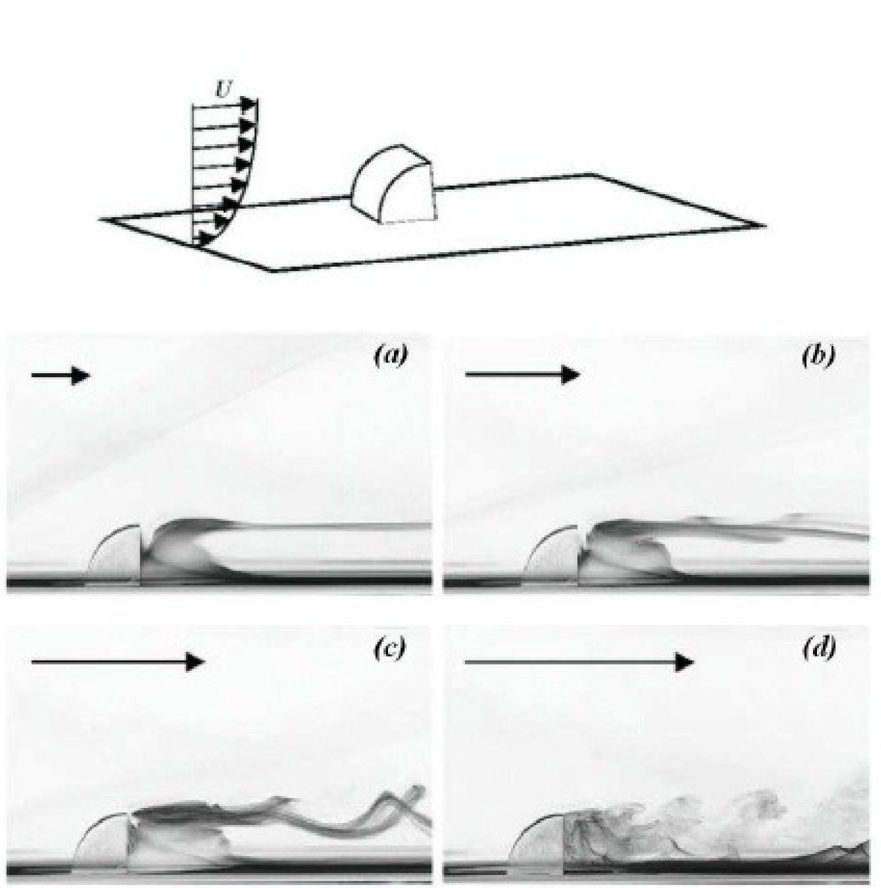
An introduction to turbulence in the atmospheric boundary layer

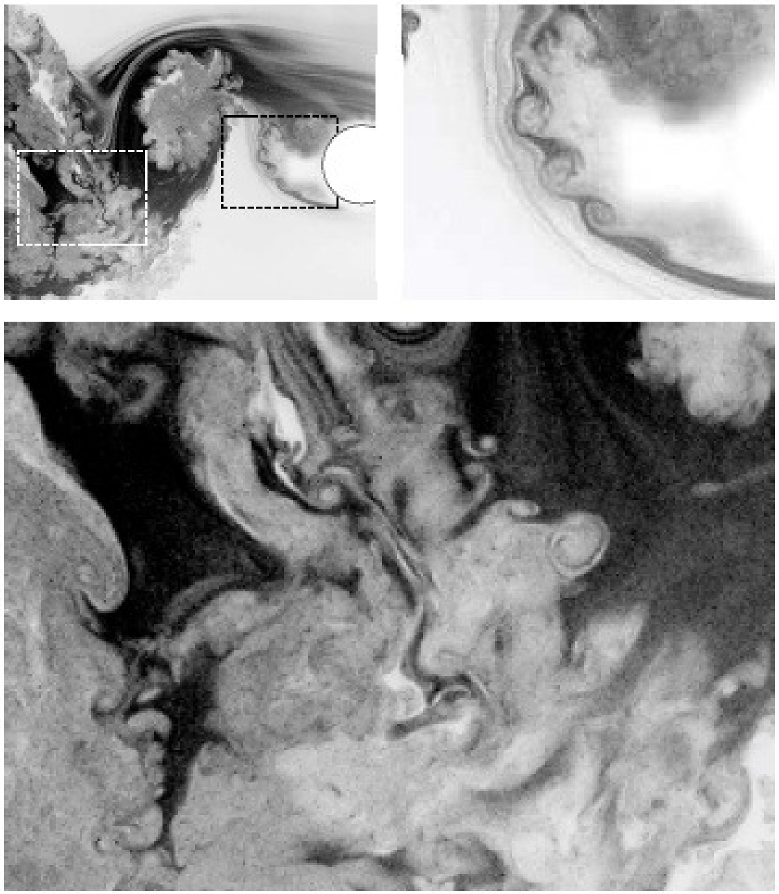
12/04/2024 – Xavier Fettweis

(and Charles Amory)

1. Introduction



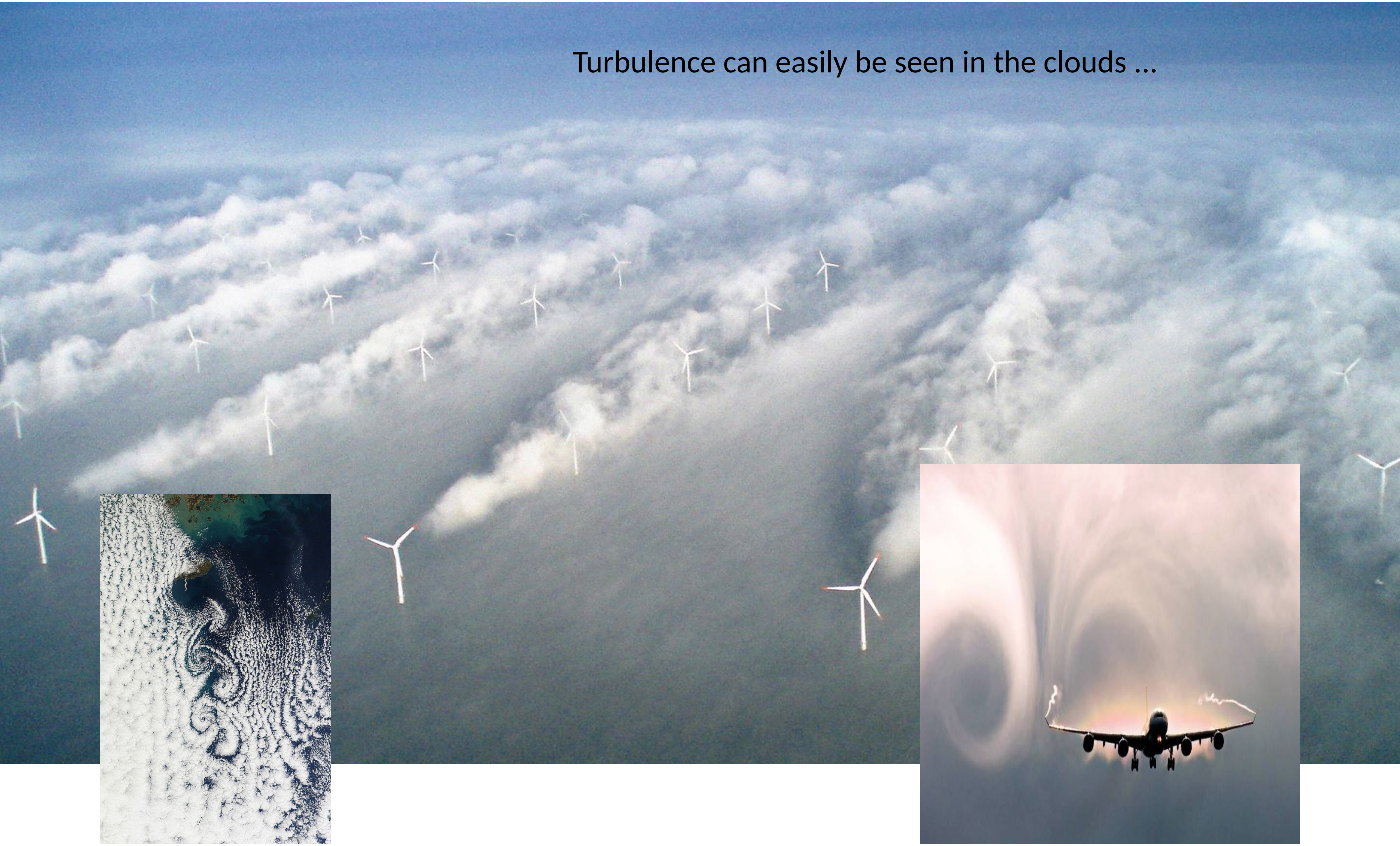
Changes in flow structure as a function of speed U.
a) Stationary trail, b) and c) periodic unsteady trail
d) turbulent trail



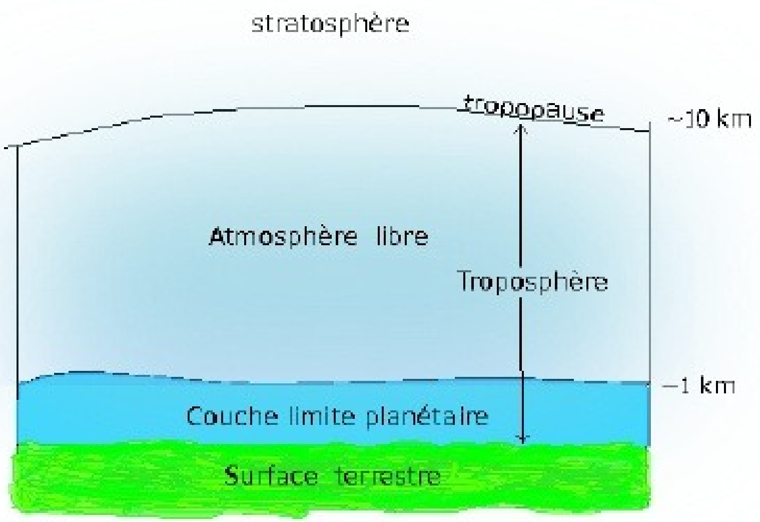
Turbulent trail of a cylinder

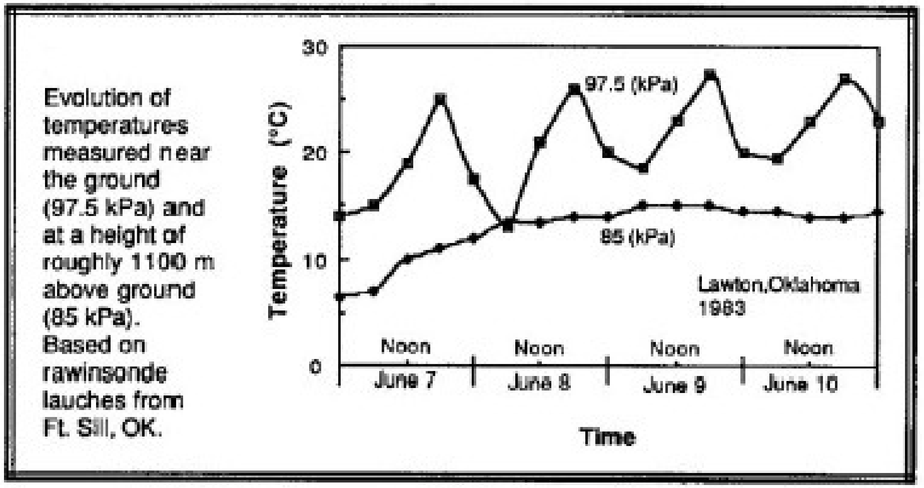
1. Introduction

Turbulence can easily be seen in the clouds ...



2. Turbulence in the ABL





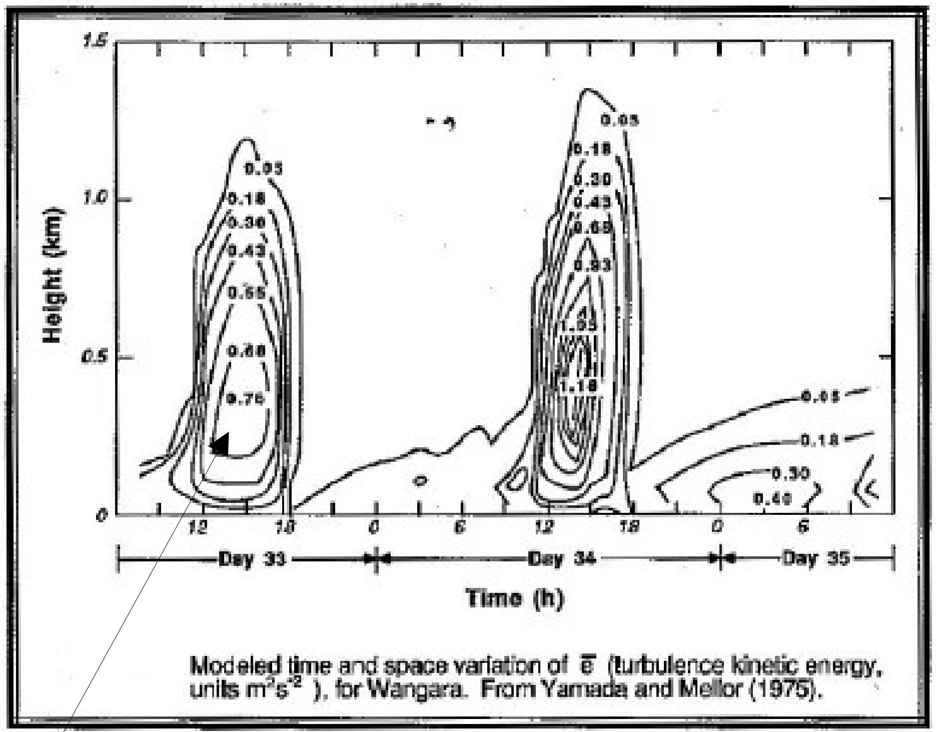
- $\tau_{max} \leq$ day ($\tau_{min} =$ s)
- Interacts with the surface : - friction (strong shear)
  - energy exchanges (solar heating, radiative cooling)
  - source of humidity
  - source of compounds (aerosols, dust, sand, snow etc.)
- Spatially and temporally variable height (day 1-2 km, night 100-300 m)
- Lower wind speeds than upper layers ($u_s \rightarrow 0$)
- Strong diurnal cycles (wind, temperature, humidity, compounds)
- **Strongly (almost continuously) turbulent**

"Aim" of turbulence: removing heterogeneities

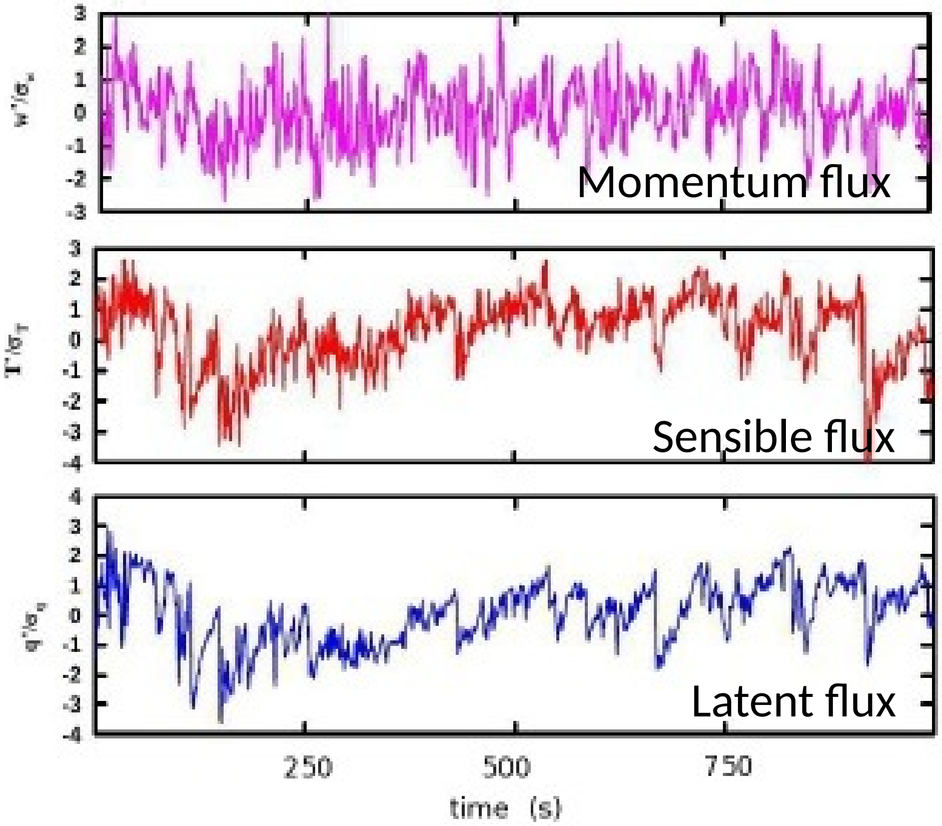
2. Turbulence in the ABL: ABL vs free atmopshere

Property	Boundary Layer	Free atmosphere
Turbulence	**Almost continuously turbulent** over its whole depth	Turbulence in convective clouds Sporadic clear atmosphere turbulence in thin layers of large horizontal extent
Friction	Strong drag against the surface Large energy dissipation	Small viscous dissipation
Dispersion	Rapid horizontal and vertical turbulent mixing	Small molecular diffusion Often rapid horizontal transport by wind
Winds	Near logarithmic wind speed profile in the surface layer Sub-geostrophic, cross-isobaric flow common	Nearly geostrophic
Vertical transport	Turbulence dominates	Mean wind and cumulus-scale dominate
Thickness	Varies between 100 m and 3 km in time and space Diurnal oscillations over land	Less variable 8-19 km slow time variations

2. Turbulence in the ABL: properties and origin



Modeled time and space variation of $\overline{e}$ (turbulence kinetic energy, units m^2s^{-2}), for Wangara. From Yamada and Mellor (1975).

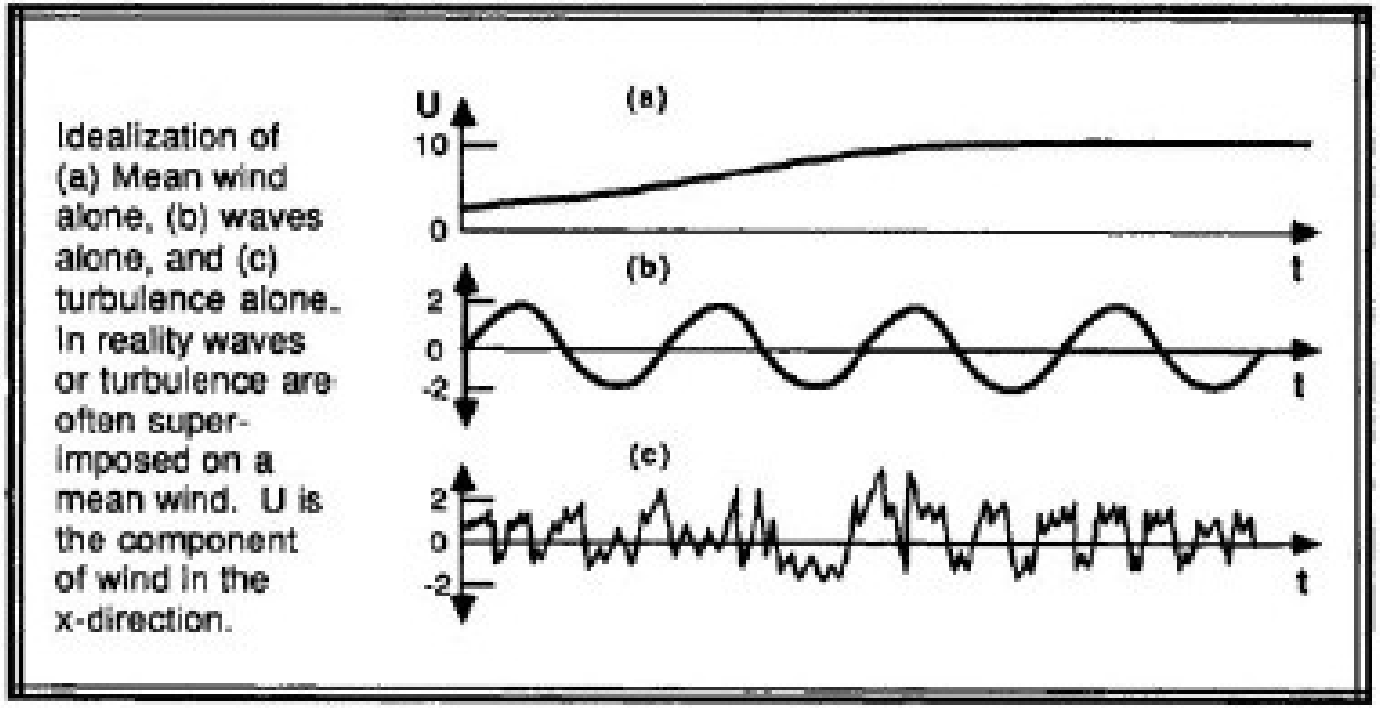


TKE $= \frac{1}{2}(\sigma_u^2 + \sigma_v^2 + \sigma_w^2) = \frac{1}{2}\left(\overline{(u')^2} + \overline{(v')^2} + \overline{(w')^2}\right)$

- Transmit to the atmosphere the variable characteristics of the Earth's surface
- Vertical (temperature) and horizontal mixing (wind)
- Random, fast irregularities (non-linear effects and unsteadyness) of atmospheric quantities resulting from the superposition of swirling motions (eddies) of different time and spatial scales

2. Turbulence in the ABL: role and definition

Components of the airflow



Synoptic scale

(i.e. mean wind)

Mesoscale

(i.e. gravity waves)

Small scale

(i.e. turbulent eddies)

- Diffusive character (mixing, transport of atmospheric quantities) ≠ waves
- Dissipative character (existence of small-scale strong gradients subject to viscous dissipation)
- Random character (small perturbations amplify rapidly)
- Two classic causes: **hydrodynamic** and **thermal** instabilities

2. Turbulence in the ABL: role and definition

hydrodynamic instabilities (wind shear)

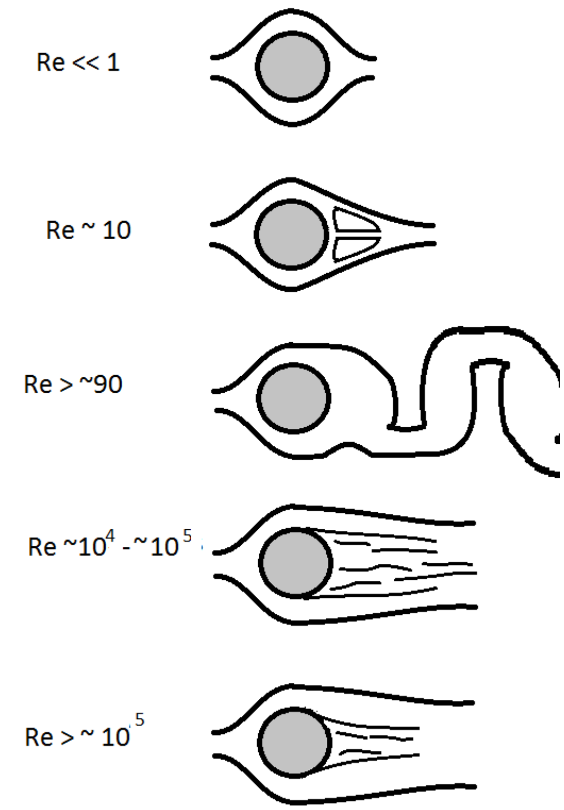
$$R_e = \frac{inertia\ force}{viscous/friction\ force} = \frac{dynamic\ pressure}{shearing\ stress}$$

- **Reynolds number**

$$R_e = \frac{U^2/L}{\nu U/L^2} = \frac{UL}{\nu}$$

where:

ρ is the density of the fluid (SI units: kg/m3)
u is the flow speed (m/s)
L is a characteristic linear dimension (m)
μ is the dynamic viscosity of the fluid (Pa·s or N·s/m2 or kg/(m·s))
ν is the kinematic viscosity of the fluid (m2/s).



