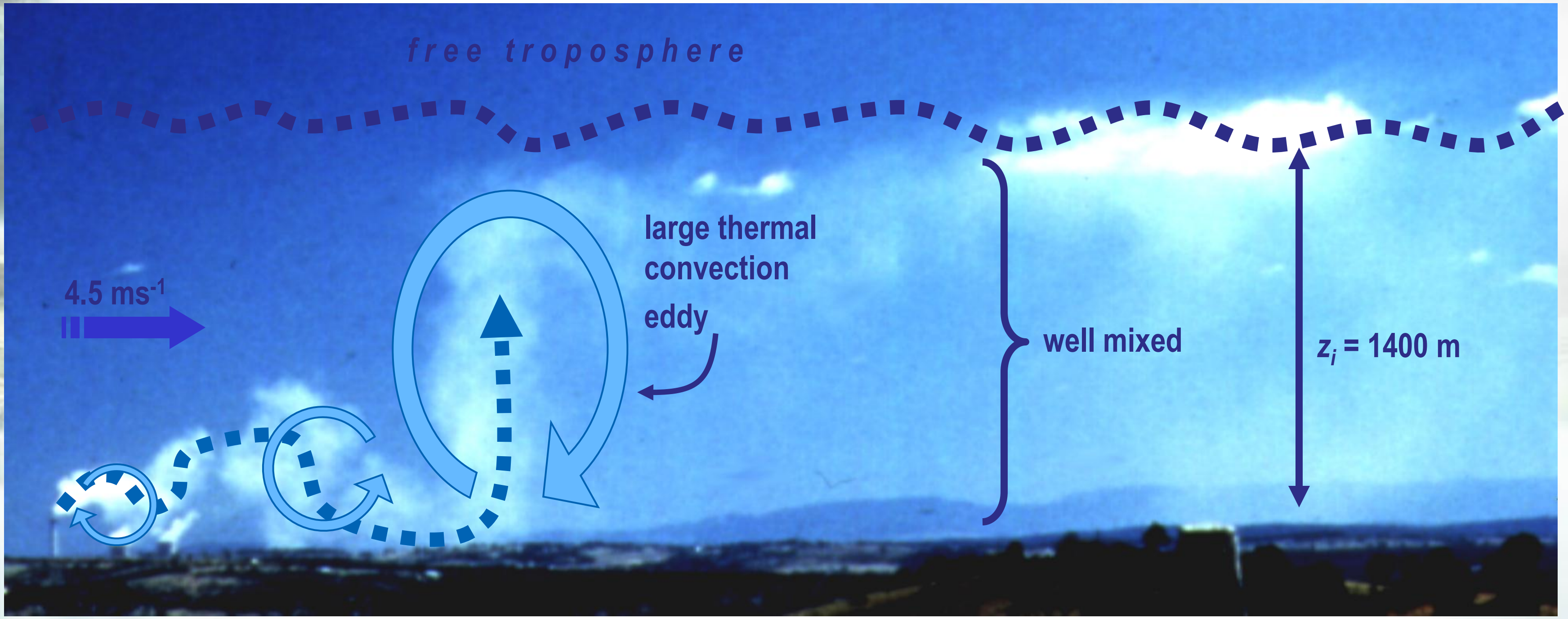
from Stull 1988

- Daytime convective mixed layer + clouds (sometimes)
- Nocturnal stable boundary layer + residual layer

Atmospheric Boundary Layer (ABL)

Daytime Convective Boundary Layer

- Looping plume, in the presence of large convective thermal eddies
- Lifting limited by capping inversion; free troposphere above
- Well mixed conditions downwind, in mixed layer of ~1400 m depth



Tarong, Queensland (AUS), stack height: 210 m, z_i = 1400 m, w^* = 2.5 ms^{-1}. Photo: Geoff Lane, CSIRO (AUS)

Atmospheric Boundary Layer (ABL)

Entrainment / Encroachment: ABL Growth Mechanism

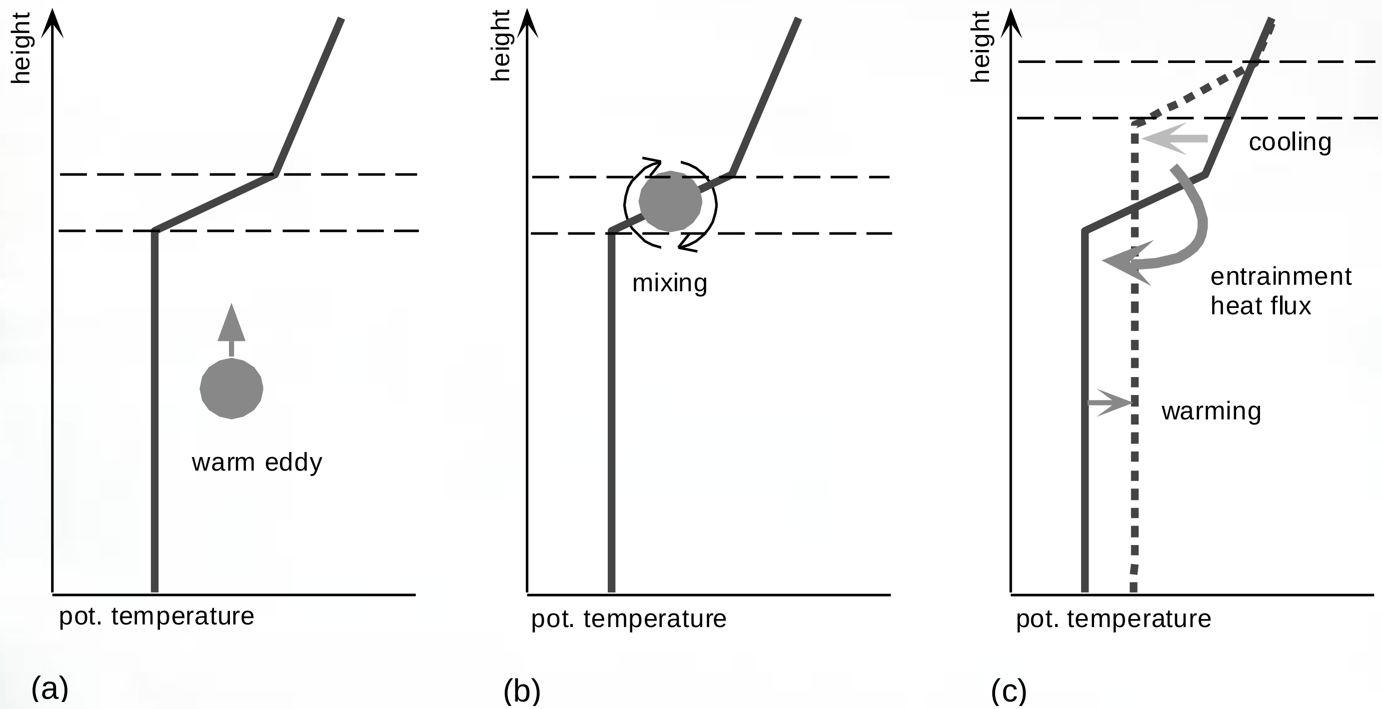
a: a *warm eddy* (incrtement of H_0) rises up from the surface, with positive buoyancy (higher potential temperature than its surroundings)

b: no more buoyance at the inversion height; "bubbling up" and the kinetic energy of the eddy causes *mixing* (also vertical!) …

c: … of *warmer air above* into the ABL = entrainment.