$R_e \ll 1$: linear equation (diffusive processes dominant) ⇒ steadyness (laminar flows)

$R_e \gg 1$: non − linear equation (convective processes dominant) ⇒ unsteady effects (turbulence)

2. Turbulence in the ABL: role and definition

hydrodynamic instabilities (wind shear)

Kelvin-Hemlholtz waves

$$R_e = \frac{UL}{\nu}$$

U_{ABL} : 10 m/s

L_{ABL} : 1000 m

ν_{ABL} : 10^{-5} m^2/s (air viscosity)

$$R_{e\,ABL} \sim \frac{10 \times 1000}{10^{-5}} = 10^9 \gg 1$$

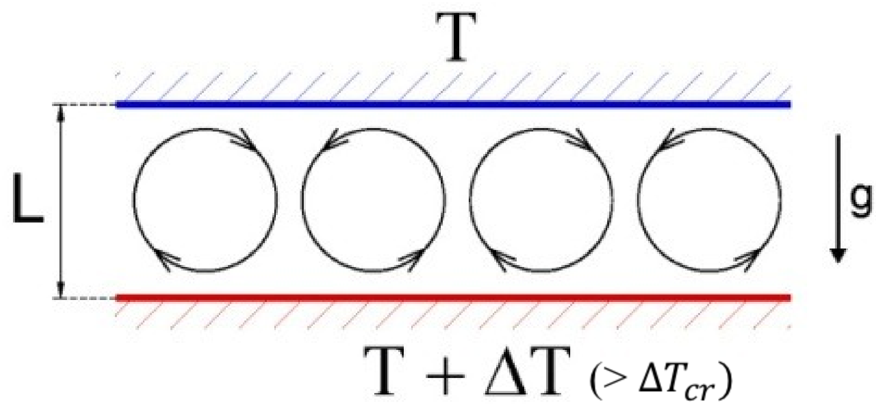
Atmospheric tropical cyclone R_e~ 1 x 10^{12}

2. Turbulence in the ABL: role and definition

$$\mathrm{Ra} = \frac{\text{time scale for thermal transport via diffusion}}{\text{time scale for thermal transport via convection at speed } u}.$$

Rayleigh number

thermal instabilities (temperature stratification)



$$R_a = \frac{\alpha g L^3 \Delta T}{\nu \nu_\theta}$$

$R_a > -1700$: stable (molecular fluxes only)

$-5\ 10^4 < R_a < -1700$: convection

$R_a < -5\ 10^4$: convective tubulent motions

ΔT : 1 K

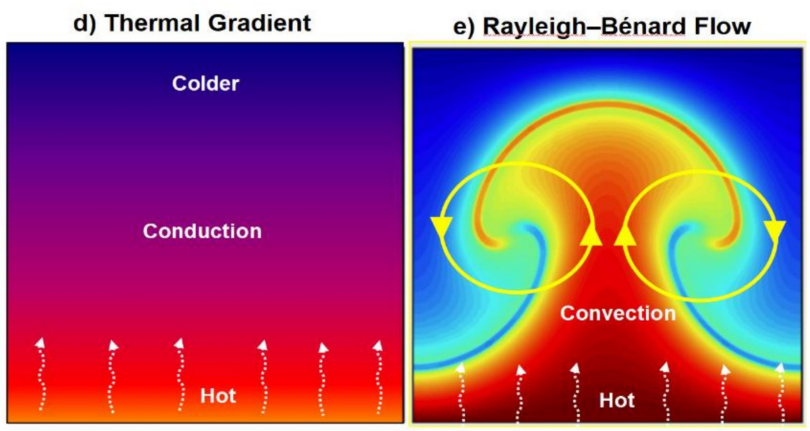
L_{ABL} : 1000 m

ν_{ABL} : 10^{-5} m^2/s (air viscosity)

$\nu_{\theta\, ABL}$: 2 10^{-5} m^2/s (air thermal diffusivity)

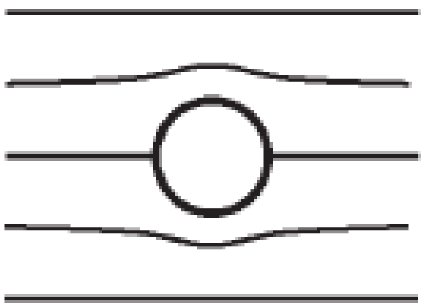
g : 10 m/s^2

α : 3 10^{-3} K^{-1} (perfect gaz)



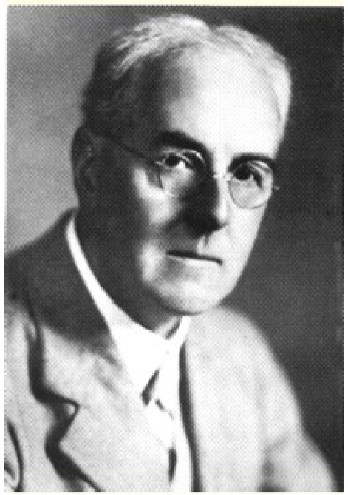
2. Turbulence in the ABL: production of a range of scales

Turbulent flow
$R_e \gg 1$
Multiple flow scales

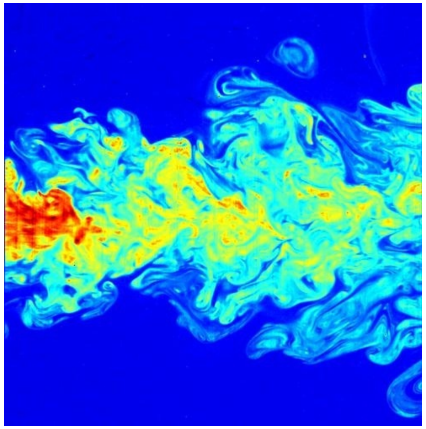


Viscous (laminar) flow
$R_e \ll 1$
One flow scale :
$\delta \propto d/\sqrt{R_e}$

2. Turbulence in the ABL: Richardson's energy cascade



Lewis Fry Richardson
1881-1953



L, U

$R_e = \frac{UL}{\nu} \gg 1$

No viscous dissipation

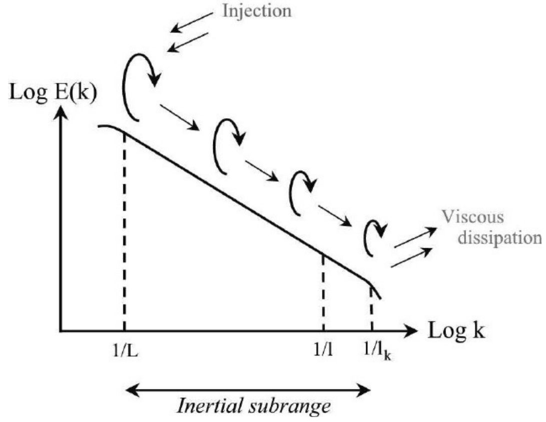
η, u_η

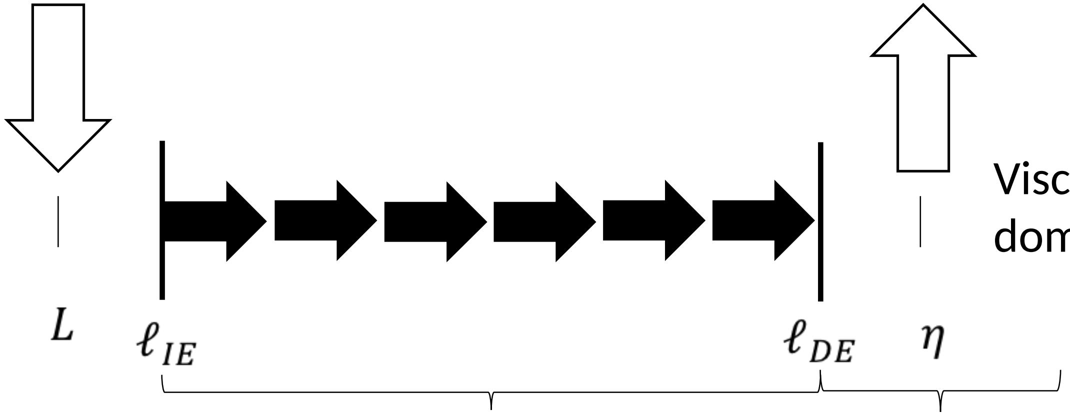
$R_e = \frac{\eta u_\eta}{\nu} = 1$

Viscous dissipation

big eddies → energy transfer → *little eddies* → energy transfer → *lesser eddies*

Production ***Dissipation***



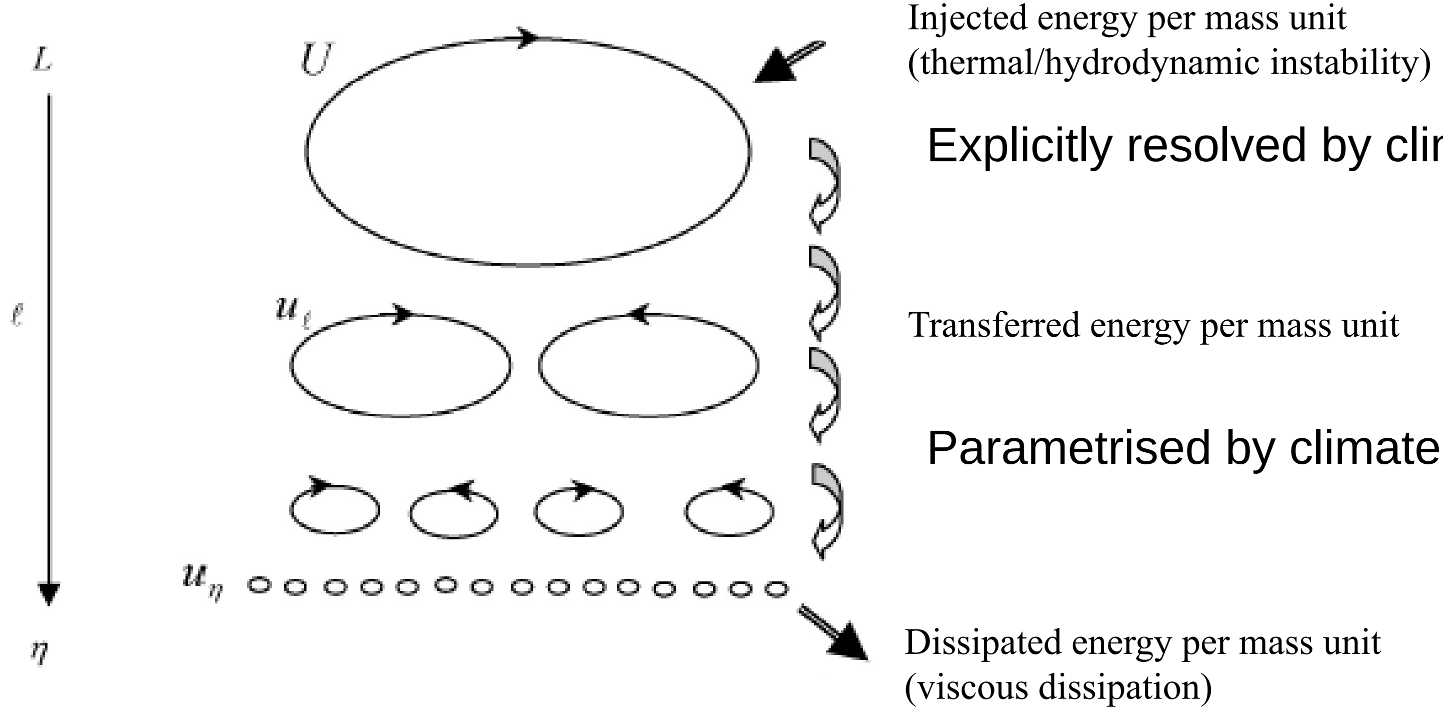


Viscosity is dominating

Inertial ranges

Dissipative range

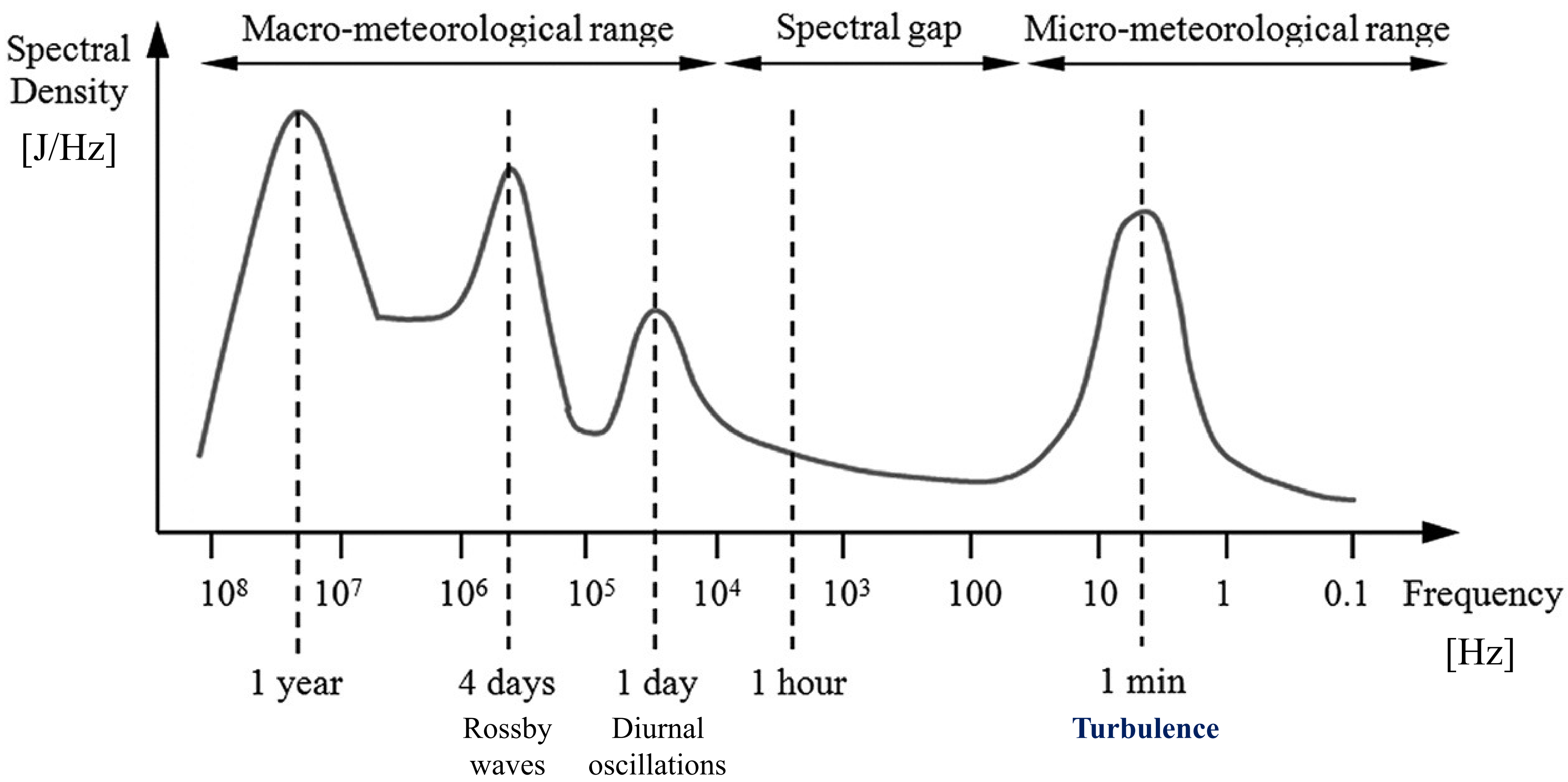
2. Turbulence in the ABL: Kolmogorov theory (1941)



- Kinetic energy transferred between eddies from the larger to the smaller structures
- Energy of (initial) larger structures provided by the mean flow (rapid dissipation if not maintained)
- Size of larger structures constrained by the geometry of the flow
- Size of smaller structures linked to dissipative effects (Kolmogorov scale η)
- In practice, theory valid for $R_e \geq 10^6$ (expansive numerical costs..)

For a plane of 10m flying at 250m/s: Re ~ 100 meaning η ~ 1 cm

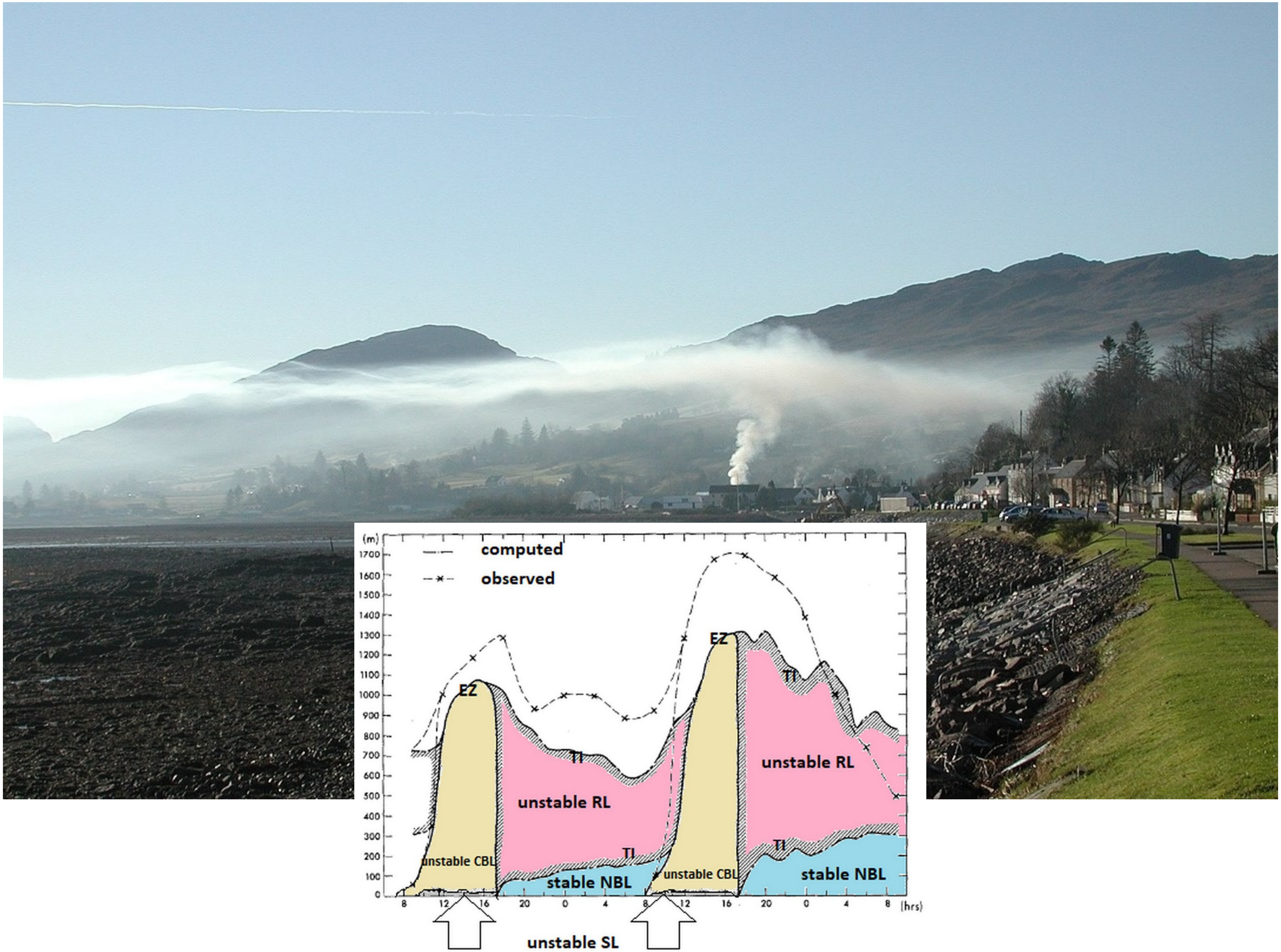
2. Turbulence in the ABL: Spectral gap



Explicitly resolved by climate models

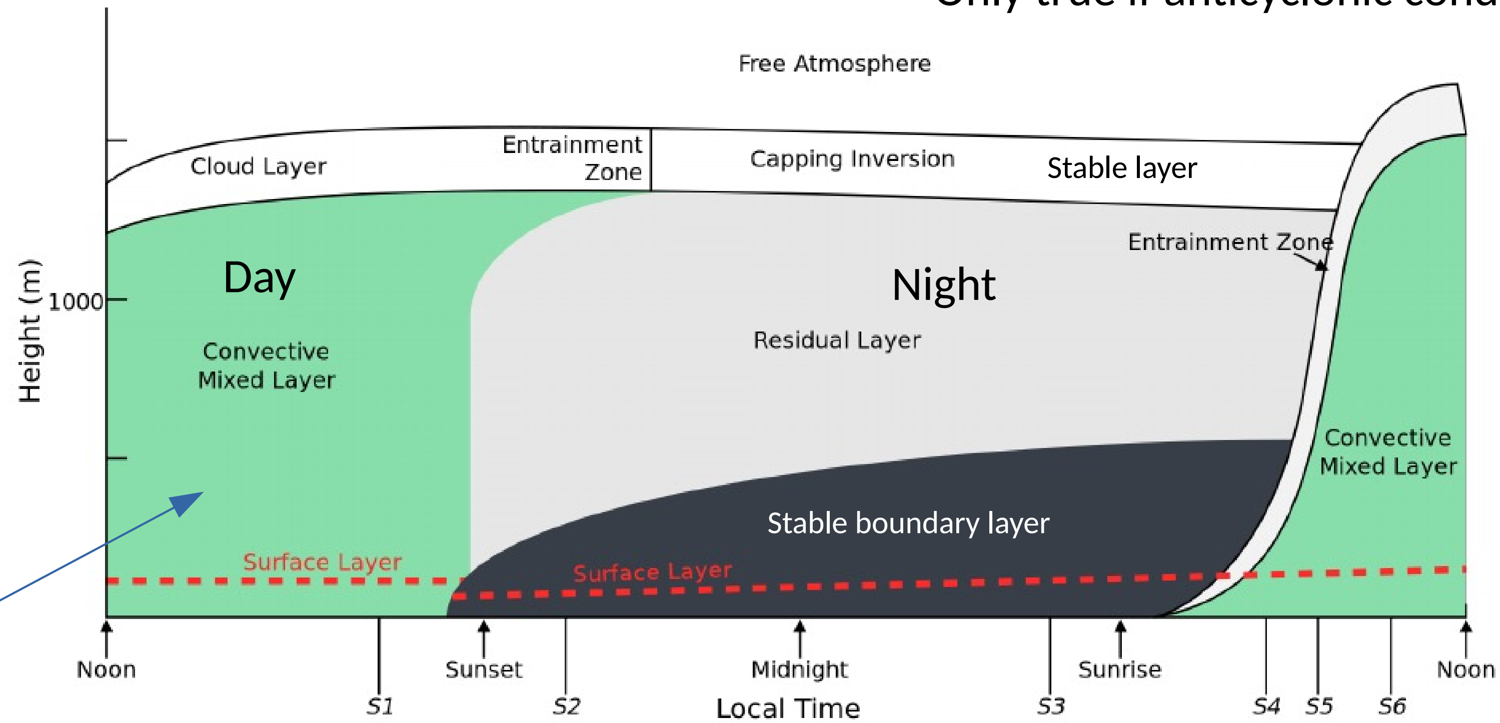
Parametrised by climate models

3. Daily cycle of the boundary layer

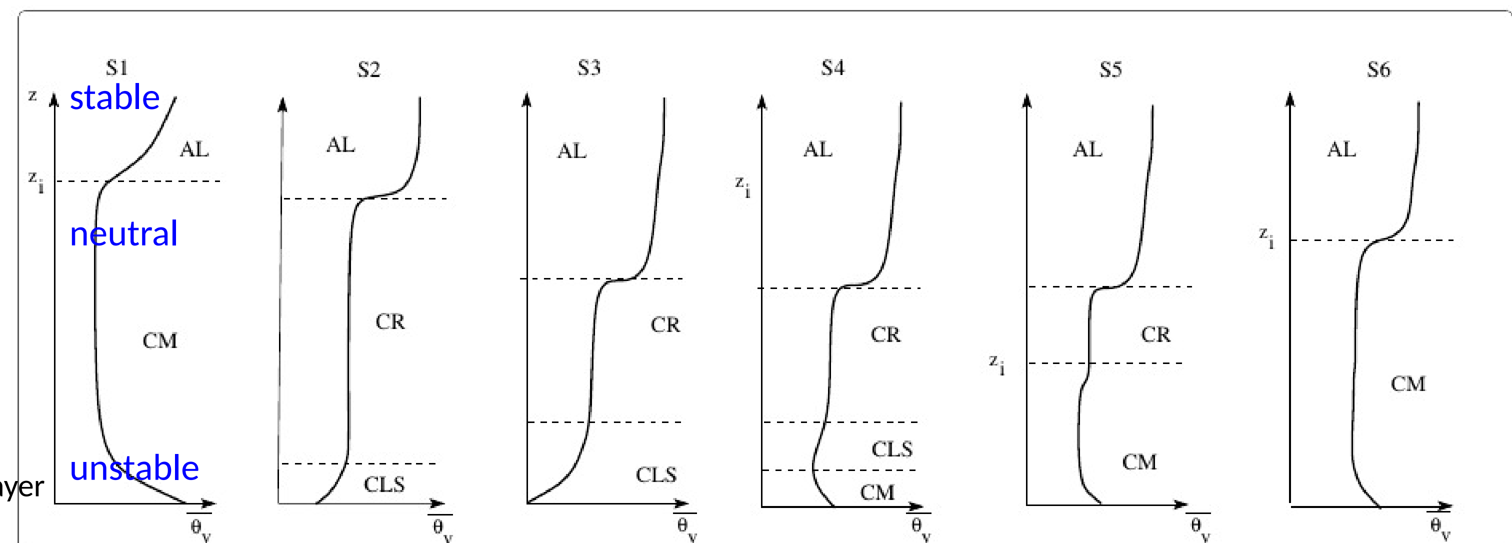


3. Daily cycle of the boundary layer

Only true if anticyclonic condition !!



Free or forced convection

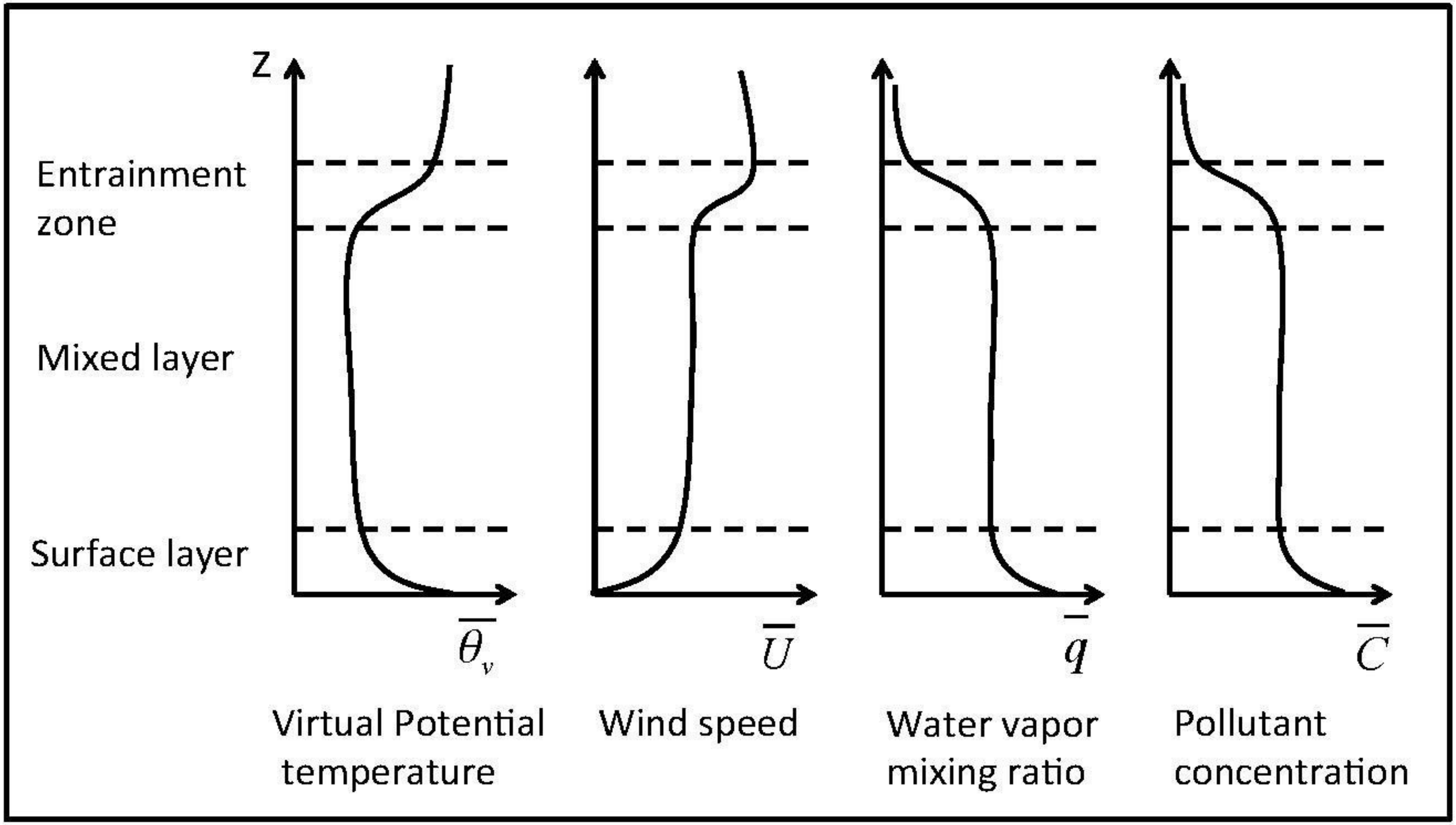


Profile of the virtual potential temperature

CM = mixing layer
CLS = stable boundary layer
CR = residual layer
AL = free atmosphere

3. Daily cycle of the boundary layer

At noon, if anticyclonic conditions, ...



3. Daily cycle of the boundary layer: pollutant

During night ...

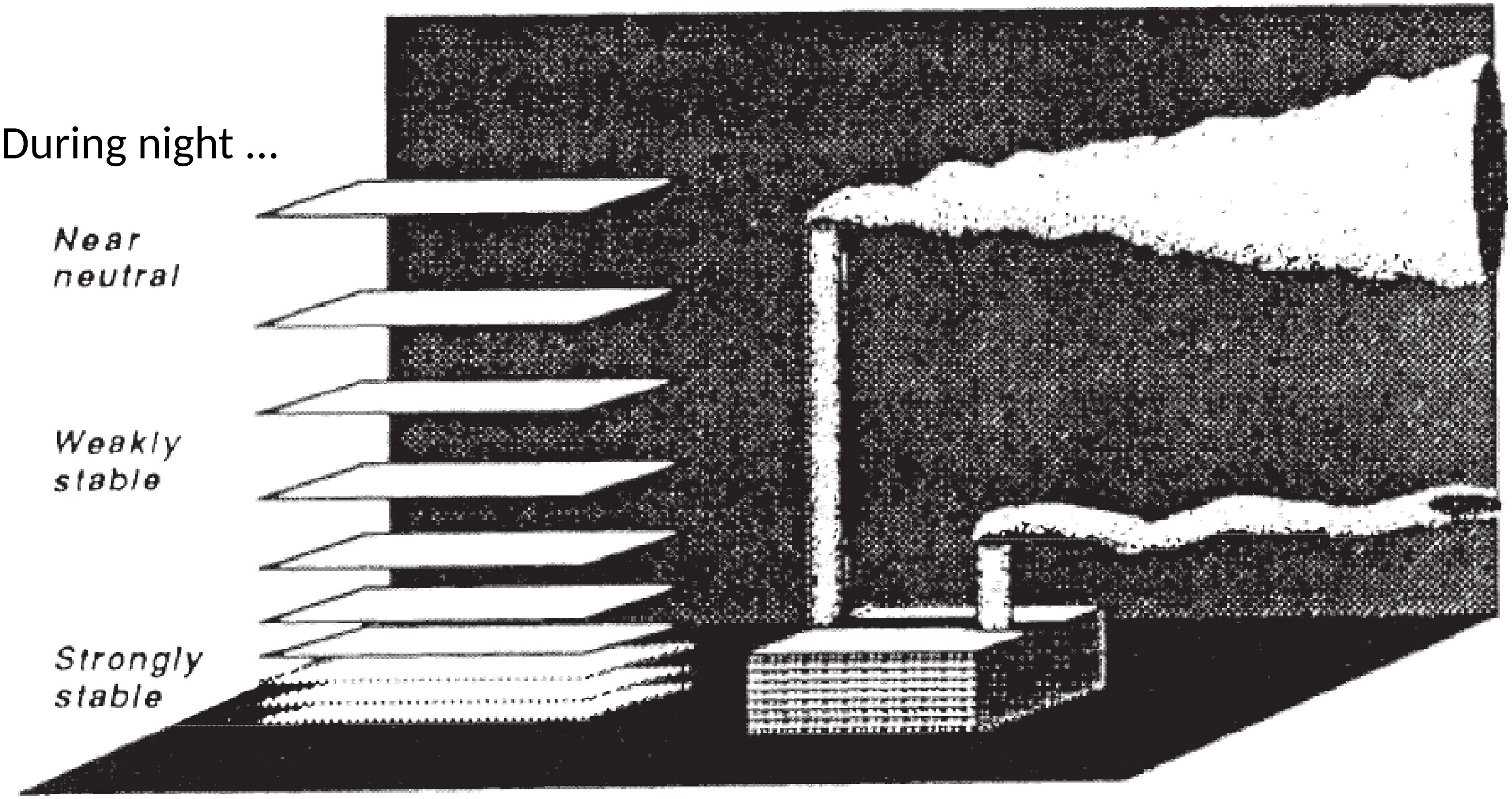


Fig. 1.10 The static stability decreases with height in the nocturnal boundary layer, gradually blending into the neutrally-stratified residual layer aloft, as indicated by the isentropic surfaces sketched on the left. Smoke emissions into the stable air fan out in the horizontal with little vertical dispersion other than wavelike oscillations. Smoke emissions in the neutral residual-layer air spread with an almost equal rate in the vertical and horizontal, allowing the smoke plume to assume a cone-like shape.

3. Daily cycle of the boundary layer: low level jet

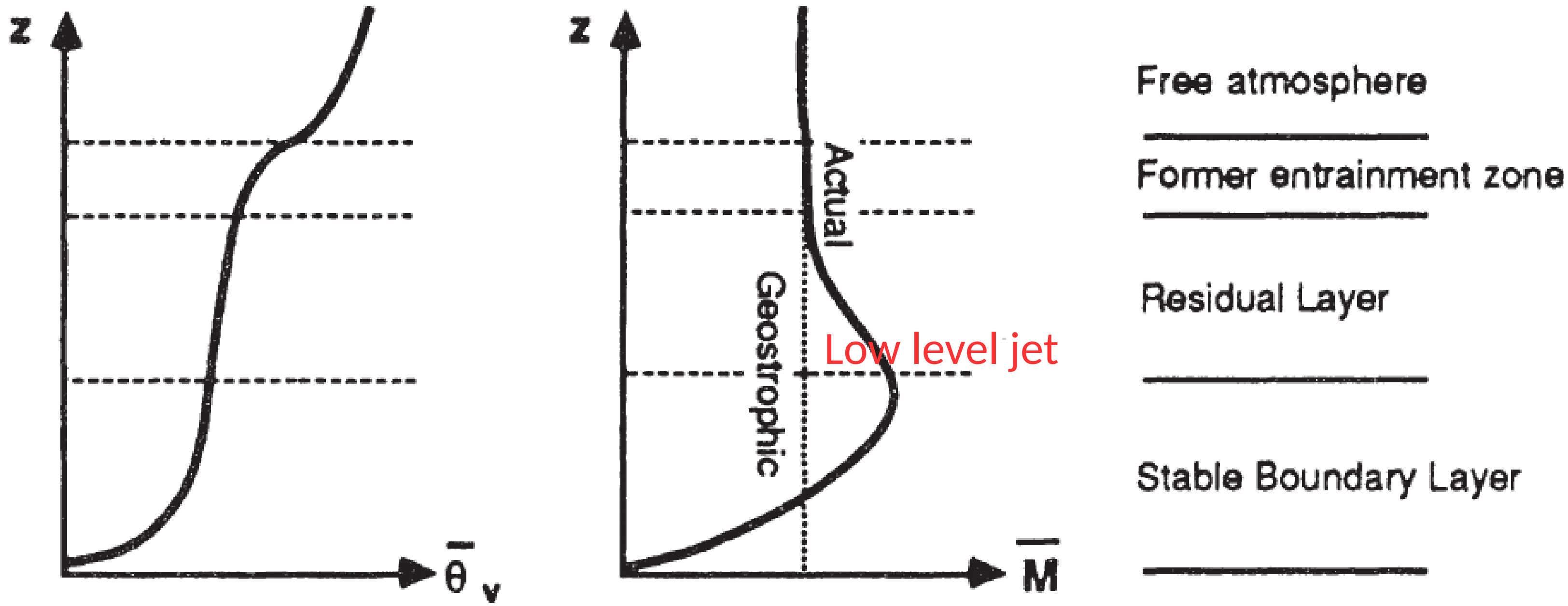


Fig. 1.11 Mean virtual potential temperature, $\overline{\theta}_v$, and wind speed, $\overline{M}$, profiles for an idealized stable boundary layer in a high-pressure region.

Low level jet: decrease of drag during night due to very stable boundary layer

3. Daily cycle of the boundary layer: low level jet

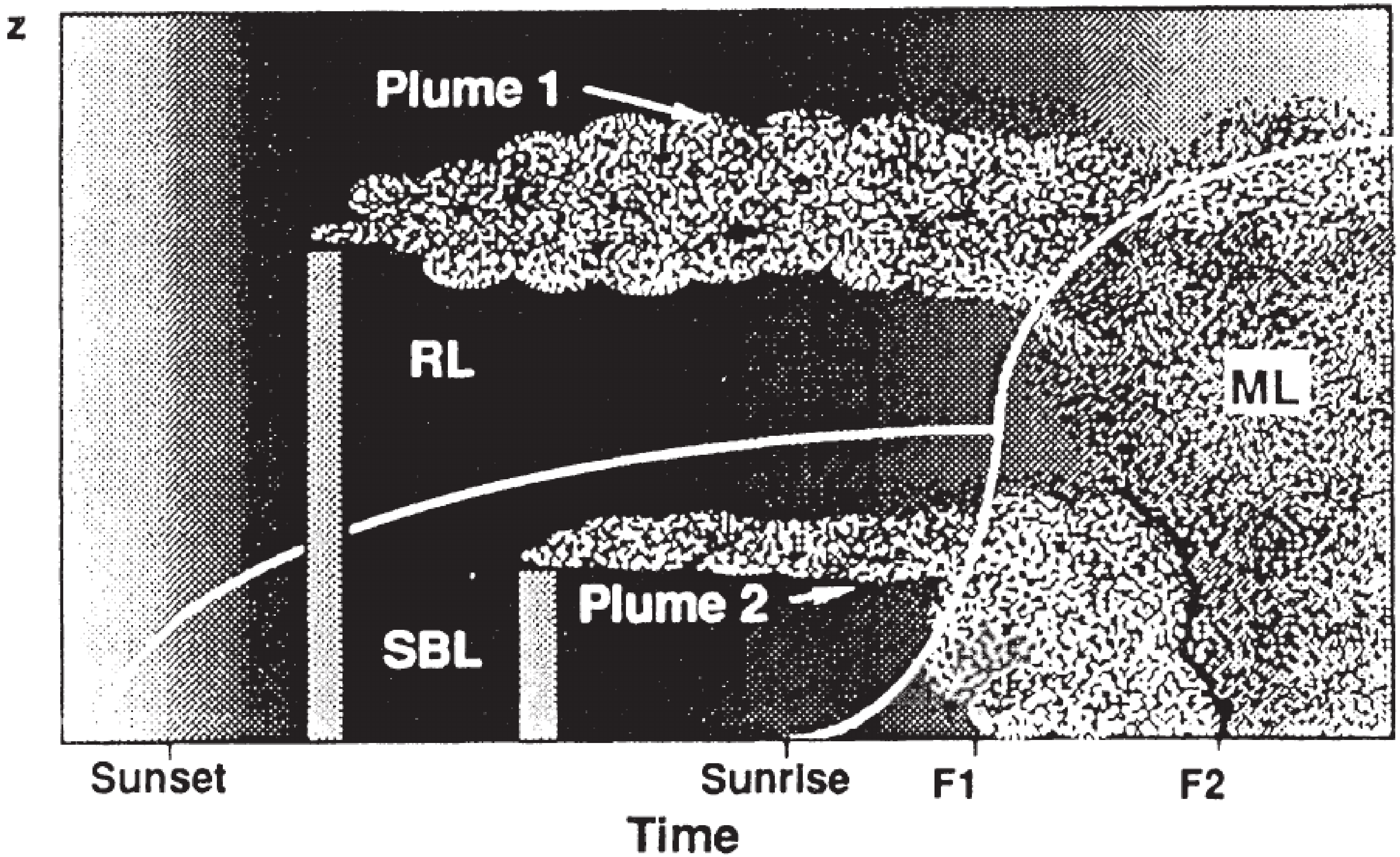
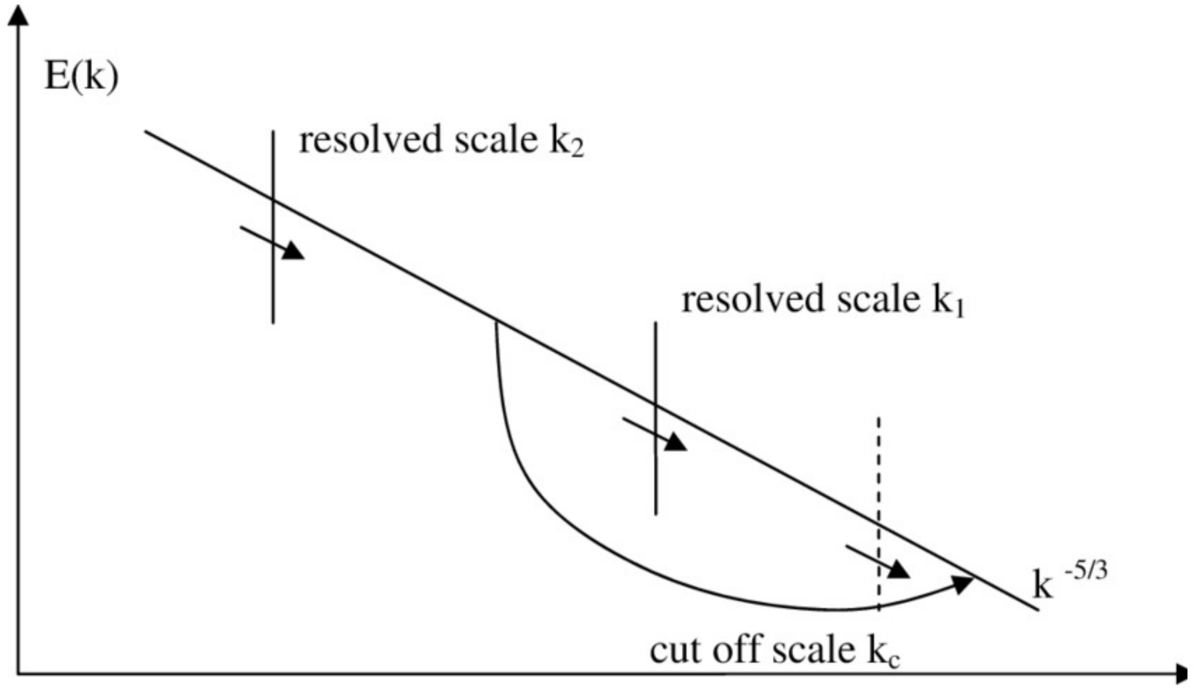
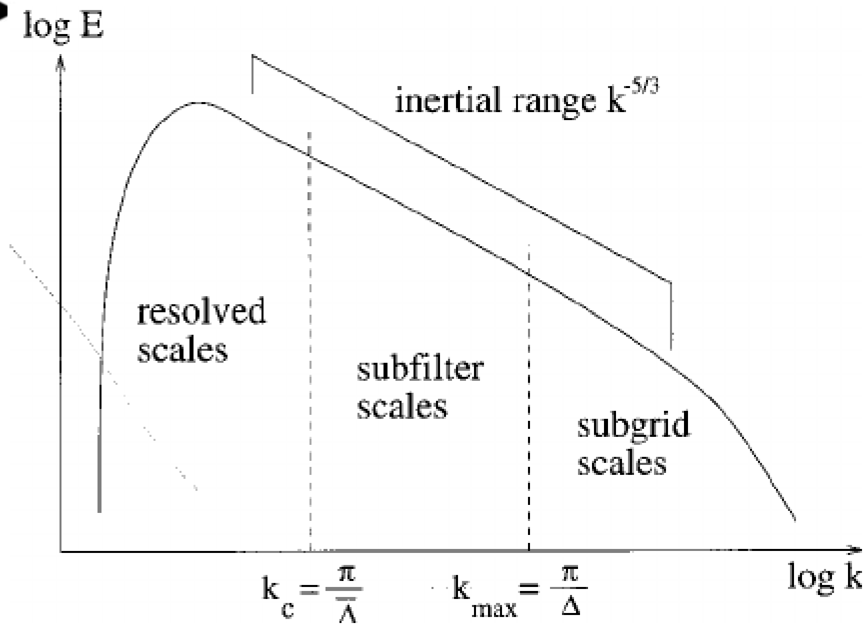


Fig. 1.14 Sketch of the fumigation process, where a growing mixed layer mixes elevated smoke plumes down to the ground. Smoke plume 1 is fumigated at time F1, while plume 2 is fumigated at time F2.