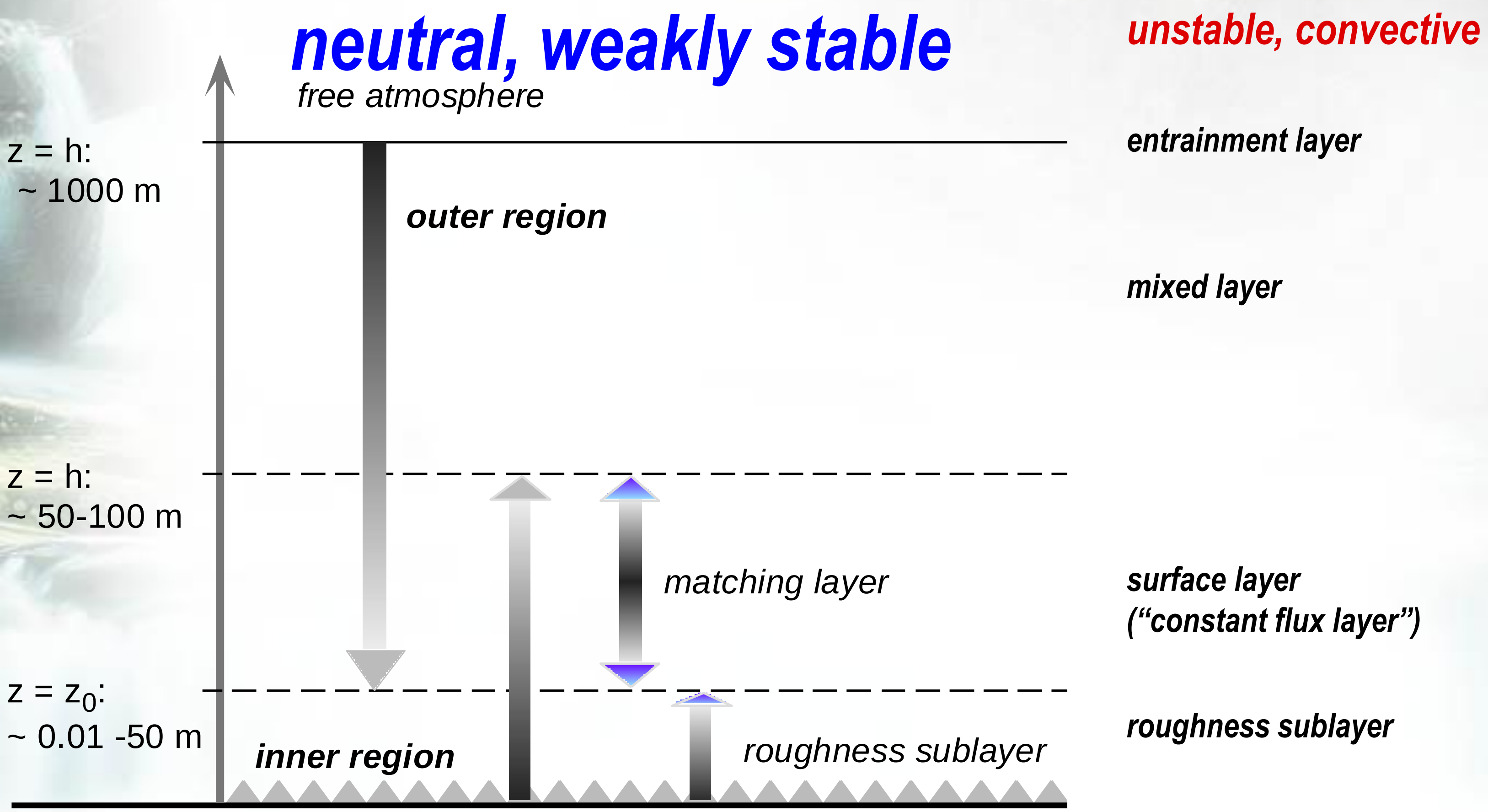
entrainment: mixing warm "free" air into ABL ⇒ ABL grows

encroachment: mixed layer warms against capping inversion ⇒ ABL grows



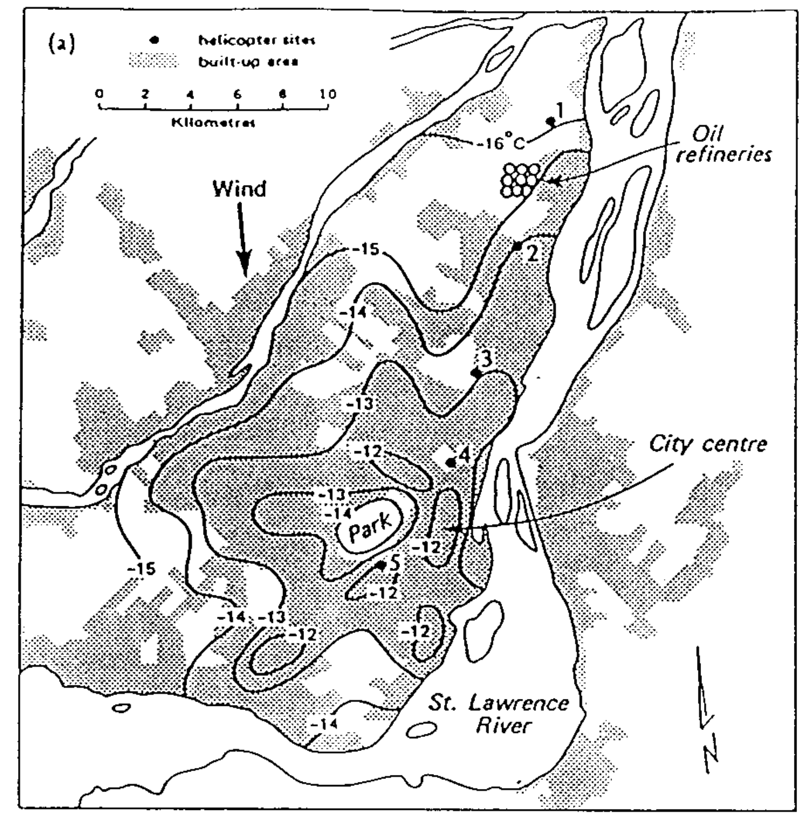
Atmospheric Boundary Layer (ABL)

Subdivision into Sublayers

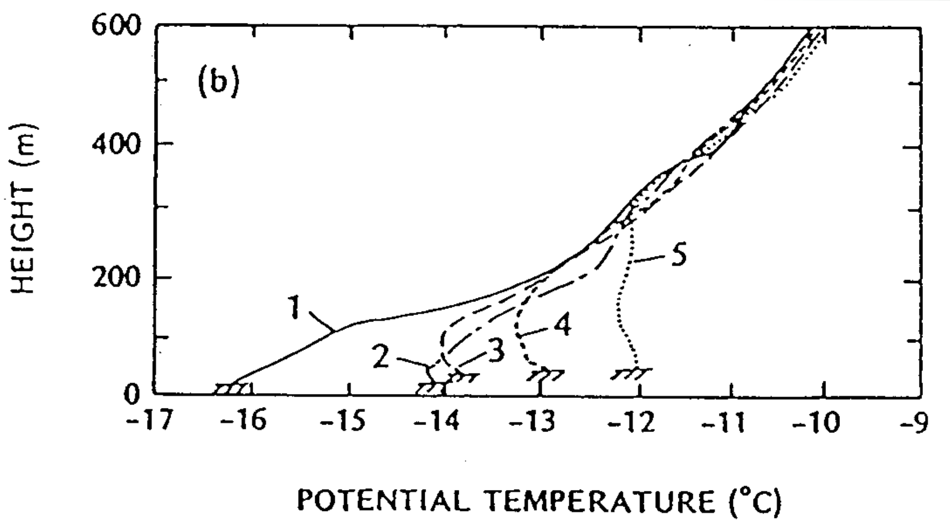


Scales and Similarity

Spatial Variation of ABL Development

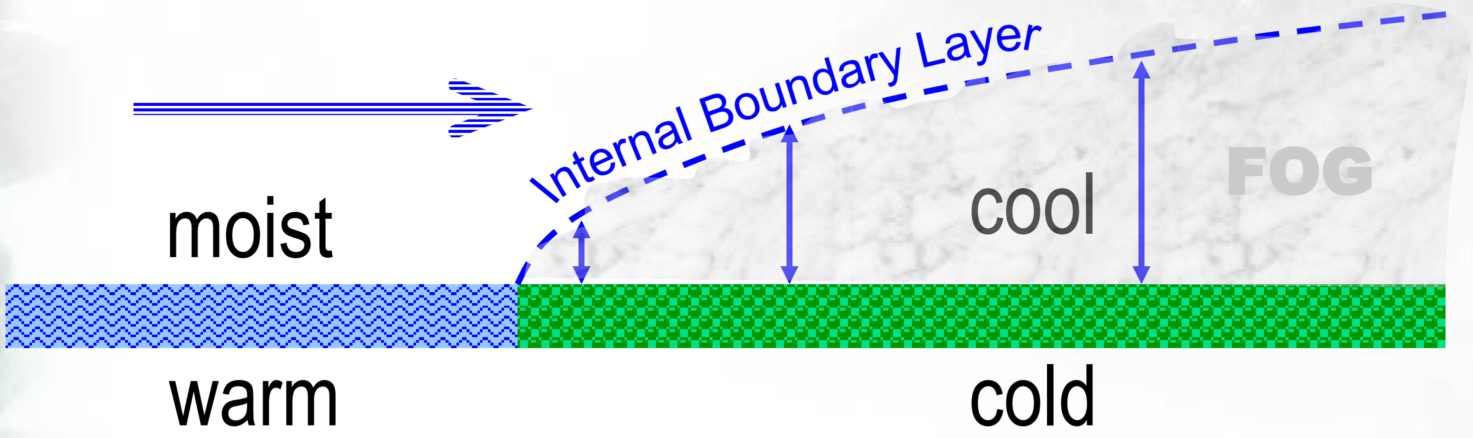


- urban heat island, winter morning
- higher H_0 in city center
- wind from ~ north
- pot. temperature profiles from rural surroundings to urban center
- growing mixed layer towards center



(Figures from Oke, 1987)

Scales and Similarity



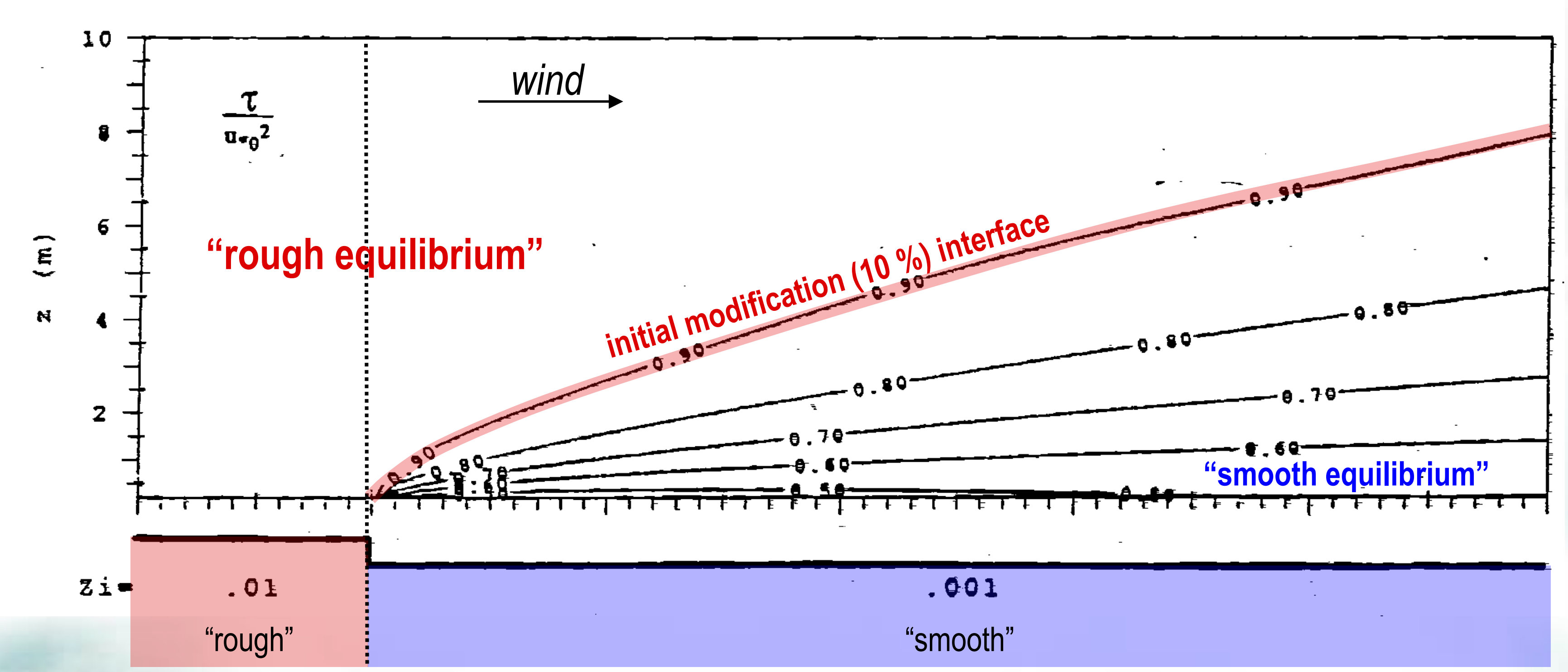
"The depth over which the ABL adjusts to new surface conditions is related to time it is allowed to do so"