4. Turbulence closure





4. Turbulence closure: Reynolds decomposition

instantaneous wind → mean wind → turbulence contribution

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

30-min averages (typically) to isolate turbulent fluctuations

and similarly for θ and q..

$$uw = (\bar{u} + u')(\bar{w} + w') = \overline{uw} + \bar{u}w' + \bar{w}u' + u'w'$$

$$\overline{x'} = 0$$

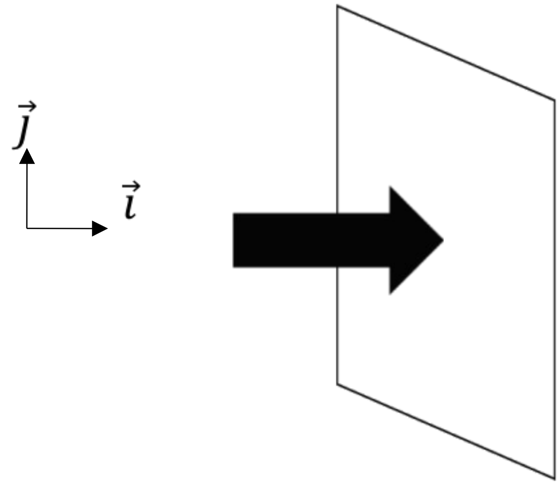
$$\overline{\bar{x}y'} = \bar{x}.\overline{y'} = 0$$

$$\overline{uw} = \bar{u}\,.\,\bar{v} + \overline{\boldsymbol{u'w'}}$$

Covariances = turbulent fluxes (2nd order moments)

4. Turbulence closure: covariance

$cov_{xy} = \overline{x'y'}$: degree of shared variability between two variables
$cov_{xy} = 0$: turbulent fluctuations x' and y' are independent

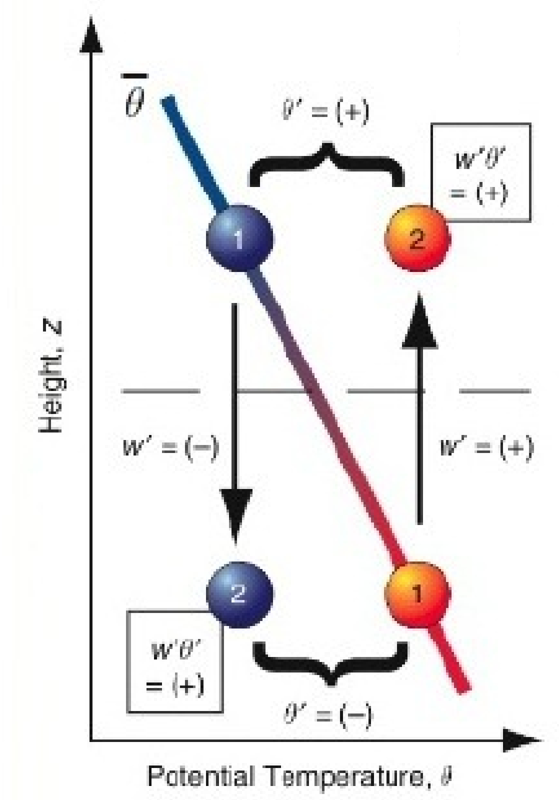


Turbulent flux $\overline{u_i'\theta'}$, $\overline{u_i'q'}$, $\overline{u_i'u_j'}$: transport of the corresponding quantity through a plane perpendicular to the direction i, by wind speed fluctuations due to turbulence motions in the direction i

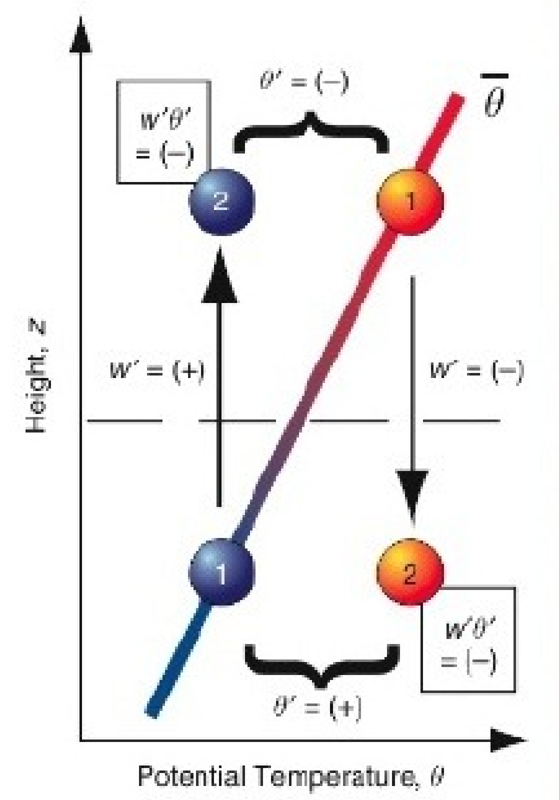
Quantity	Associated (kinematic) flux	Formulation	Units
Temperature	Vertical sensible heat flux	$\rho L\overline{w'\theta'}$	W/m^2
Humidity	Vertical latent heat flux	$\rho C_p\overline{w'q'}$	W/m^2
Wind	Vertical flux of horizontal momentum	$\rho\overline{u'w'}$	N/m^2 = Pa

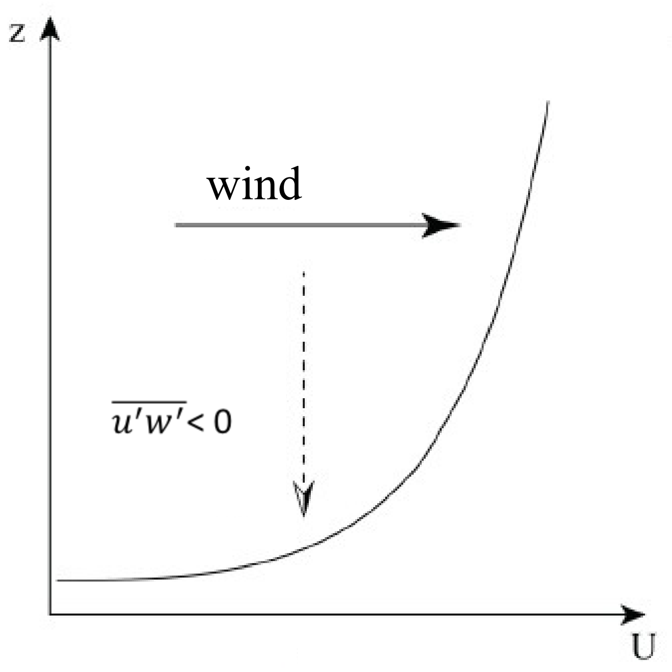
4. Turbulence closure: covariance

Unstable ABL



Stable ABL





Upward heat transport

Downward heat transport

Downward momentum transport

4. Turbulence closure: covariance

- Reynlods decomposition into a mean field and a fluctuating field add unknowns to the system
- The number of unknowns is larger than the number of equations
- When we start deriving equations for unknowns, more variables appear

Closure problem:
Total statistical description of turbulence requires an infinite set of equations

Prog. Eq. for :	Moment	Equation	Nb of Eqs.	No. unknowns
$\overline{U_i}$	First	$\frac{\partial \overline{U_i}}{\partial t} = \cdots - \frac{\partial \overline{u_i' u_j'}}{\partial x_j}$	3	6
$\overline{u_i' u_j'}$	Two	$\frac{\partial \overline{u_i' u_j'}}{\partial t} = \ldots - \frac{\partial \overline{u_k' u_i' u_j'}}{\partial x_k}$	6	10

4. Turbulence closure: the closure problem

For example, for the first order we have 6 unknowns: $\overline{u'^2}, \overline{v'^2}, \overline{w'^2}, \overline{u'v'}, \overline{u'w'}, \overline{v'w'}$.
We can derive prognostic equations for these correlation terms, but then the new equations contain additional triple correlation terms. If we were to derive prognosic equations for these they would contain fourth-order moments...

Solution:
Use only a finite number of prognostic equations and approximate the remaining unknowns.

- Closure approximations are named according to the highest order prognotic equations are retained

ex: 1^{st} order closure = prognostic equation for the mean variables, approximation for the 2^{nd} moments

- Approximations made using the derivated quantities of moments of lower order.

4. Turbulence closure: the K-theory

K-theory : Parametrizing the turbulent fluxes as proportional to local gradients (~ Fick law)

Molecular diffusion		K-theory
$\phi_i(a) = -\upsilon \frac{\partial \bar{a}}{\partial x_i}$	→ generalized by	$\overline{u_i'a'} = -K \frac{\partial \bar{a}}{\partial x_i}$

$$\overline{u'w'} = -K_m \frac{\partial \bar{u}}{\partial z}$$

$$\overline{v'w'} = -K_m \frac{\partial \bar{v}}{\partial z}$$

$$\overline{\theta'w'} = -K_\theta \frac{\partial \bar{\theta}}{\partial z}$$

$$\overline{q'w'} = -K_q \frac{\partial \bar{q}}{\partial z}$$

K: eddy viscosity, eddy diffusivity, eddy-transfer coefficient, turbulent-transfer coefficient...

4. Turbulence closure: properties of eddy diffusivities

Molecular diffusion		K-theory
$\phi_i(a) = -\upsilon \frac{\partial \bar{a}}{\partial x_i}$	→ generalized by	$\overline{u_i' a'} = -K \frac{\partial \bar{a}}{\partial x_i}$

- K [m^2/s] ~ viscosity
- = capacity of transferring a quantity (momentum, heat etc.) by turbulent mixing for a given gradient
- K > 0 (down-gradient transport from strongest values to lowest values)
- K >> υ (turbulence is much more effective at causing mixing than viscosity)
  - K ≈ 1 - 10 m^2/s
  - $\upsilon \approx 10^{-5}$ m^2/s
- $K_m \neq K_\theta \neq K_q$
- Horizontal gradients considered as negligible compared to vertical gradients
- K is a flow property ($\neq \upsilon$ = fluid property): varies in time and space; depends on local turbulence intensity

How to determine K?

4. Turbulence closure: the mixing length theory

Prandtl (1925): the mixing length l is the typical vertical depth over which parcels of air move before mixing with their surroundings (~ radius of a typical eddy)

$$K = u_* l$$

with l characteristic length scale of eddies and u_* characteristic speed scale of eddies ; [m^2/s]

K_m increases with wind shear (intensity of eddies) and the mixing length (size of eddies)

In the surface layer eddies are limited by the proximity of the surface (i.e. $\propto z$).
It is assumed that

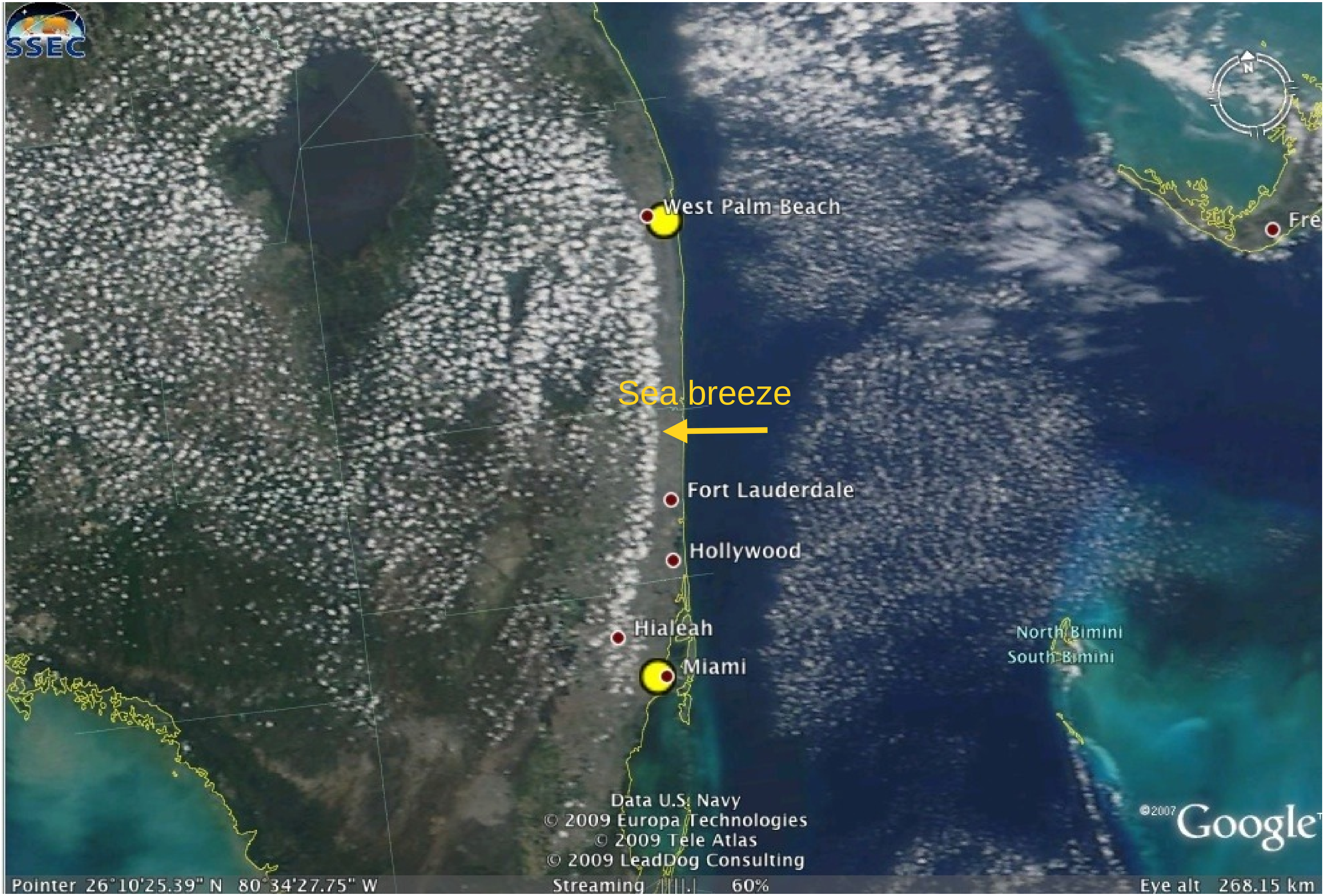
$$l = \kappa z$$

with κ the von Karman constant (=0.4)

Then

$$K_m = u_* \kappa z$$

5. Local winds and circulations



5. Local winds and circulations: sea breeze

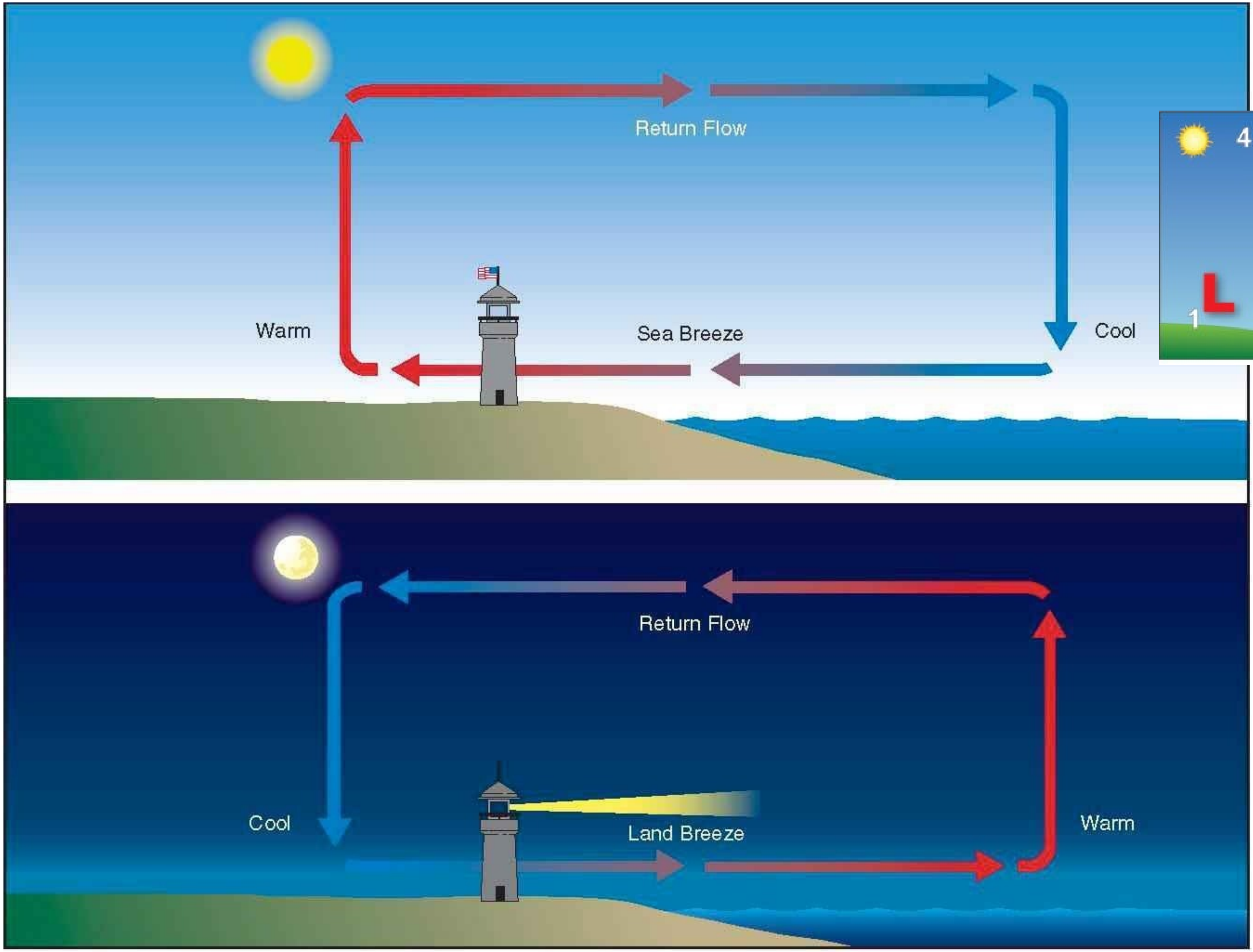
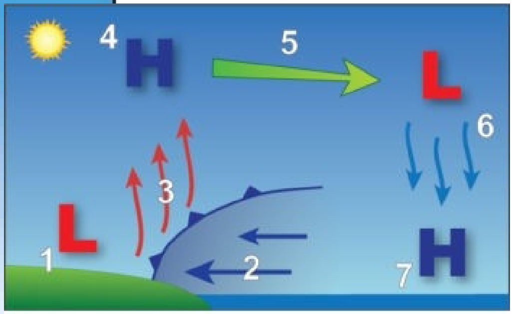
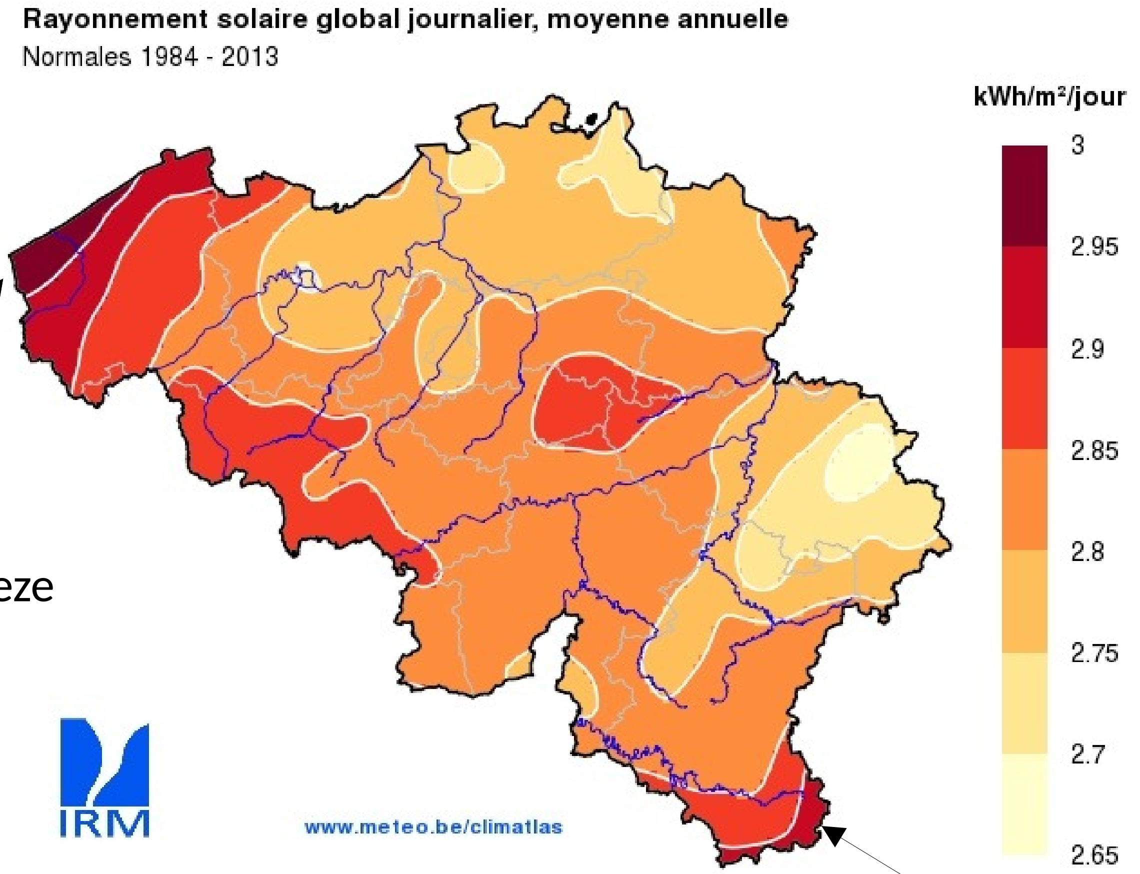


Figure 10-13. Sea breeze and land breeze wind circulation patterns.