"The strength of a thermal adjustment depends on the magnitude of the thermal forcing"

Scales and Similarity

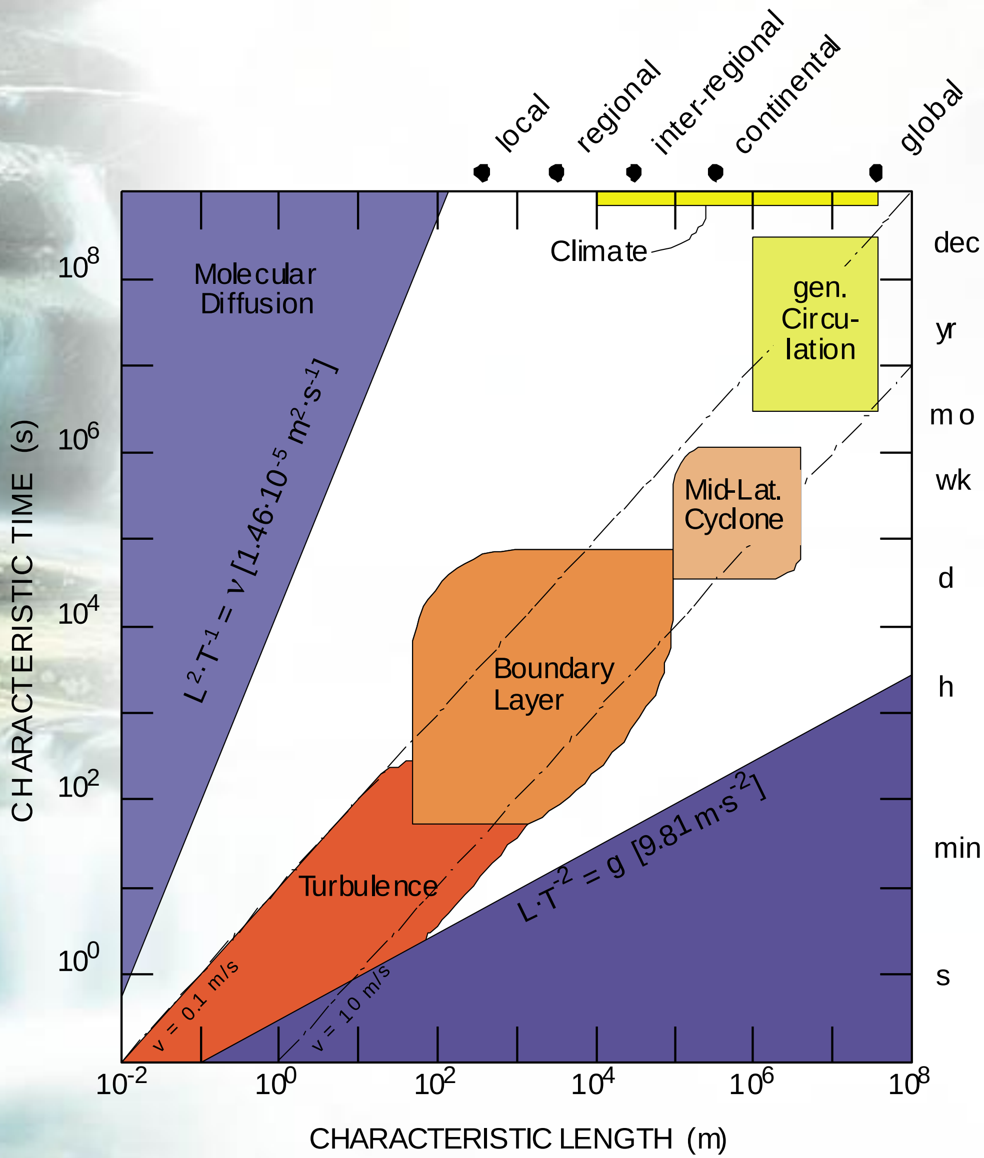
Internal Boundary Layer Development

- growth of layer with new surface effects downstream from leading edge
- "rough" equilibrium upstream and above leading edge
- "smooth" equilibrium far downstream from leading edge at low height
- Internal boundary layer (e.g. initial modification interface) grows downstream depending on the magnitude of the change



(Figures from Wild, 1991)

Scales and **Similarity**: Space-Time Frame



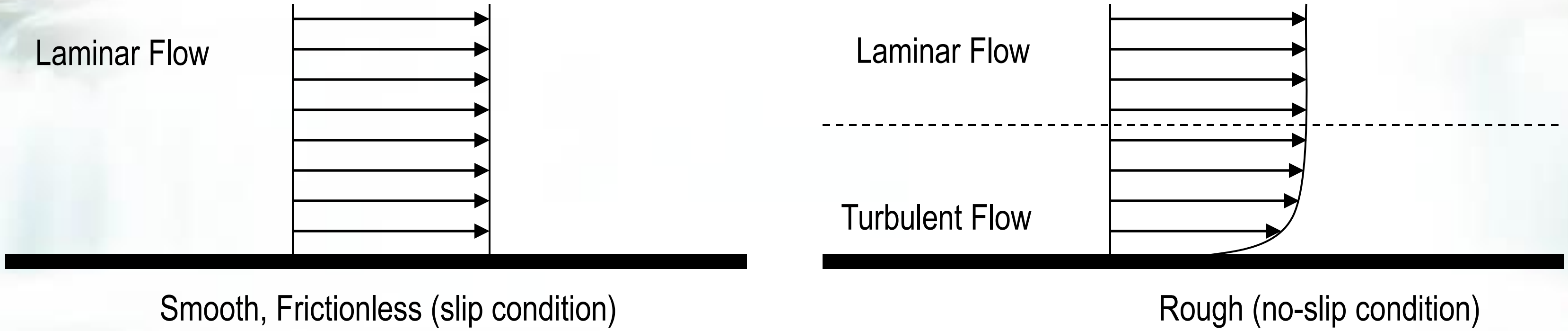
"Distinct regimes" implies that processe in different scaling regimes **cannot interact directly**: we can analyze them independently of each other!!! (= **makes solutions possible!!!**)

Scales and Similarity

- The **similarity hypothesis** postulates that:
  processes and phenomena within a given ***scaling group*** *behave in a similar way, irrespective of time and location.*

Similarity theory:

Similarity theories for given scaling regimes are the most significant frameworks of analysis in the boundary layer and micrometeorology (actually: in all fluid dynamics).



Flux Measurement

Evapotranspiration, and turbulent mixing

Turbulent Fluxes

$$u = \bar{u} + u'$$

Total velocity — Mean velocity — Turbulent velocity (mean of $u' = 0$).

$$T = \bar{T} + T'$$

Temperature — Mean temperature — Turbulent temperature fluctuation (mean of $T' = 0$).

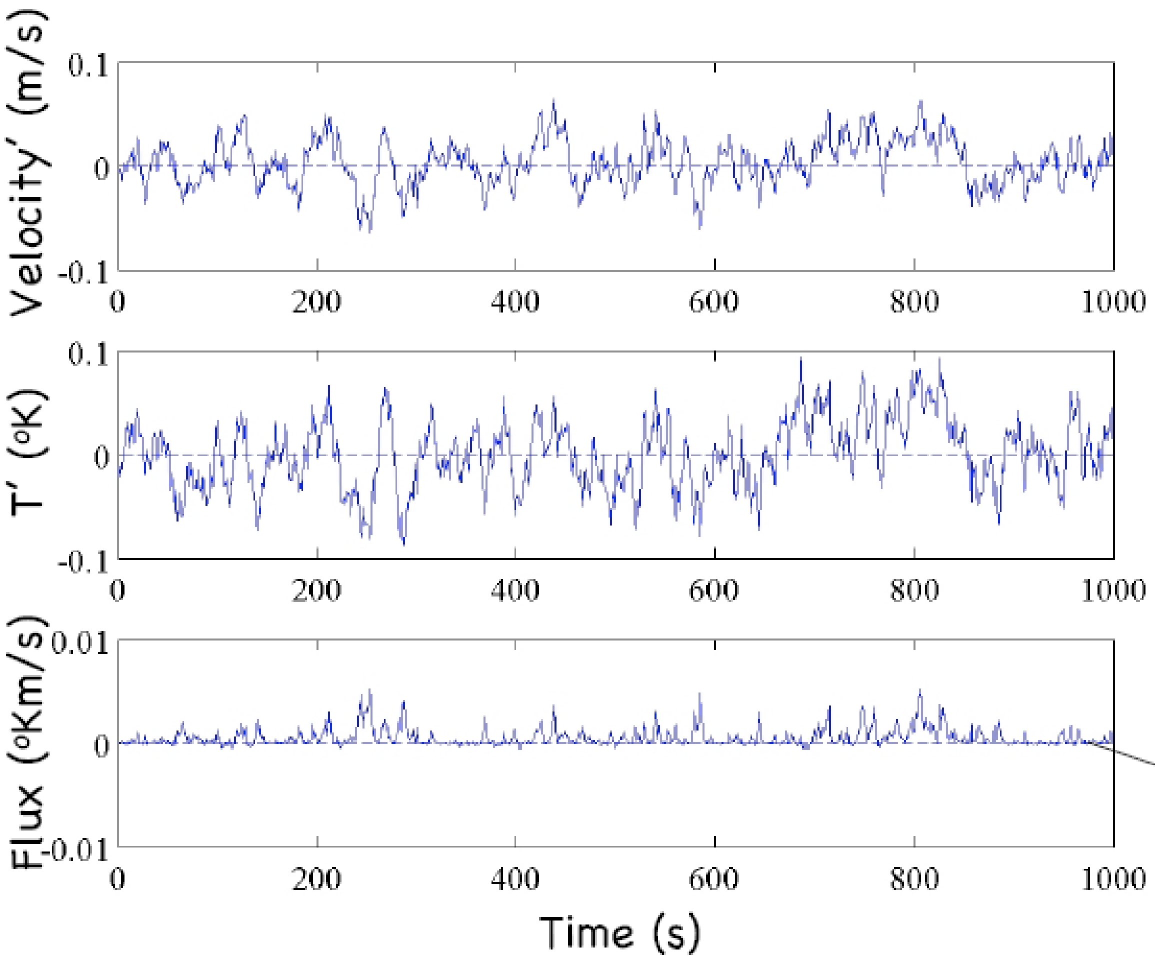
- Heat flux equals $u \times T$.
- Mean heat flux equals mean of $u \times T = \overline{u \times T}$.
- Without any assumptions, a little algebra leads to

$$\overline{uT} = \bar{u} \times \bar{T} + \overline{u' \times T'}$$

Mean flux — Flux from mean flow (average u and T, then multiply). — Flux from turbulence (multiply u′ and T′, then average).

- In practice u' and T' are unpredictable, but $\overline{u' \times T'}$ depends on systematic relationship between u and T, so it is predictable.

Turbulent mixing of heat



Measuring turbulent velocity and temperature requires expensive, rapidly-sampling instruments.

Vertical heat flux (temperature' × velocity') is usually directed upward.

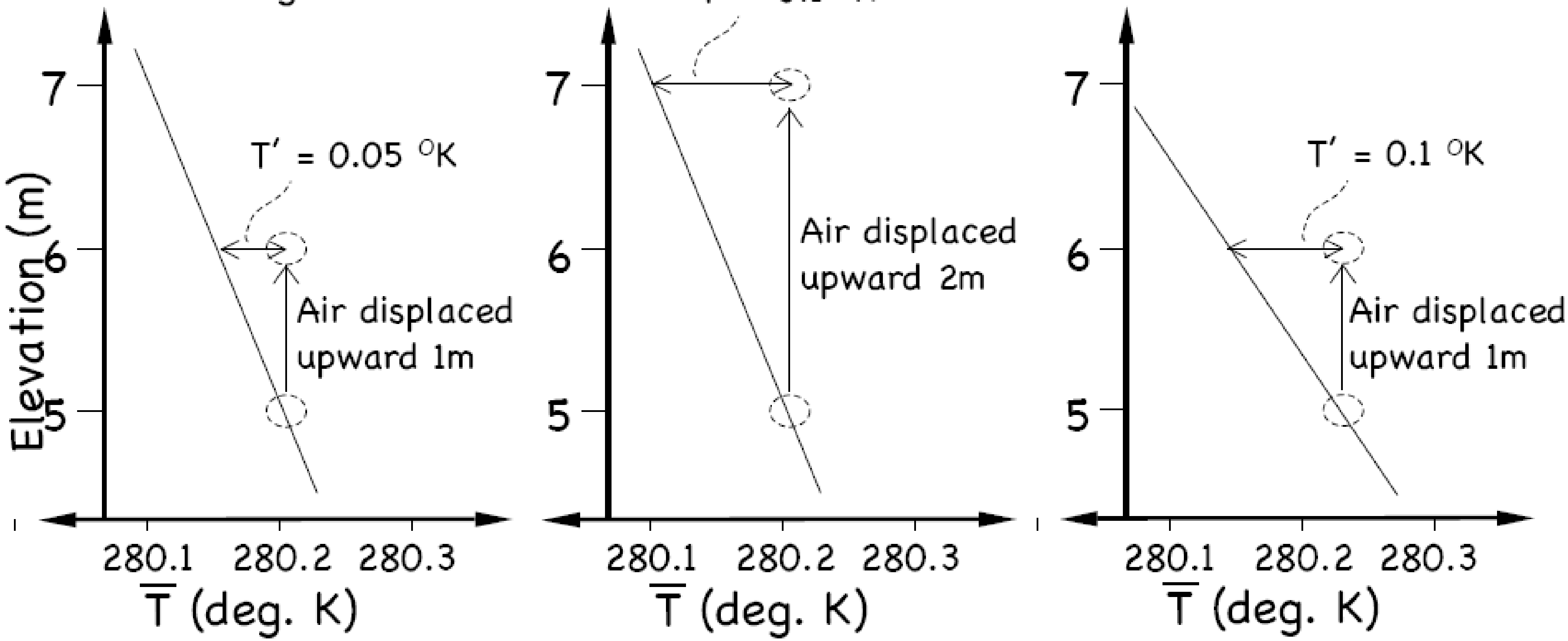
- Average turbulent heat flux = average of (velocity × temperature) is directed upward, even though average upward velocity is zero.