5. Local winds and circulations: sea breeze

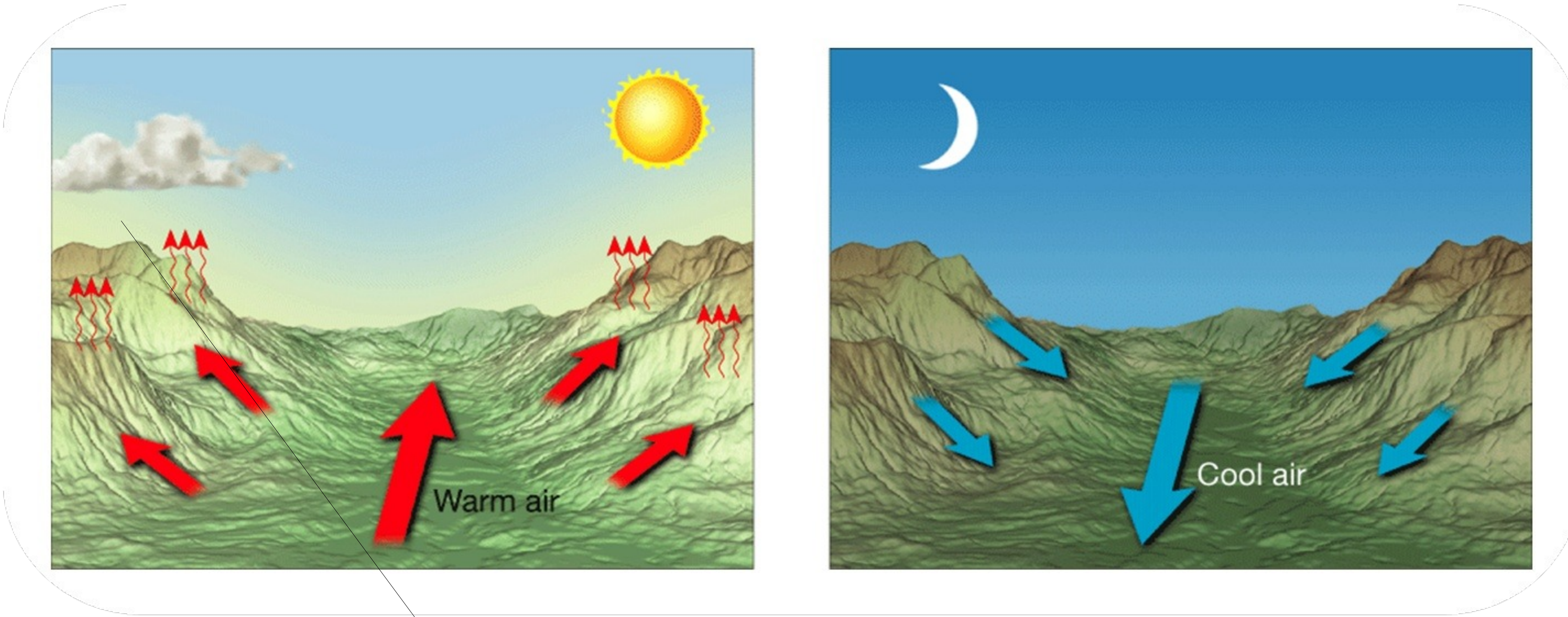
In Belgium … as observed



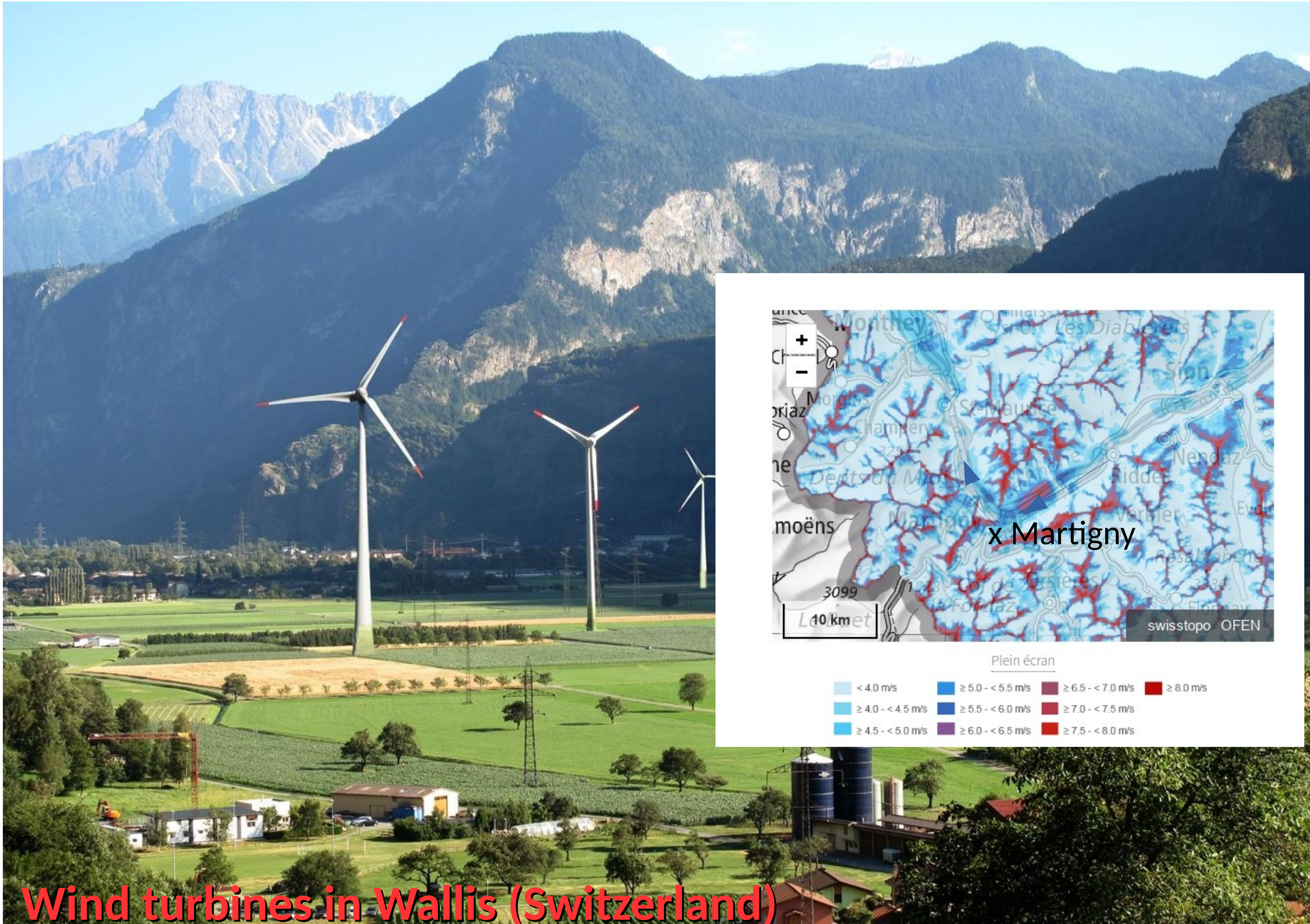
Due to the sea breeze

Due to the latitude +
more “continental” climate

5. Local winds and circulations: valley breeze

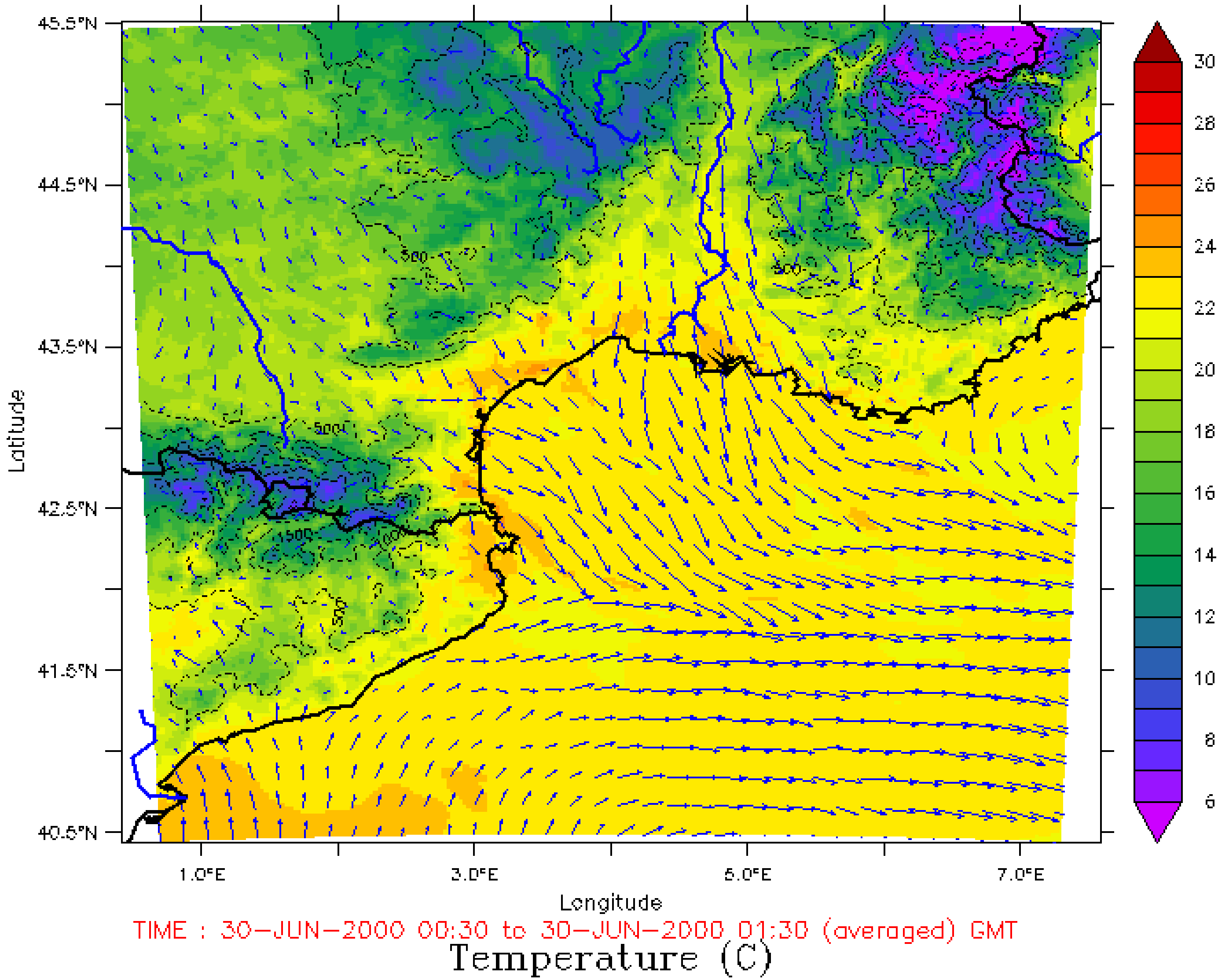


5. Local winds and circulations: katabatic wind



5. Local winds and circulations: Valley + sea breeze

As simulated by MAR...



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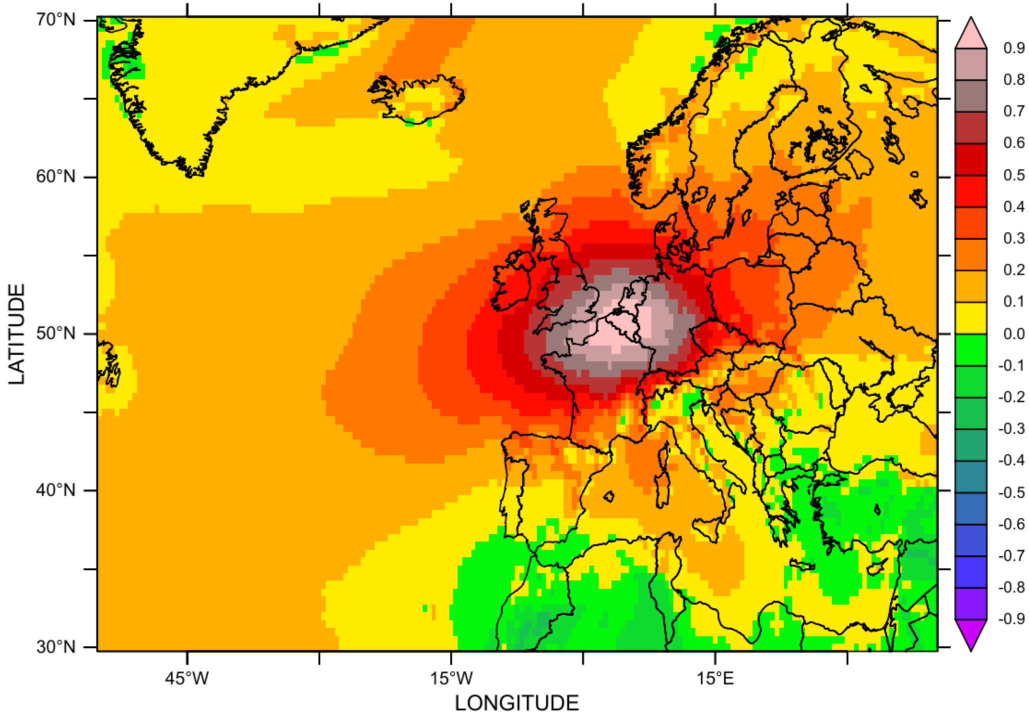
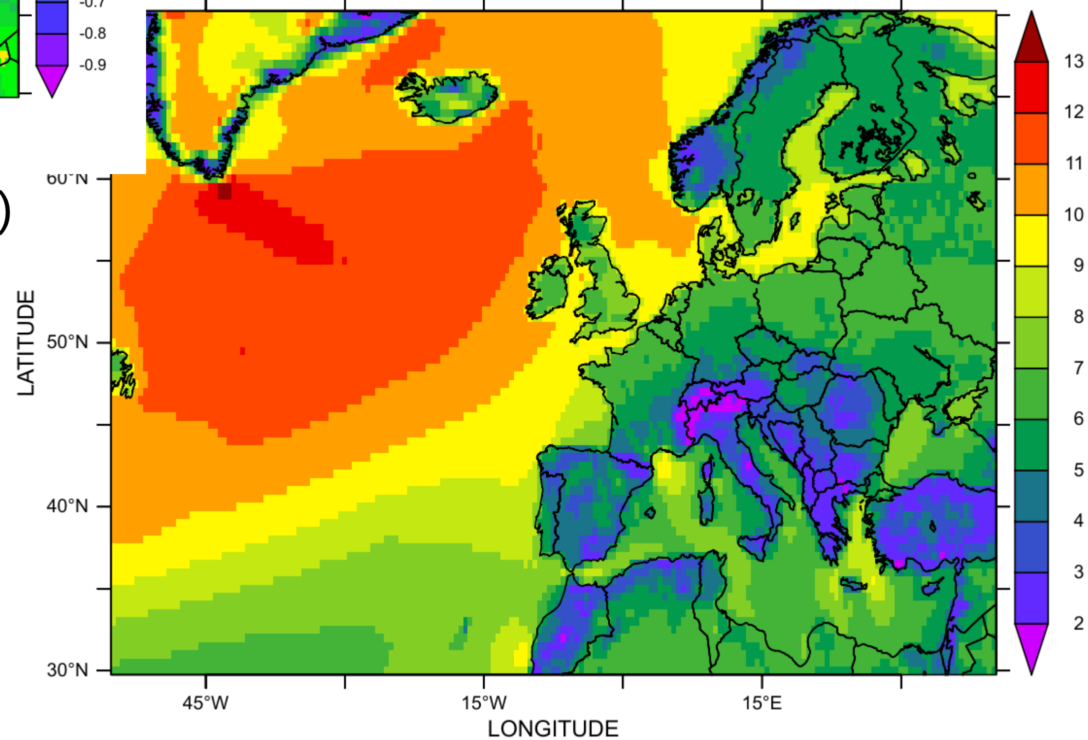


Correlation with Belgian wind speed (-)

Mean wind speed (m/s)



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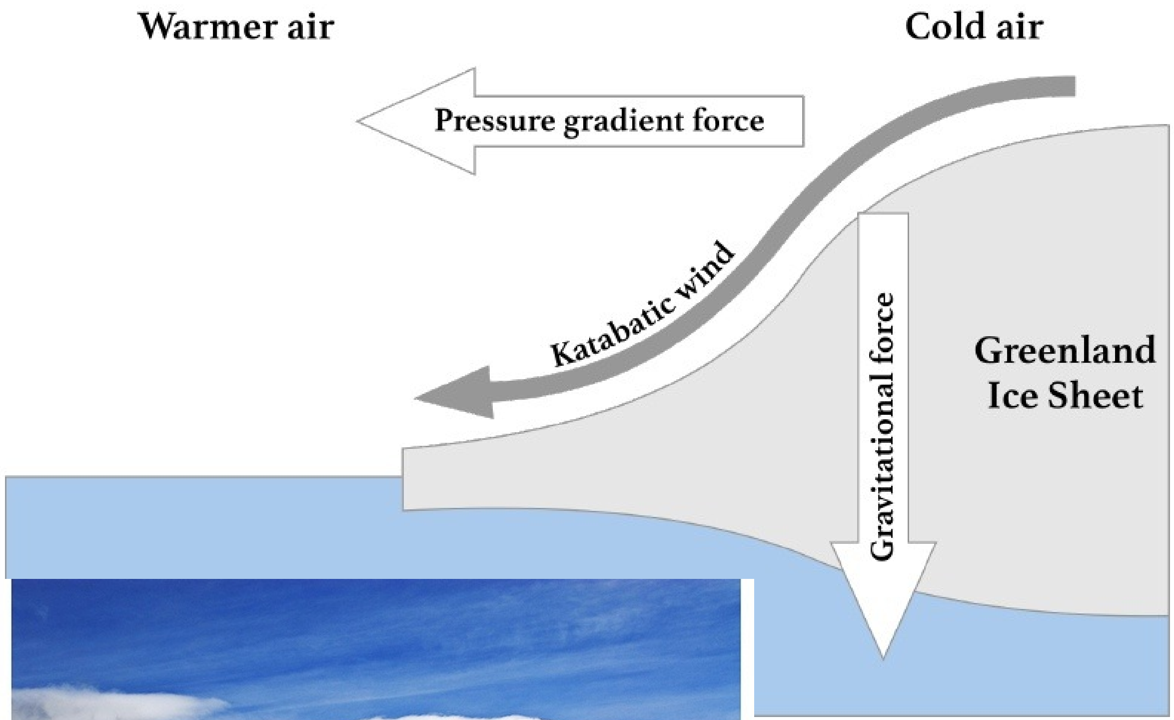


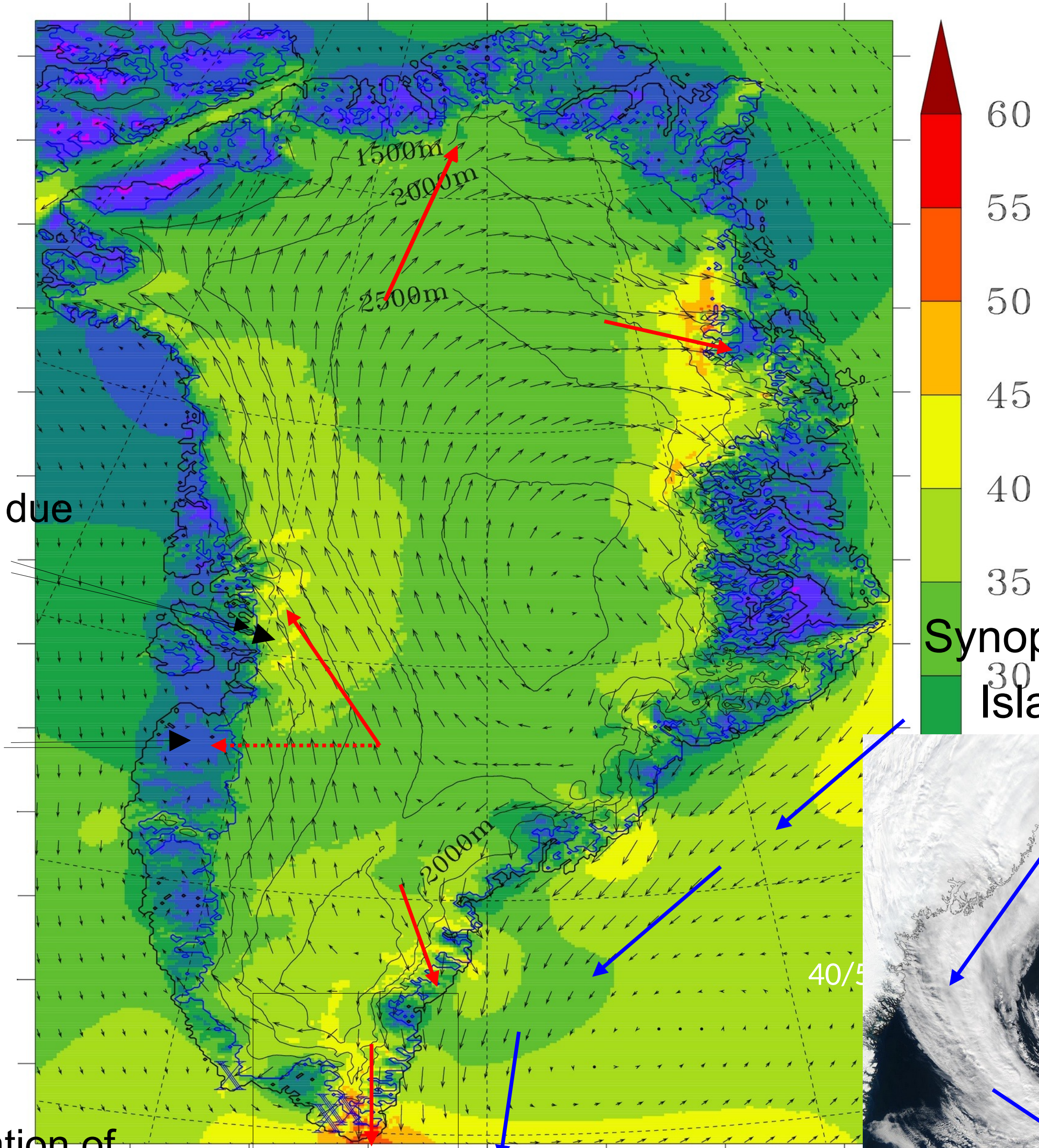
Katabatic wind

Katabatic winds are the result of heat transfer processes between the cold ice cap and the warmer air mass above it.

When the air mass temperature is higher than that of the ice sheet, the former is cooled down by radiation, thus the air density increases forcing it down the sloping terrain.

The flow of katabatic winds is driven by gravity, temperature gradient and inclination of the slope of the ice sheet.

5. Local winds and circulations: katabatic wind

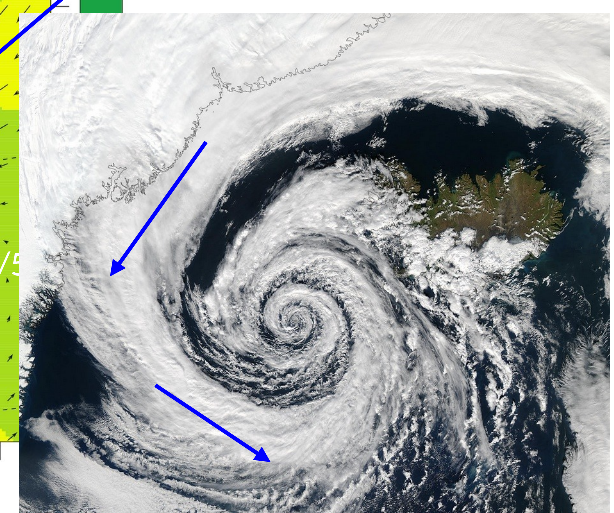


Real katabatic wind (deviated on its right due to the Coriolis force + Ekman)

Pure katabatic wind

In this area, combination of both katabatic and synoptic winds

Synoptic wind due to the Island Low Pressure

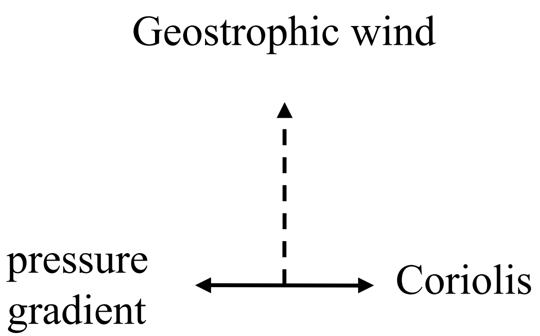


Mean Wind speed at 100m (km/h)

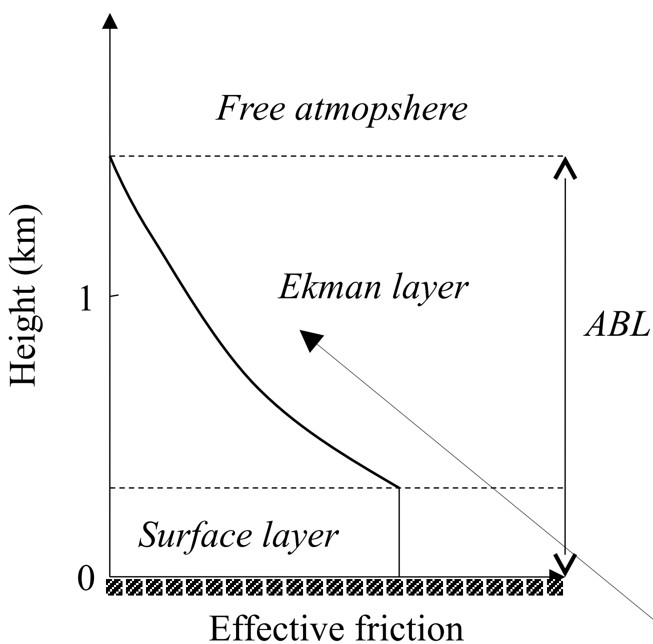
5. Local winds and circulations: katabatic wind

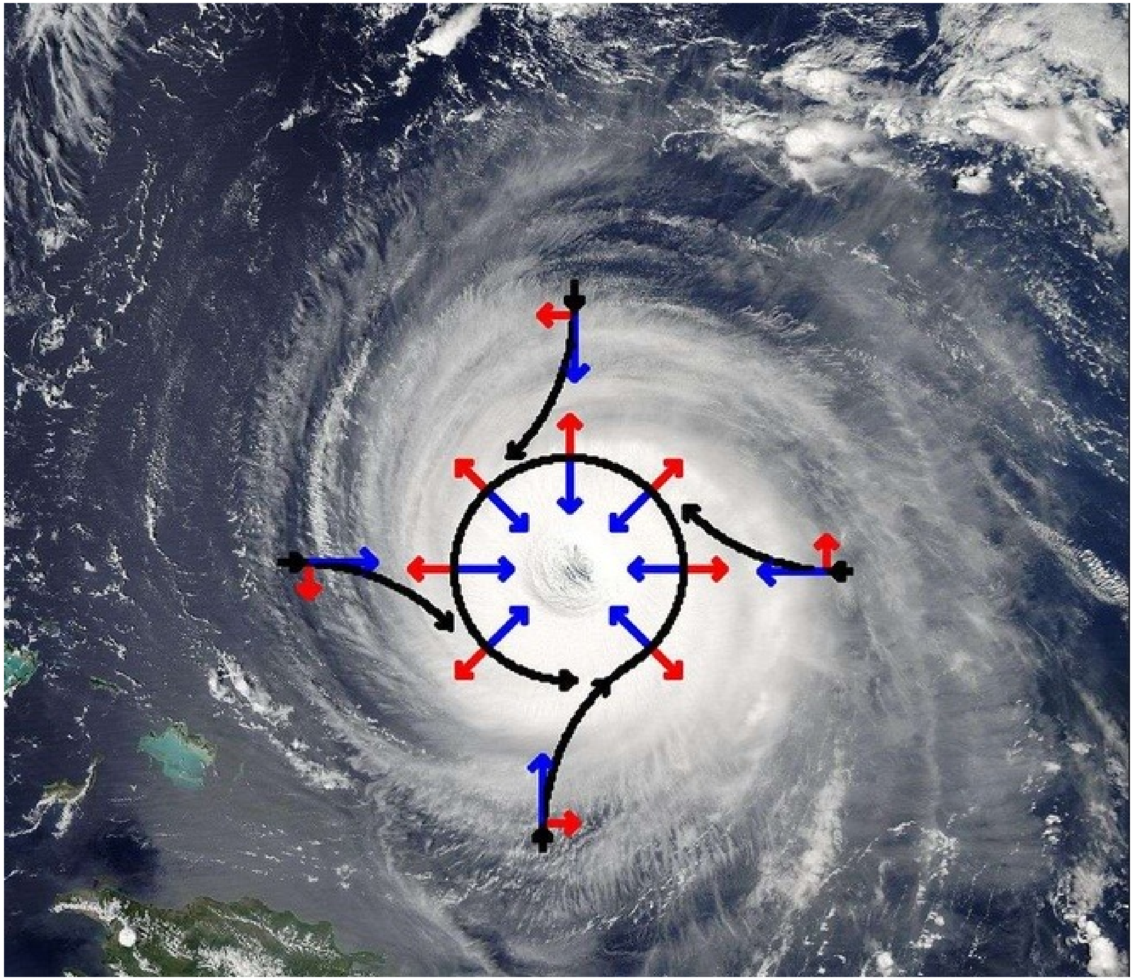
Wind turbines in Greenland: https://www.katabata-project.uliege.be/

5. Local winds and circulations: Ekman Layer



wind flows parallel to the isobars

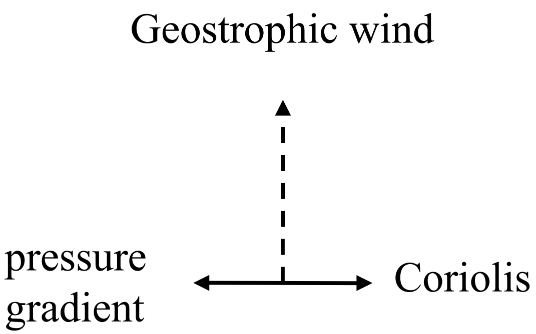




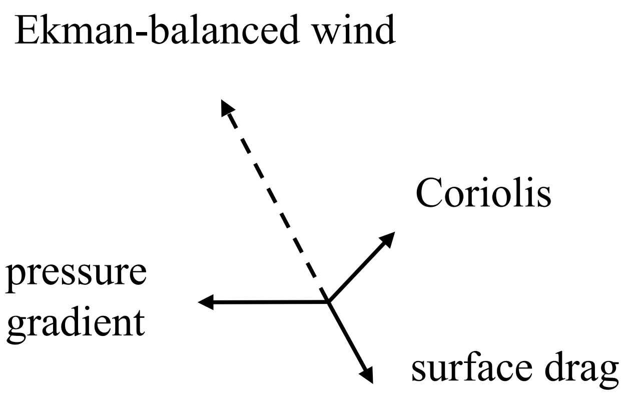
Depression in the Northern hemisphere. Windflow trajectory (black) as a resultant of pressure (blue) and Coriolis (red) forces.

Wind direction impacted by the surface...

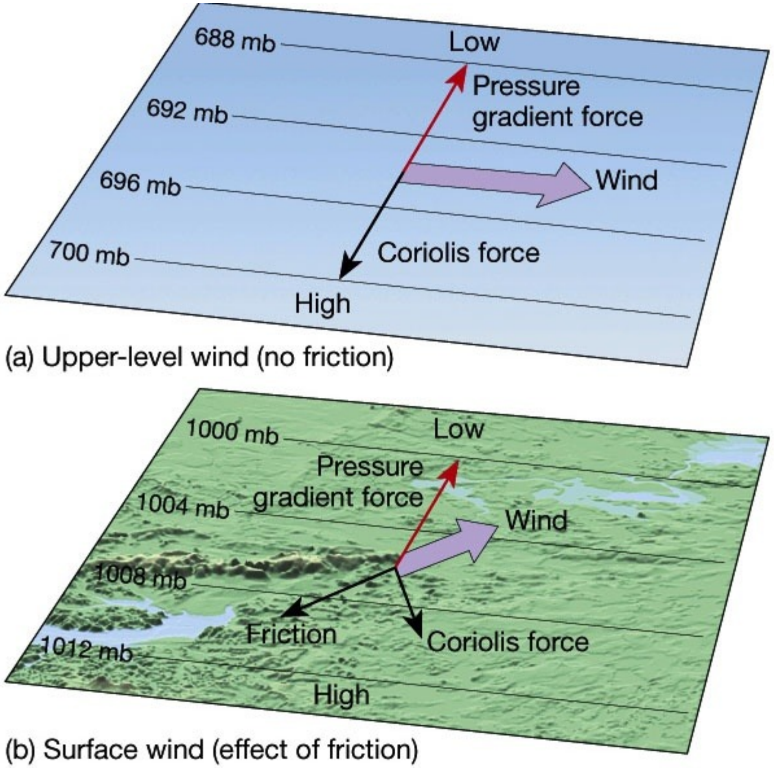
5. Local winds and circulations: Ekman Layer



wind flows parallel to the isobars

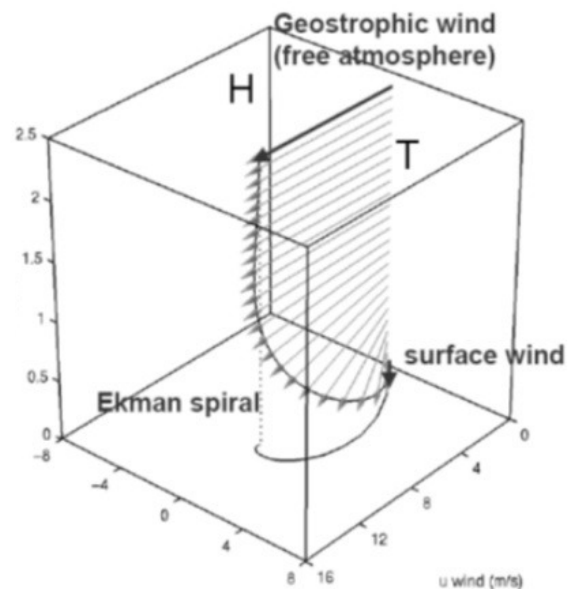


vertical rotation due the balance of surface drag by Coriolis forces



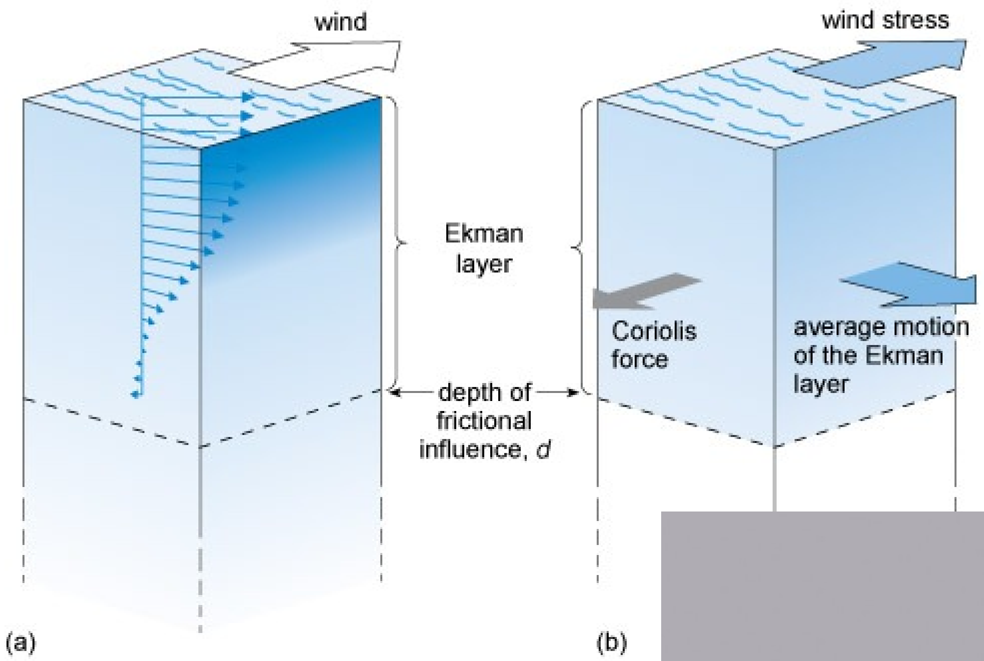
(a) Upper-level wind (no friction)

(b) Surface wind (effect of friction)



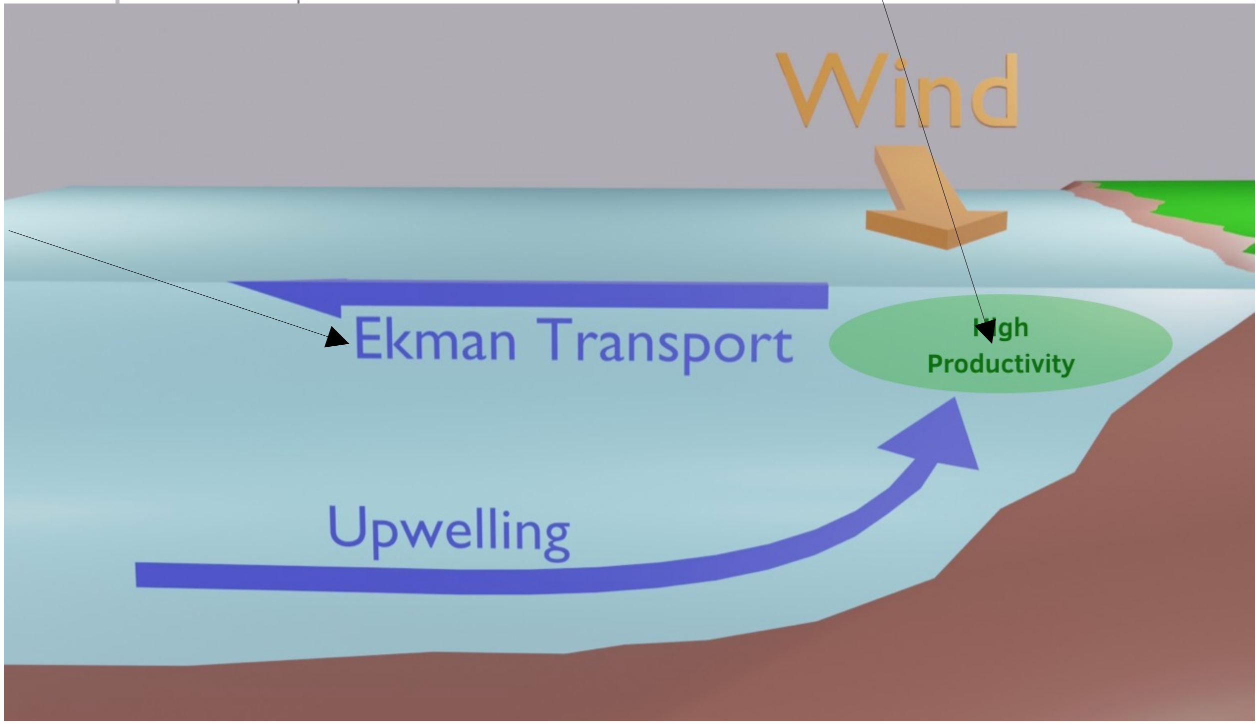
Atmospheric Ekman spiral matching the geostrophic flow in the free atmosphere

5. Local winds and circulations: Ekman Layer

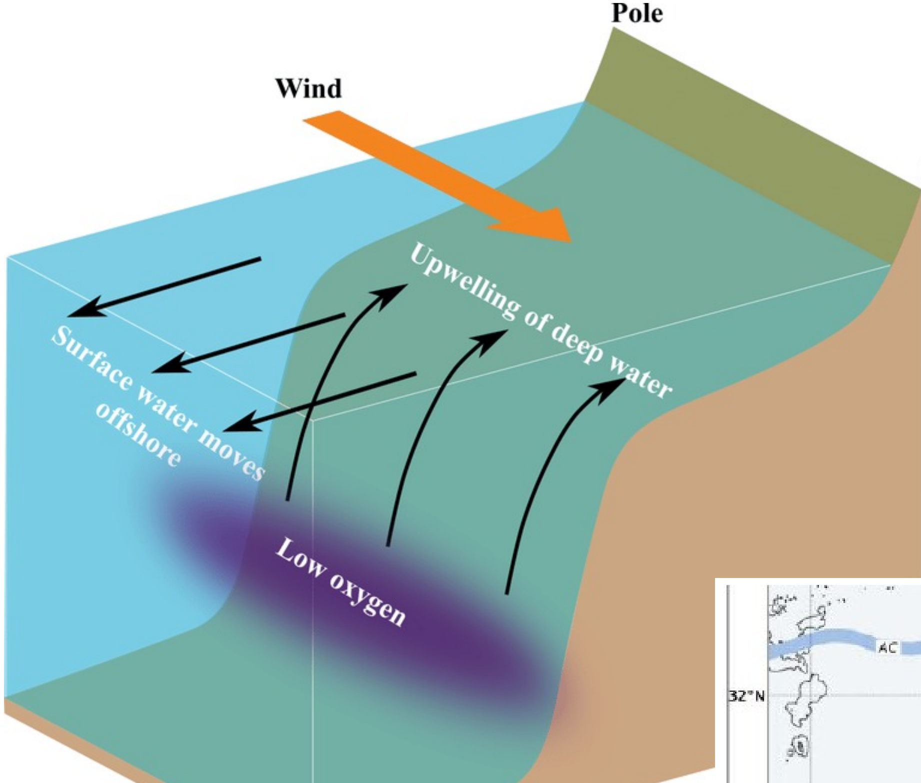


Upwelling of cold water favourable for fishing

Ekman transport at 45°

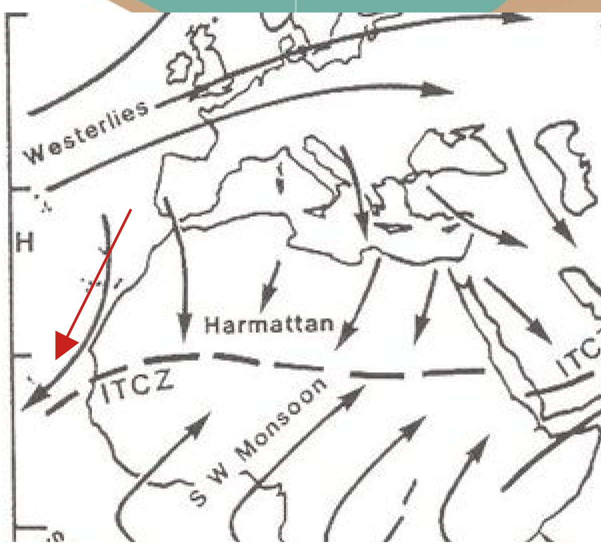


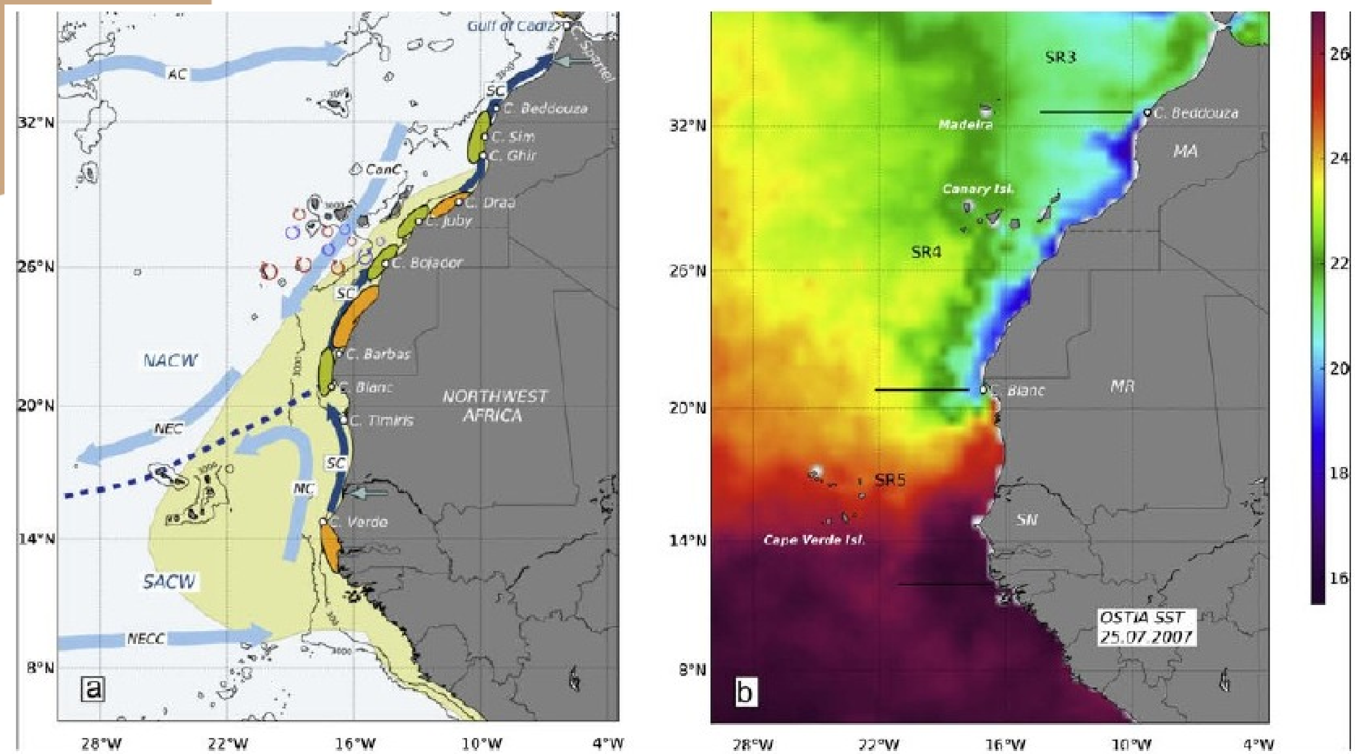
5. Local winds and circulations: Ekman Layer