Turbulent temperature fluctuations

Case 1: Turbulence lifts air parcel 1 m before vertical motion ceases and parcel mixes with surrounding air. $T'=0.05^oK$

Case 2: Doubling displacement doubles T' from case 1.

Case 3: Doubling mean temperature gradient doubles T' from case 1.



- In general,

$T' \approx -$ (displacement distance) $\times$ (mean temperature gradient)

Expression for sensible heat flux

- Substituting
  $T' \approx -$ (vertical displacement) $\times$ (mean temperature gradient)
  into
  $$\text{Heat flux} = \overline{u' \times T'}$$
  gives

$$\text{Heat flux} \approx -\ \overline{u' \times (\text{vertical displacement})} \times (\text{mean temperature gradient})$$

- In practice, vertical displacements are comparable to elevation above ground, u' is comparable to the standard deviation of u, and
  $$\boxed{\text{Heat flux} \approx -\ D_{T,turb} \times d\overline{T}/dz,}$$
  Where
  $d\overline{T}/dz$ = mean temperature gradient,
  $D_{T,turb}$ = 0.4 $\times$ (standard deviation of u) $\times$ elevation.

Called eddy diffusivity Called von-Karman's constant, determined from experiments

Sensible heat flux in energy units

$$\text{Turbulent heat flux} = - D_{T,turb} \times d\overline{T}/dz,$$

- To convert flux measured in $^oK\ m\ s^{-1}$ into flux measured in Watts m^{-2}, need to multiply by air density (about 1.2 kg m^{-3}) and heat capacity of air (about 1000 J $kg^{-1}\ ^oK^{-1}$), giving

$$\text{Turbulent heat flux} = - D_{T,turb} \times d\overline{T}/dz \times (1200\ J\ ^oK^{-1}\ m^{-3}),$$

in Watts per m^2

- Since molecular heat transport is usually negligible, turbulent heat flux = sensible heat flux.

Latent Heat Flux

- Latent heat flux dominated by turbulent mixing (except very near surface), so

$$\text{Latent heat flux} = \overline{\mathbf{u}' \times v'}$$

where v' = turbulent fluctuation in water vapor concentration (kg m^{-3}).

- The same logic previously used to calculate the sensible heat flux leads to

$$\text{Latent heat flux} \approx - D_{v,turb} \times d\bar{v}/dz,$$

where
$d\bar{v}/dz$ = mean water vapor concentration gradient,

- To convert above flux (in kg m^{-2} s^{-1}) into flux measured in Watts m^{-2}, need to multiply by latent heat of vaporization, giving

$$\text{Latent heat flux} \approx - D_{v,turb} \times d\bar{v}/dz \times (2.3 \times 10^6 \text{ J kg}^{-1})$$

Measuring Flux with Eddy Correlation

- Can measure mean vapor flux = $\overline{u' \times v'}$ by simply measuring velocity u' and vapor concentration v', multiplying and averaging. This is called eddy correlation.
- Can provide very good measurements, but requires instruments that respond fast enough to resolve turbulence (expensive).
- To obtain measurements more cheaply, need method that doesn't require measurement of turbulent fluctuations. Eddy diffusivity seems to provide such a method…

Bowen Ratio

- Bowen ratio = B = (Sensible heat flux)/(Latent heat flux)

  Sensible heat flux $\approx - D_{T,turb} \times d\overline{T}/dz \times (1200\ J\ {}^oK^{-1}\ m^{-3})$

  Latent heat flux $\approx - D_{v,turb} \times d\overline{v}/dz \times (2.3 \times 10^6\ J\ kg^{-1})$

- If turbulence mixes heat and water vapor in same way, then $D_{T,turb} = D_{v,turb}$, and

  $B = [d\overline{T}/dz \times (1200\ J\ {}^oK^{-1}\ m^{-3})]/[d\overline{v}/dz \times (2.3 \times 10^6\ J\ kg^{-1})]$
  ...so B can be determined by measuring mean temperature and vapor gradients - diffusivities need not be known.

- In practice $D_{T,turb}$ is not very different from $D_{v,turb}$, and it is common to assume $D_{T,turb} = D_{v,turb}$

Usefulness of Bowen Ratio

- Surface energy balance is

  Incoming shortwave – outgoing shortwave +
  incoming longwave – outgoing longwave – outgoing soil heat =
  outgoing sensible heat + outgoing latent heat

  …but outgoing sensible heat = B × outgoing latent heat, so…

  Incoming shortwave – outgoing shortwave +
  incoming longwave – outgoing longwave – outgoing soil heat =
  (B + 1) × outgoing latent heat

  …can be evaluated without knowing turbulent diffusivities.

Measuring evaporation using Bowen Ratio

measures upward shortwave and longwave radiation

measures downward shortwave and longwave radiation

measured using net radiometer

temperature measured on two sides of buried plate

measured using soil heat flux plate, which records temperature gradient in soil

Incoming shortwave – outgoing shortwave + incoming longwave – outgoing longwave – outgoing soil heat =
(B + 1) X outgoing latent heat

calculated from vertical gradients in measured mean temperature and vapor content

temperature and vapor measured at two elevations to determine gradient

Measuring evaporation using Lysimeters

- A lysimeter is a block of soil and vegetation that has been isolated to allow accurate calculation of water balance.

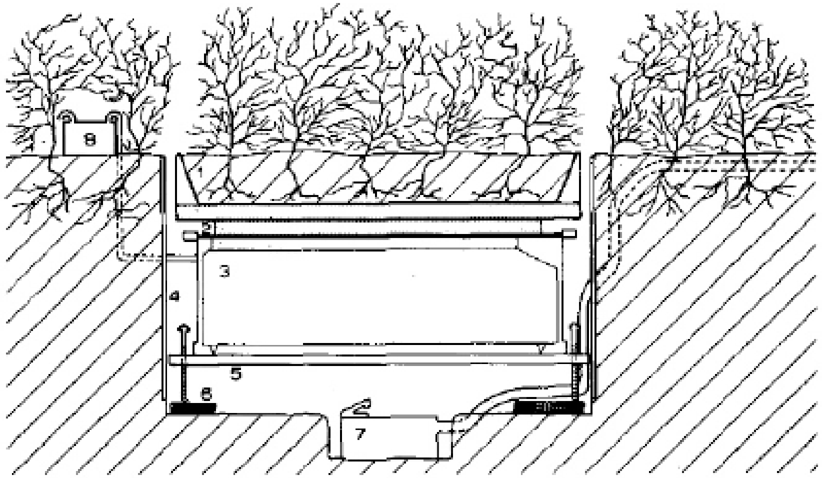


FIG. 4. A cross-sectional diagram of the lysimeter showing: 1, the heather sample in the tray (incorporating the block of insulating polystyrene); 2, polystyrene spacer block on neoprene rubber top of box; 3, the electronic balance inside its protective box; 4, aluminium hole liner; 5, leveling tray; 6, concrete blocks; 7, water pump; and 8, breather box containing silica gel.

[Calder *et al.* (1984), J. Climate and Appl. Met.]

Determined from changes in weight of entire lysimeter

- Evaporation can be calculated from water balance

Evaporation = precipitation - subsurface water loss - storage change

Measured by collecting and weighing rainfall

Measured by collecting and weighing water seeping from base of lysimeter

- Assembling the Eddy Covariance Towers in Albuquerque, NM

CLPX
Rabbit Ears Eddy Covariance

Objective: Characterize the water and carbon fluxes from the snowpack

Method: Eddy covariance at 3 elevations – Buffalo Pass, Spring Creek, and Potter Creek.

Observations: 10 min averages of 20hz samples

Spring Creek and Buffalo Pass:

3D Sonic Anemometer

CO2, H2O IR Gas Analyzer

2 level Wind Speed/Direction

2 level humidity/temperature

Buffalo Pass only

K&Z 4-way net radiometer

K&Z “nrlite” net radiation

Q7 net radiation

Geonor Precipitation

3 level leaf wetness

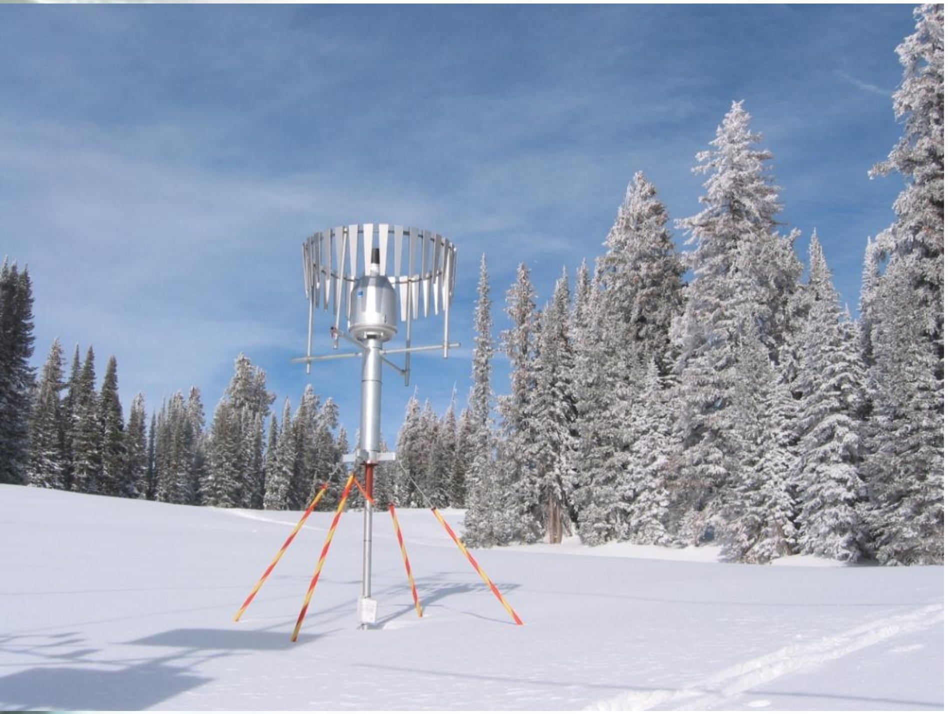
soil heat flux

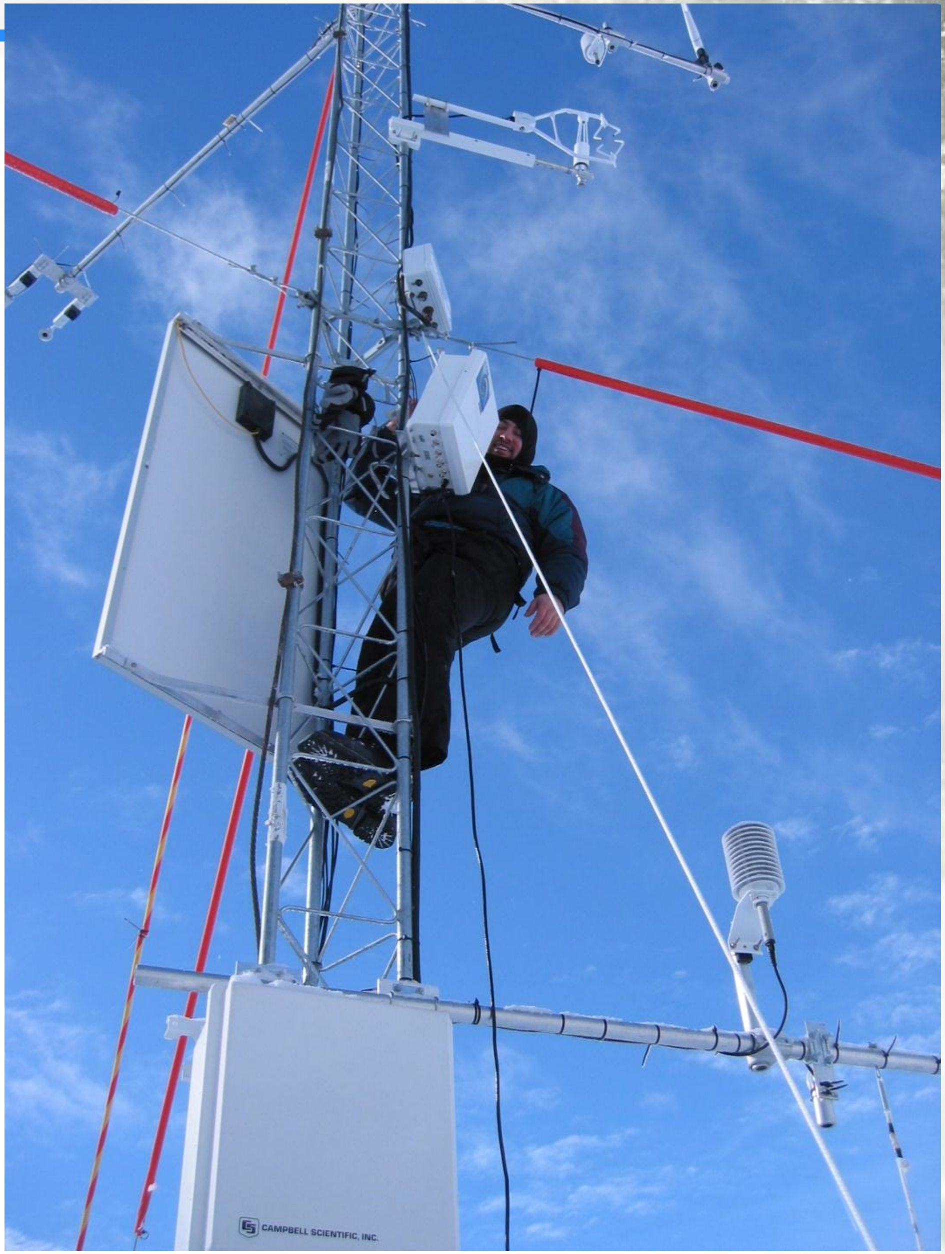
surface skin temperature (Apogee)