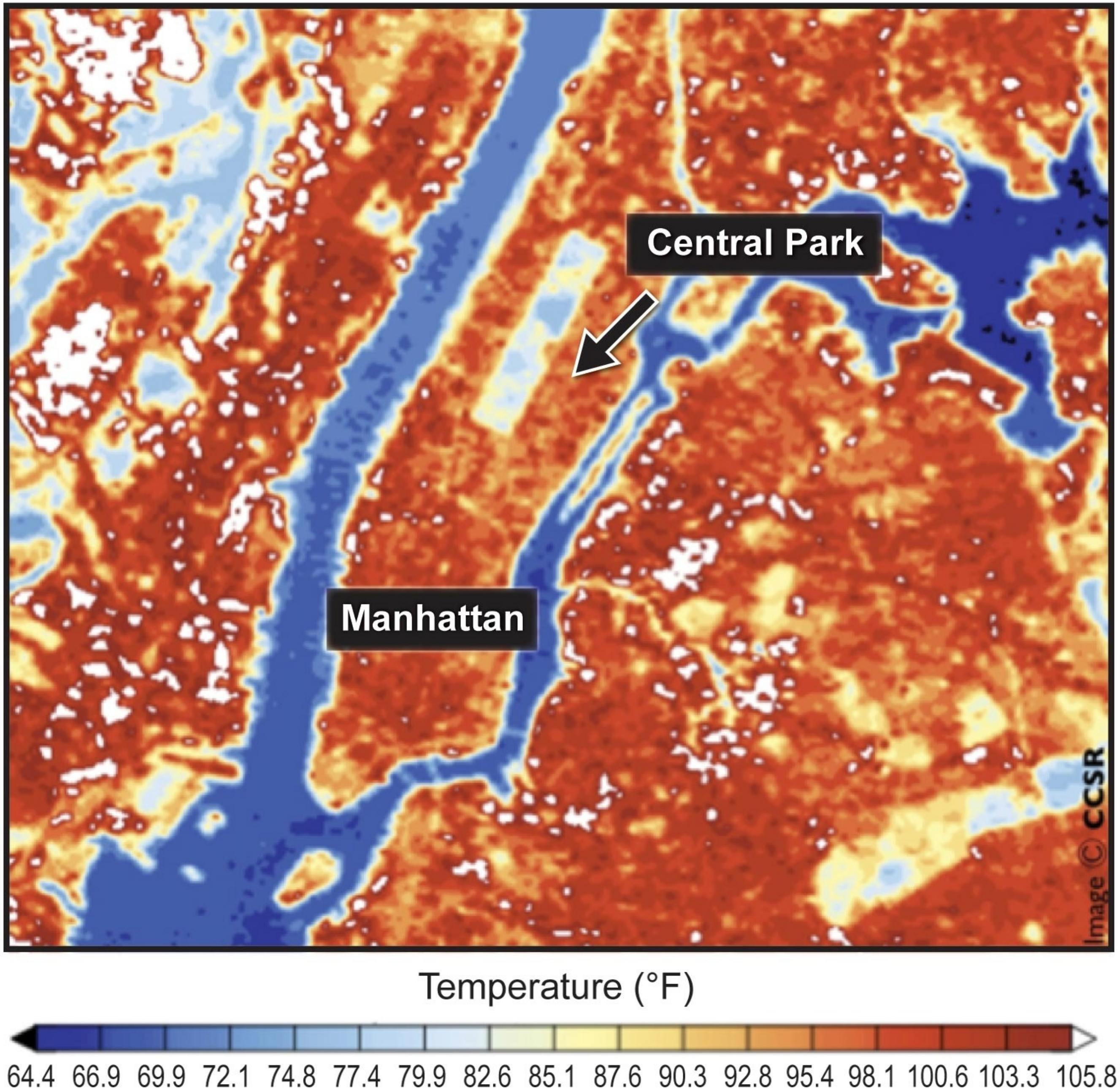
Occidental Sahara :

Azores High
=> upwelling of cold water
=> very stable condition in surface
=> no precipitation along the coast
=> but a lot of fog

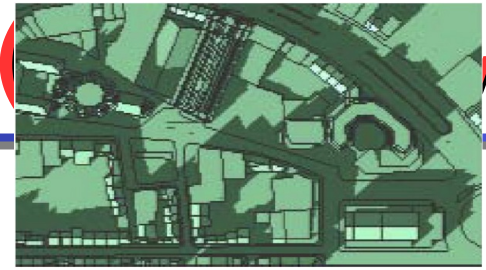




6. Urban Heat Island (UHI)

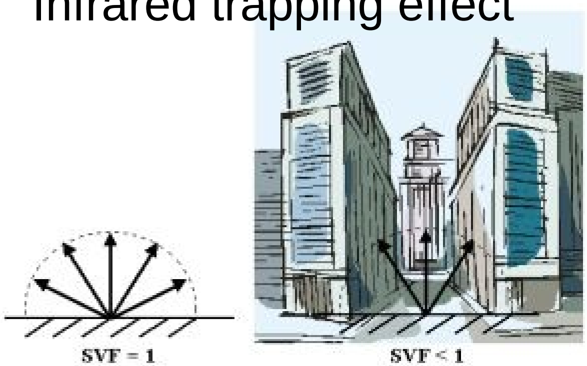


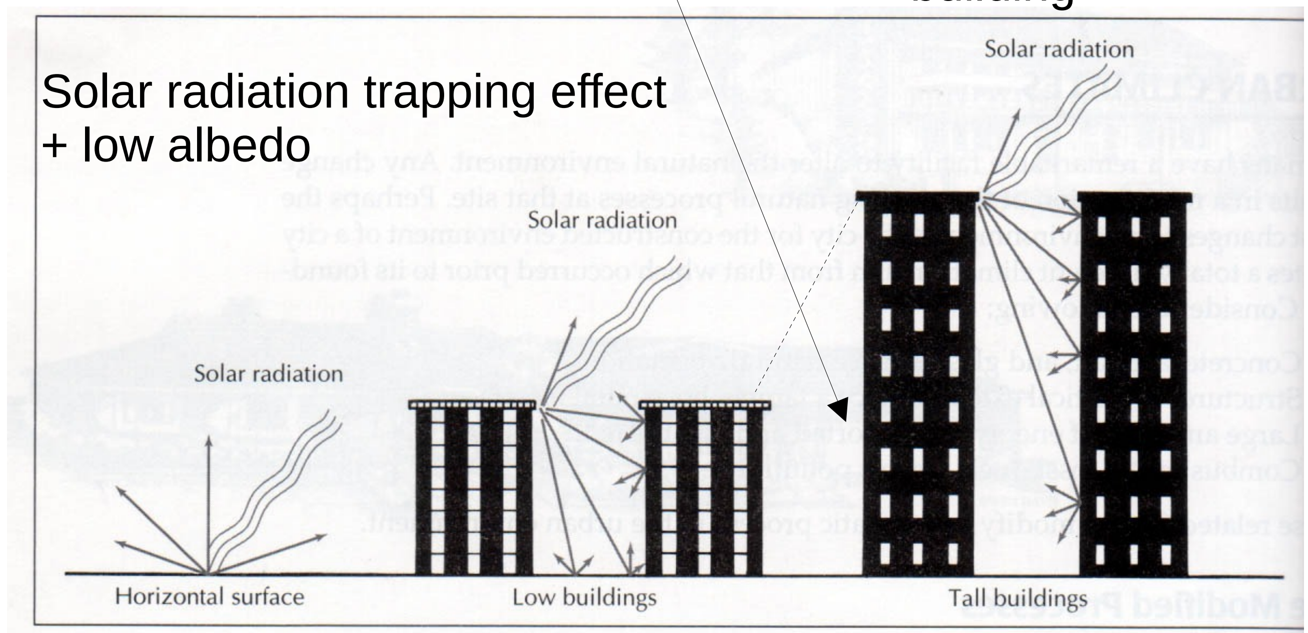
6. Urban Heat Island (… y fluxes



cast shadows behind the building

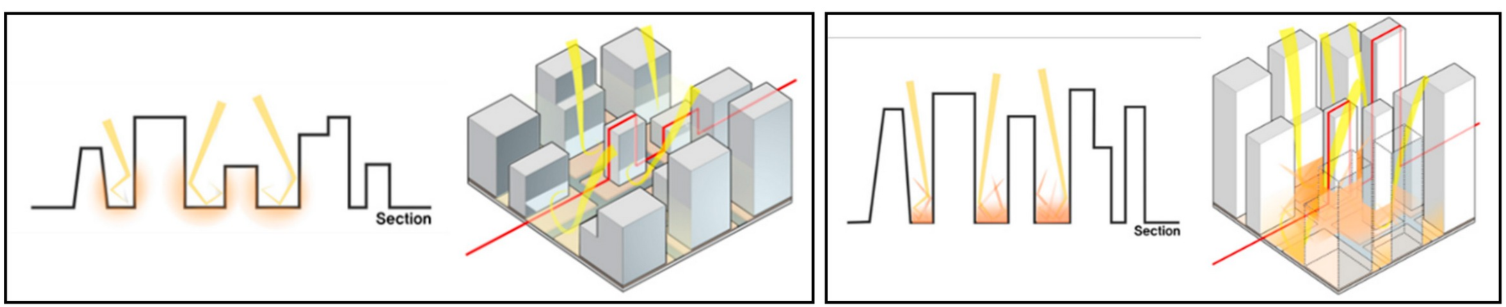
Infrared trapping effect





Longwave (LWD)

Shortwave (SWD)



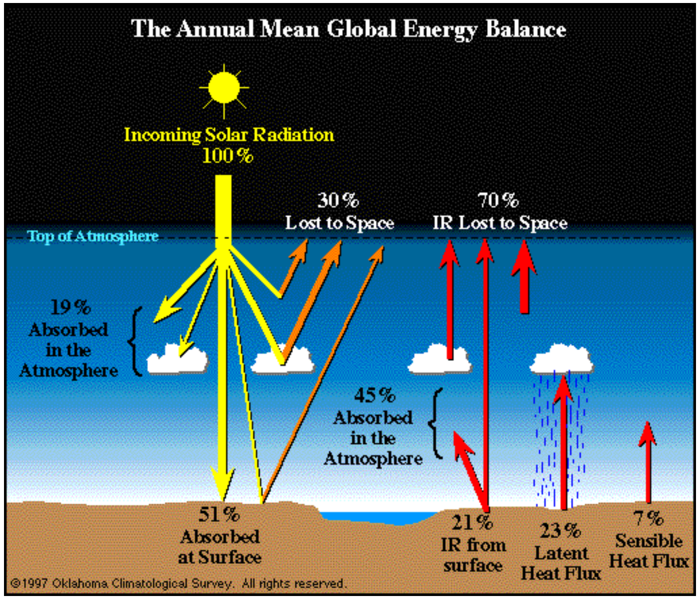
(a) (b)

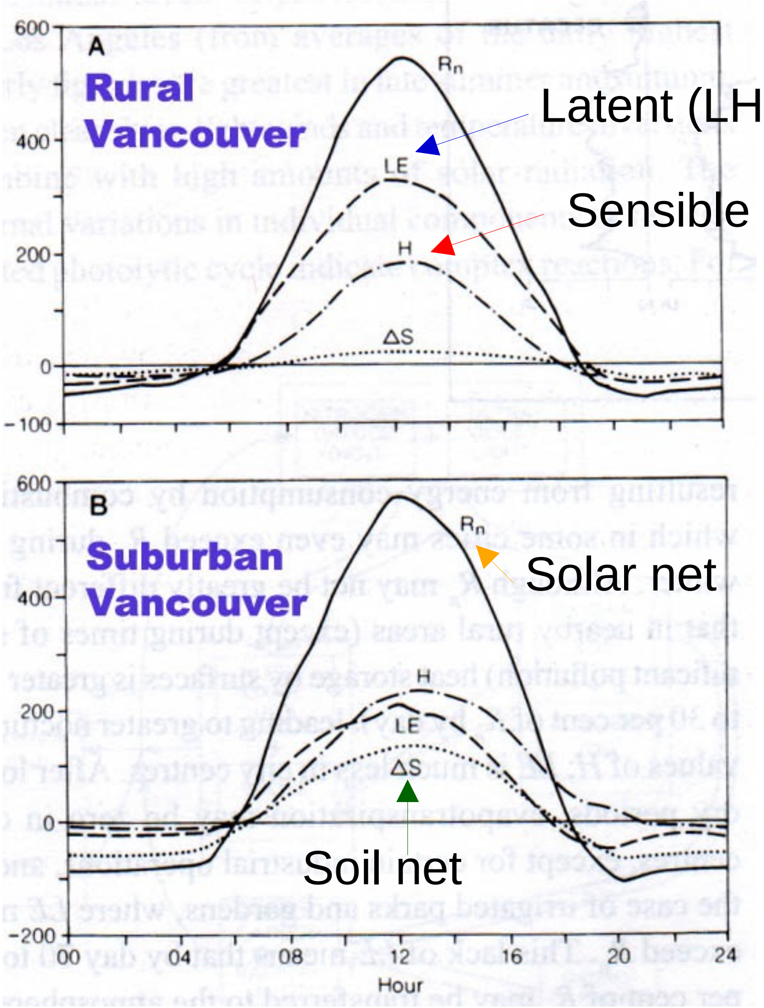
urban canyon albedo/effect but also cast shadows

6. Urban Heat Island (UHI): energy fluxes

- Latent heat flux (LHF) typically much smaller in a city

- Sensible heat flux (SHF) thus increases
  - Also increases due to increased turbulence in urban environment

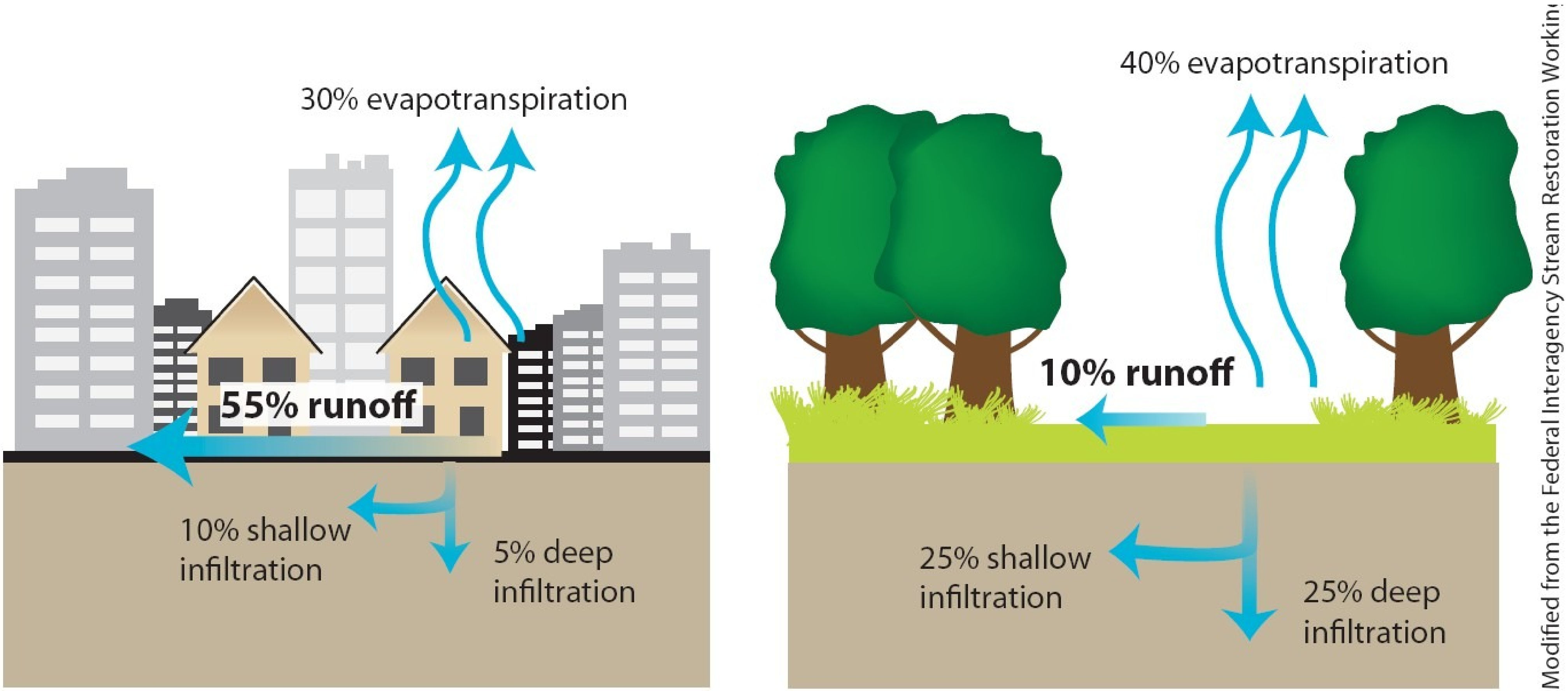




SHF ~ WS * Δ T

LHF ~ WS * Δ Q

6. Urban Heat Island (UHI): energy fluxes



Less water to cool the city during an heat wave ...

6. Urban Heat Island (UHI): energy fluxes

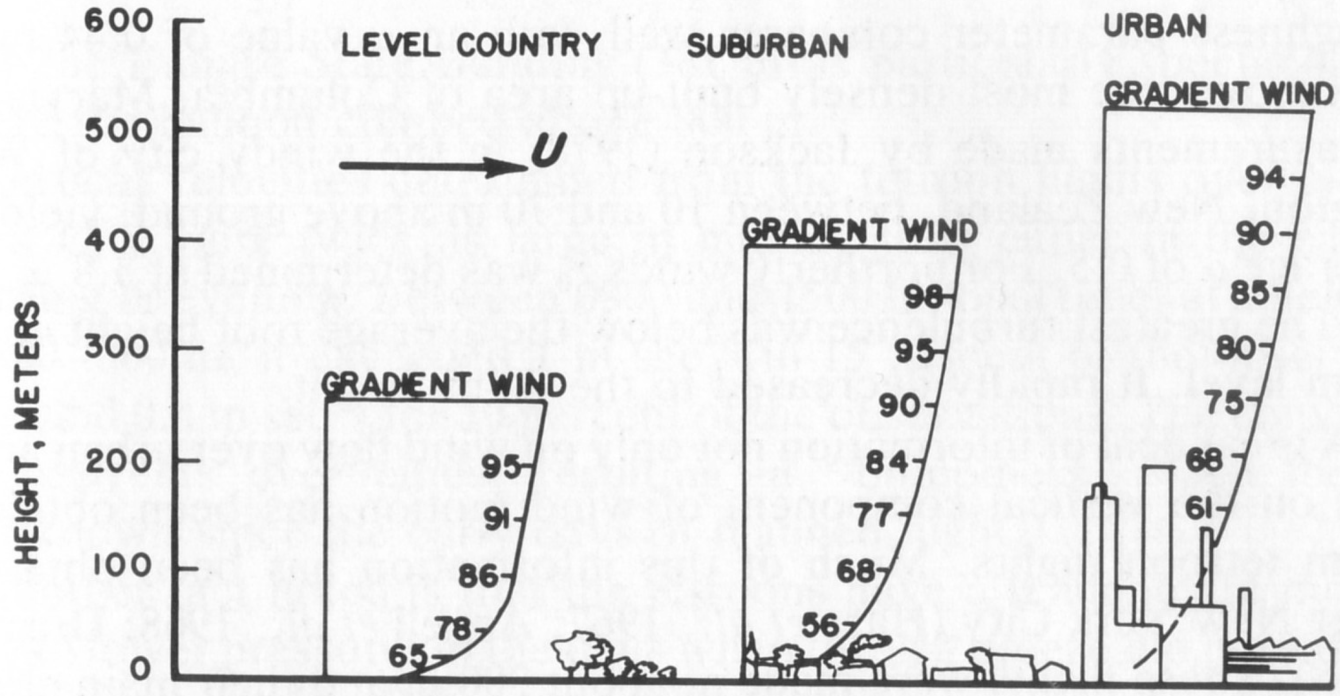


Fig. 6.7 Vertical wind speeds, in percent of the gradient wind at various heights over terrain of different roughness (after Davenport, 1965).