Snow depth (Campbell and Judd)

6 levels soil temperature, suction, moisture (to 1m)

Figure : Tower based eddy covariance system consisting of a sonic anemometer, CO_2/H_2O gas analyzer, and 4-way net radiometer. For reference, the sonic anemometer path is 20cm.

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Role of Models

Any **abstraction of reality** is a model.

Abstraction:	to capture the *essence* of something
What is essential?	is *subjective*: different models for different applications.

Will encounter **various kinds of models**:

- to explain a process in isolation (simplification)
- to examine our understanding of a process
- to organize and examine a set of data
- to obtain quantitative information without measurement

TYPES OF MODELS (an *ad hoc*, incomplete list)

Qualitative Models

- conceptual: “schematics”, informative
- visualizations: explanatory
- heuristic: exploratory (quantitative, but out of context)

Quantitative Models

Stochastic Models: results have statistical properties (e.g., uncertainty)

- empirical: *generalization* of measured data (similarity framework!), *parametrizations*: statistics deal with *unknown physics*

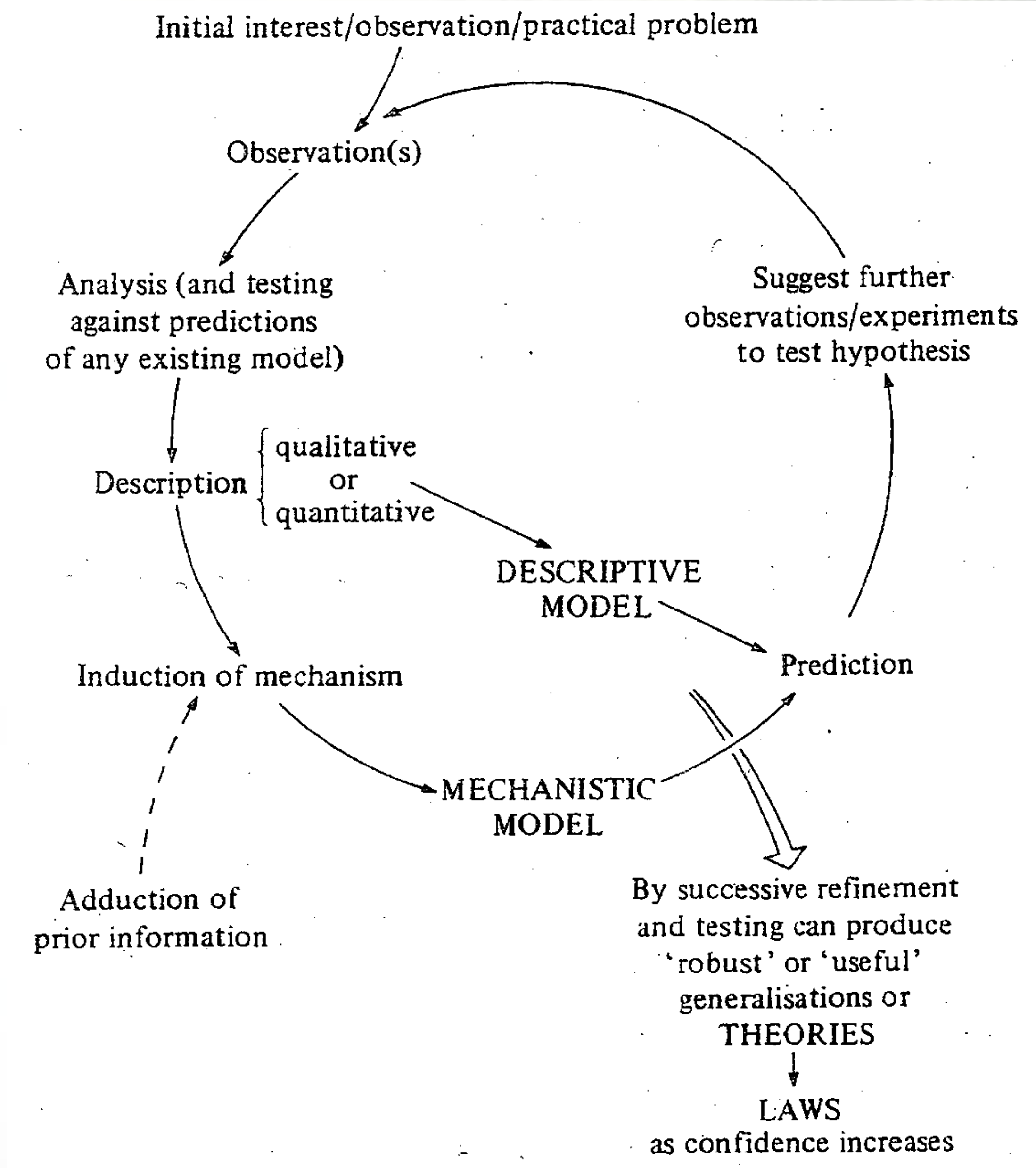
- theoretical: statistics express an inherently *random process* (e.g., statistical mechanics, turbulence)

Deterministic Models: results are unambiguous

- mathematical: *analytical* expressions of “laws” of physics (= well established formal theories) (*method of solution* may be numerical)
- mechanistic: catena of causal relationships (logical, mathematical, categorical)

Most practical models are hybrids of classes.

The Relationship of Models and Observations in the Scientific Method



(Figures from Jones, 1992)