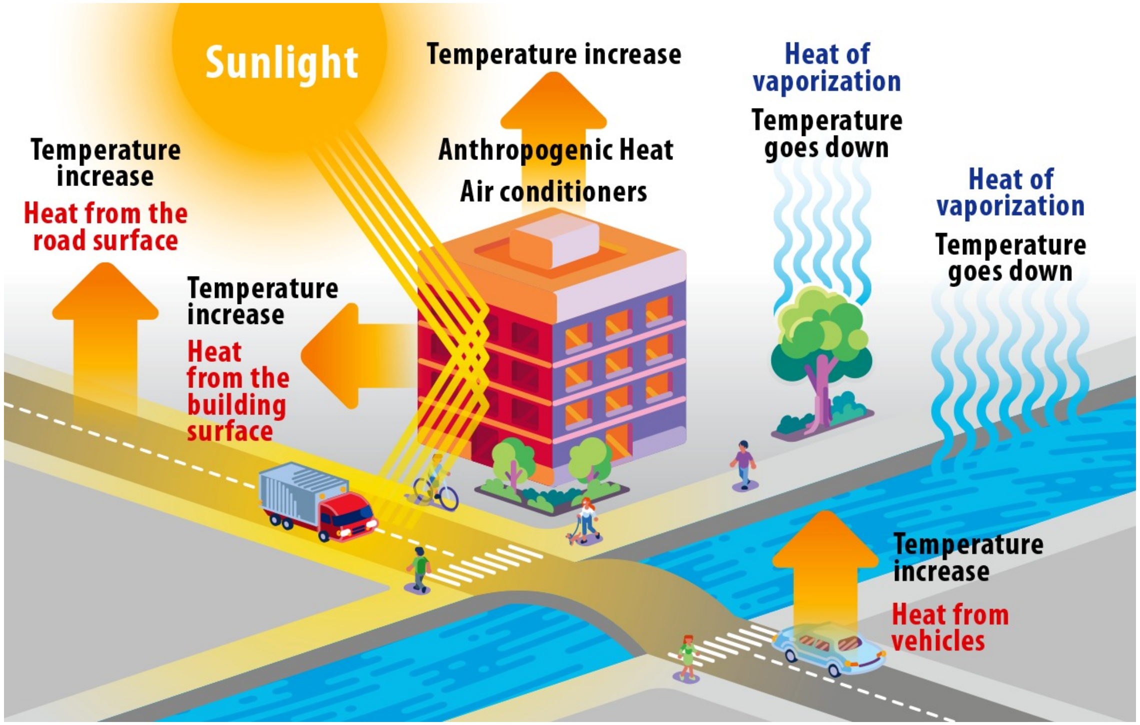
- Local circulation develops
- Wind decreases overall
- Wind tunnels + Rugosity>>

} Impacts on SHF and LHF

6. Urban Heat Island (UHI): energy fluxes

Anthropogenic heat (from combustion processes)

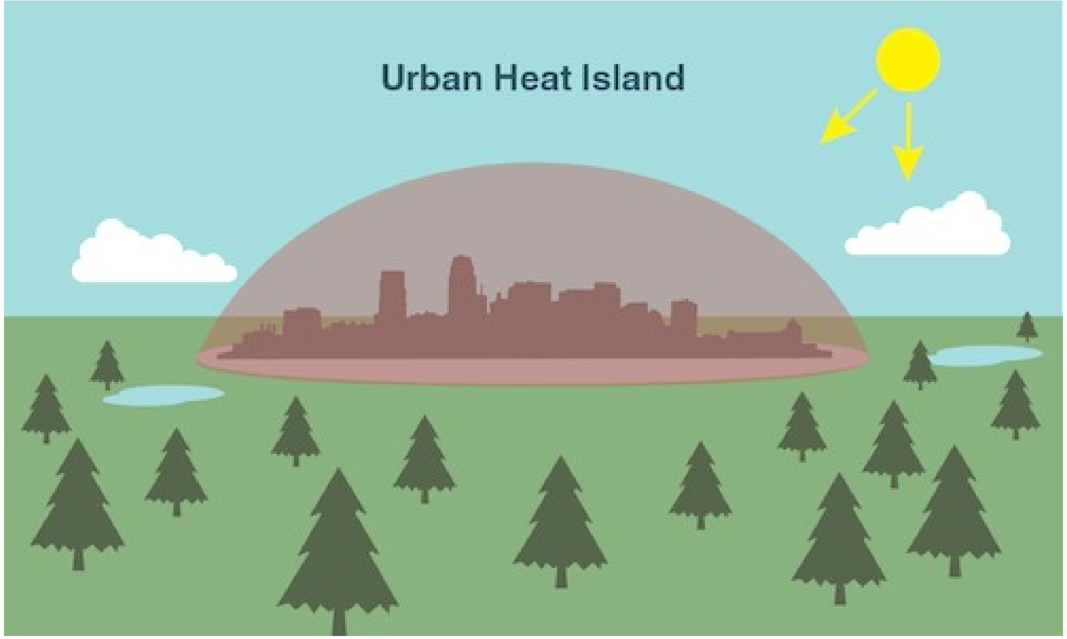
- In wintertime: thermal rejection from heaters
  - 2 x energy received from sun
- In summertime: thermal rejection from coolers
  - 20% energy received from sun
- All year, thermal rejections of
  - Vehicle
  - Industry
  - Commerce
  - …



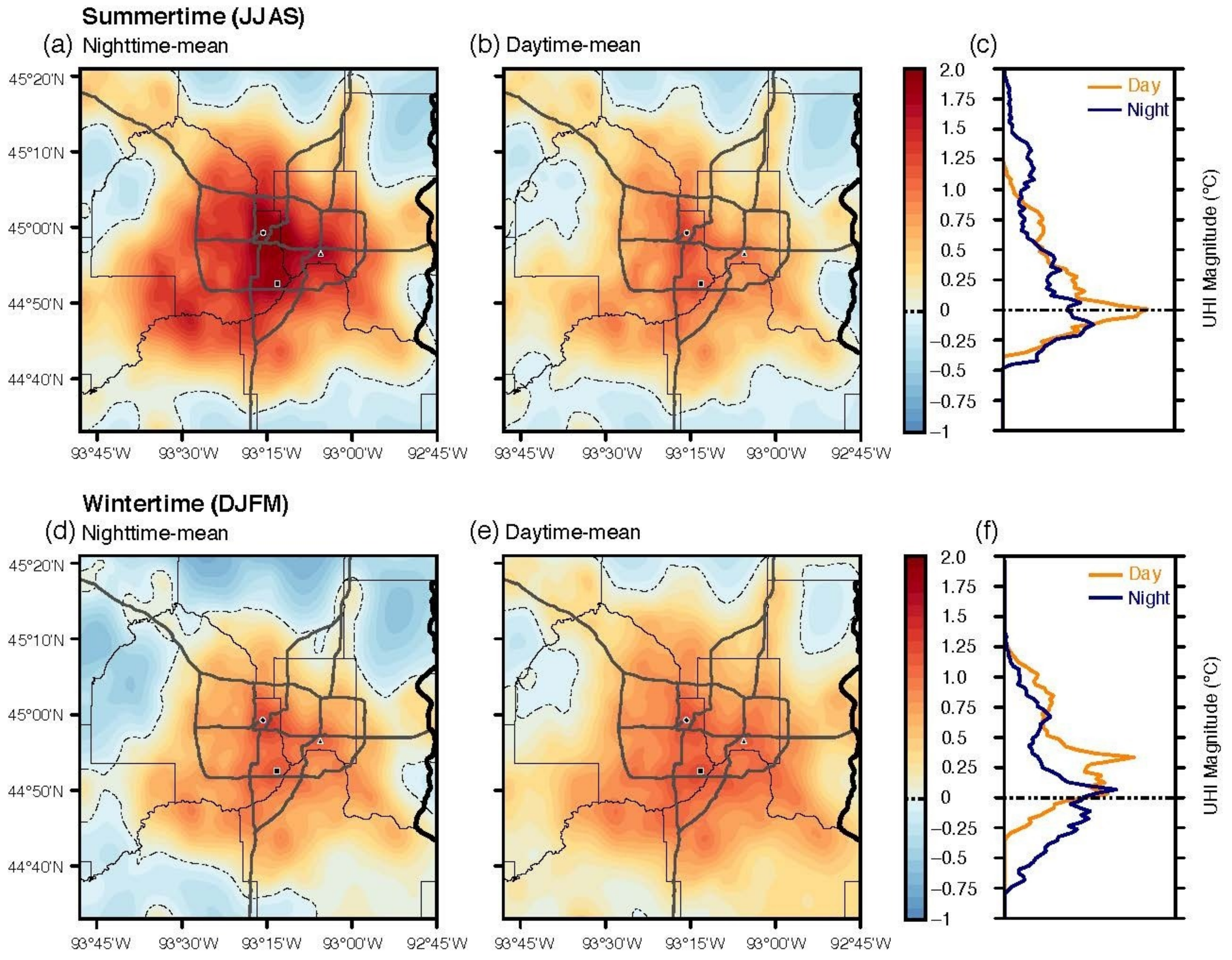
6. Urban Heat Island (UHI): energy fluxes

- **Albedo** (–; net solar radition ++)
- **Conductivity** of materials (G ++)
- **Impermeability** of materials (LHF --)
- **Sensible heat flux** (SHF ++)
- **Rugosity** (++; turbulence/mixing ++)
- **Anthropogenic heat flux** (AHF ++)

Urban Heat Island (UHI)



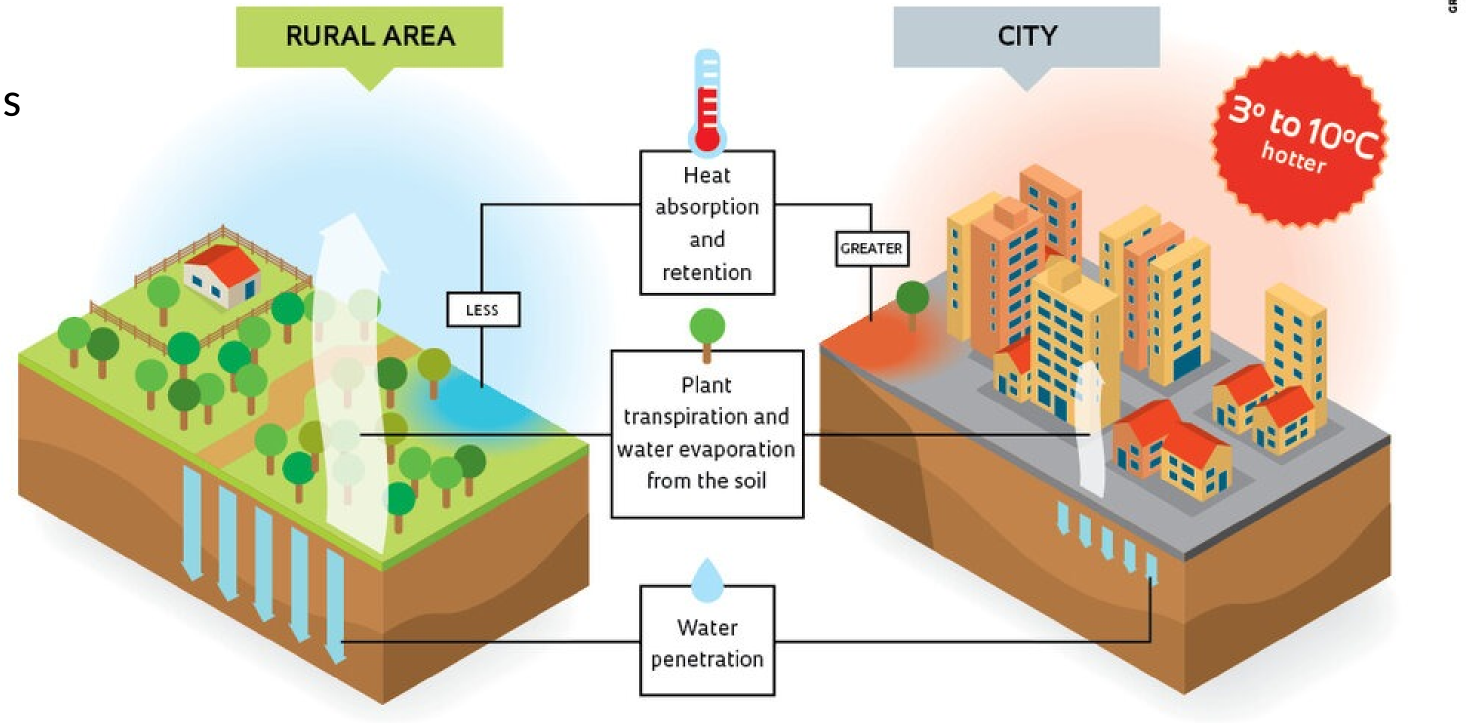
6. Urban Heat Island (UHI): temperature



6. Urban Heat Island (UHI): temperature

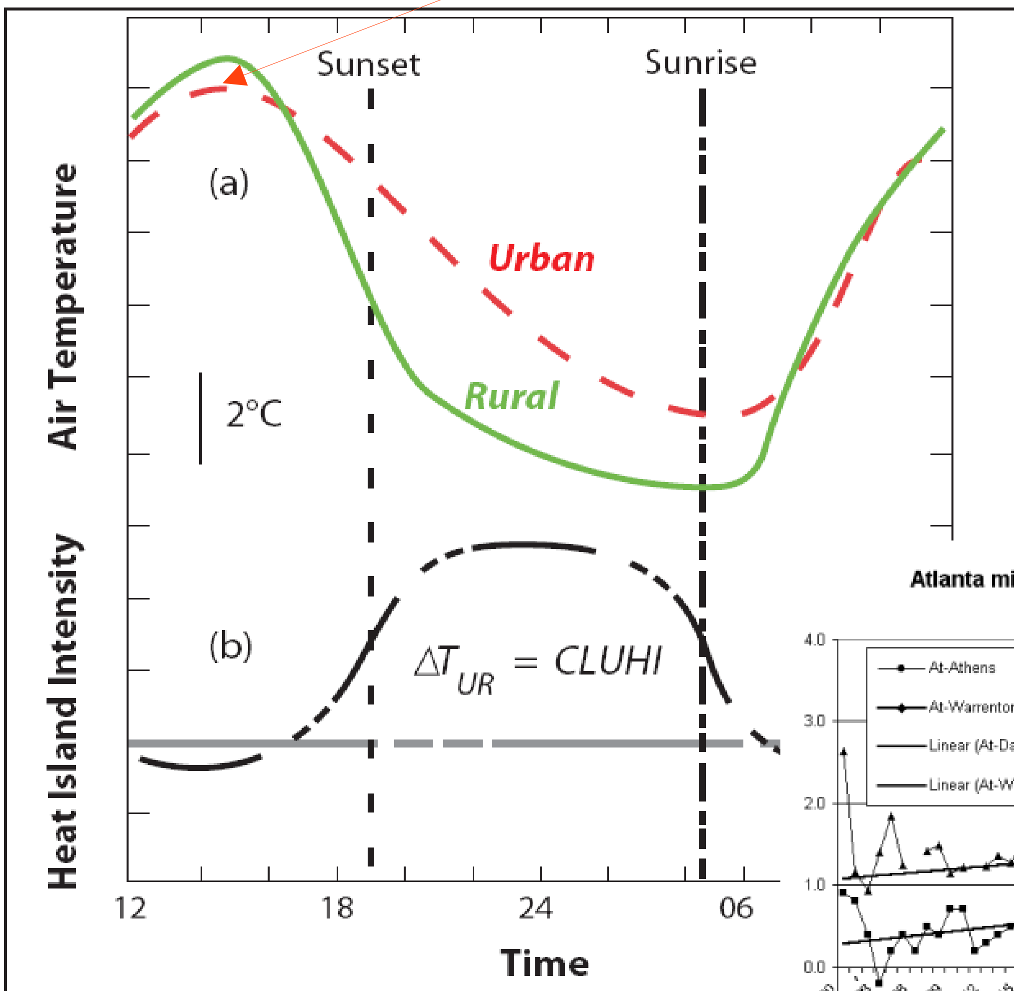
- Annual scale :
  - Temperature increase of ~+1°C warmer (depend on the city)
- During heatwave events :
  - Temperature increase of more than 10°C

- Most significant heat island
  - At night
  - Clear sky conditions
  - Calm wind
  - Winter
  - During the week

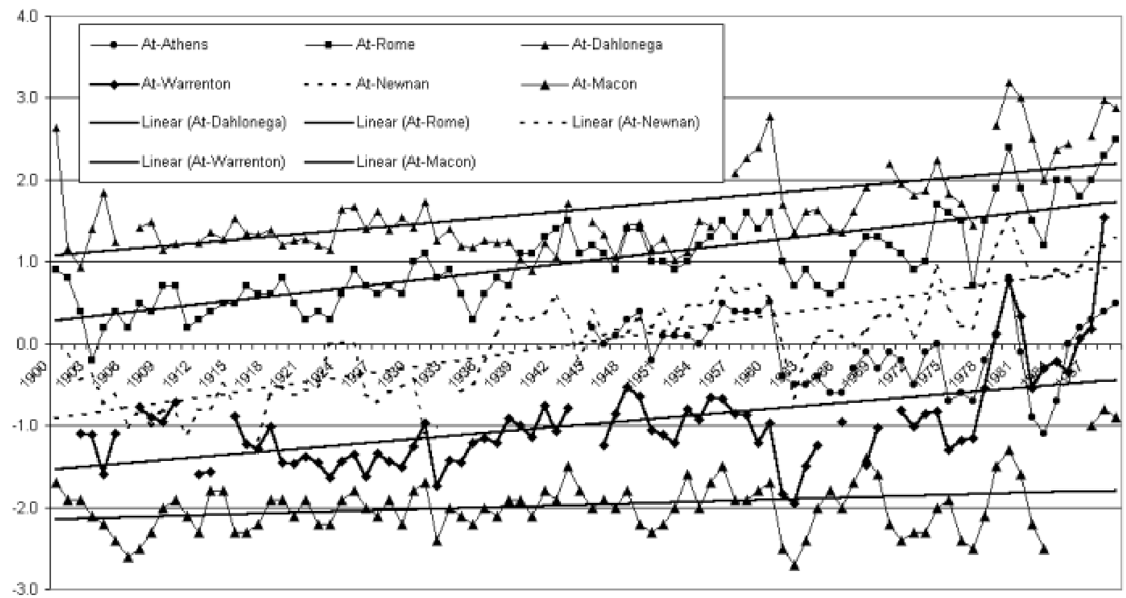


6. Urban Heat Island (UHI): temperature

Decrease of solar radiation due to pollution

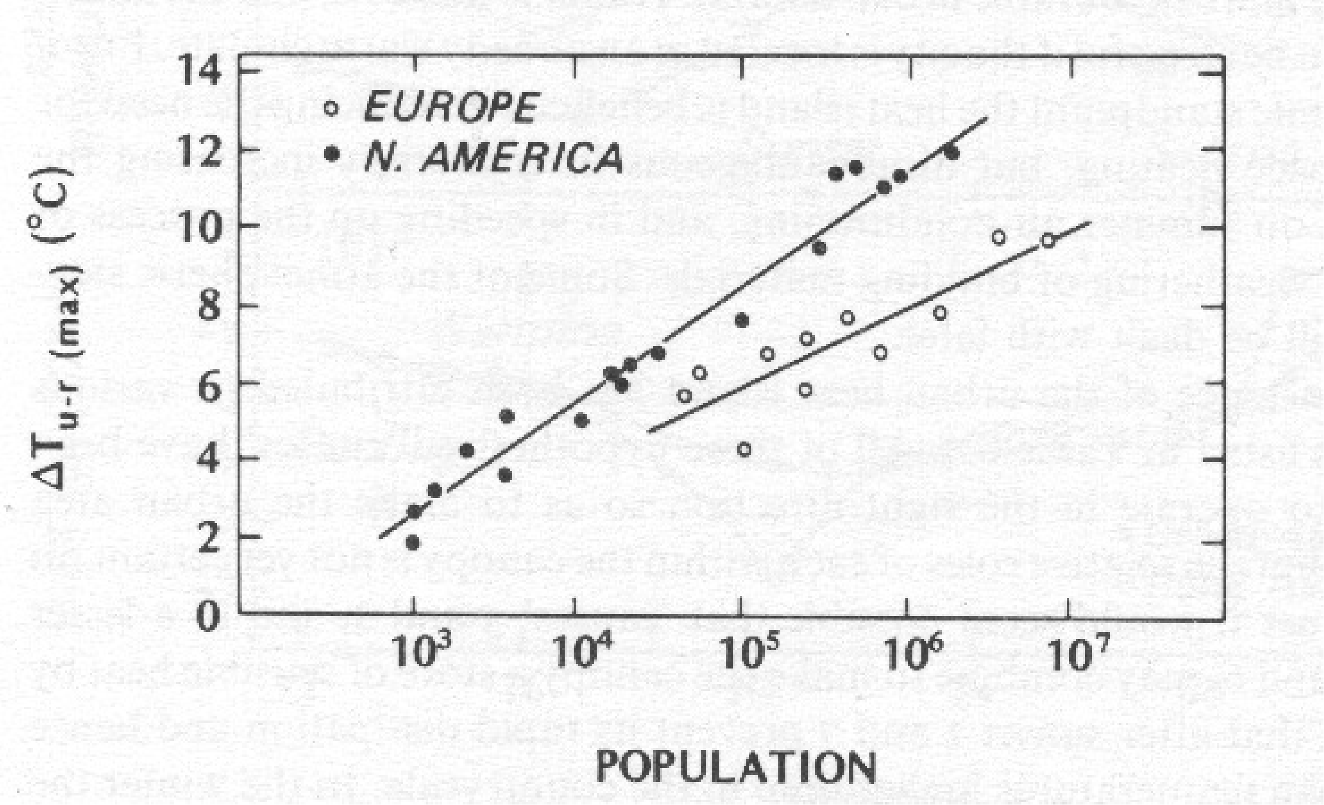


Atlanta minus six neighbours, demonstrates heat island contamination in Atlanta, Macon and Athens.



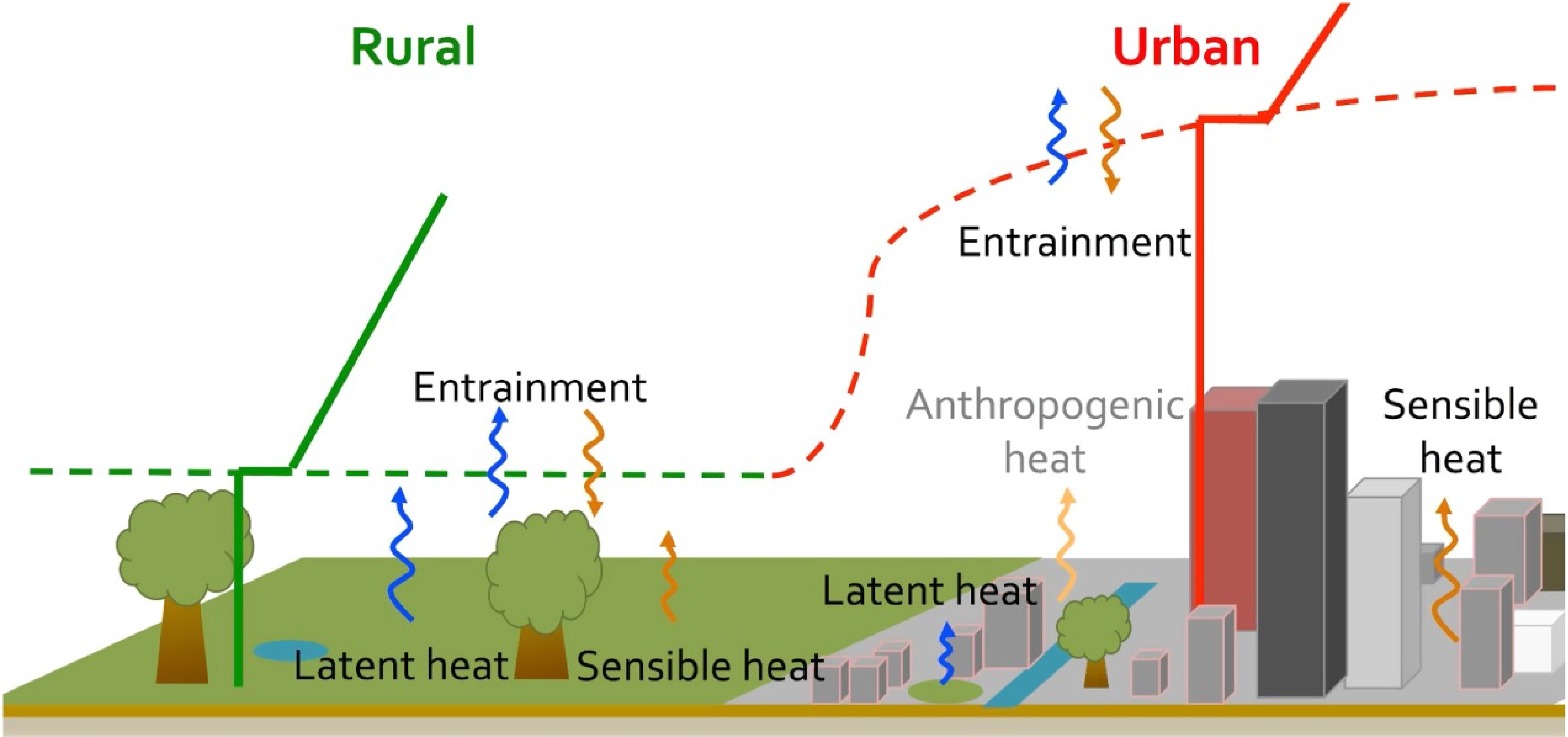
6. Urban Heat Island (UHI): temperature

Heat islands noted in cities of only 1,000 people



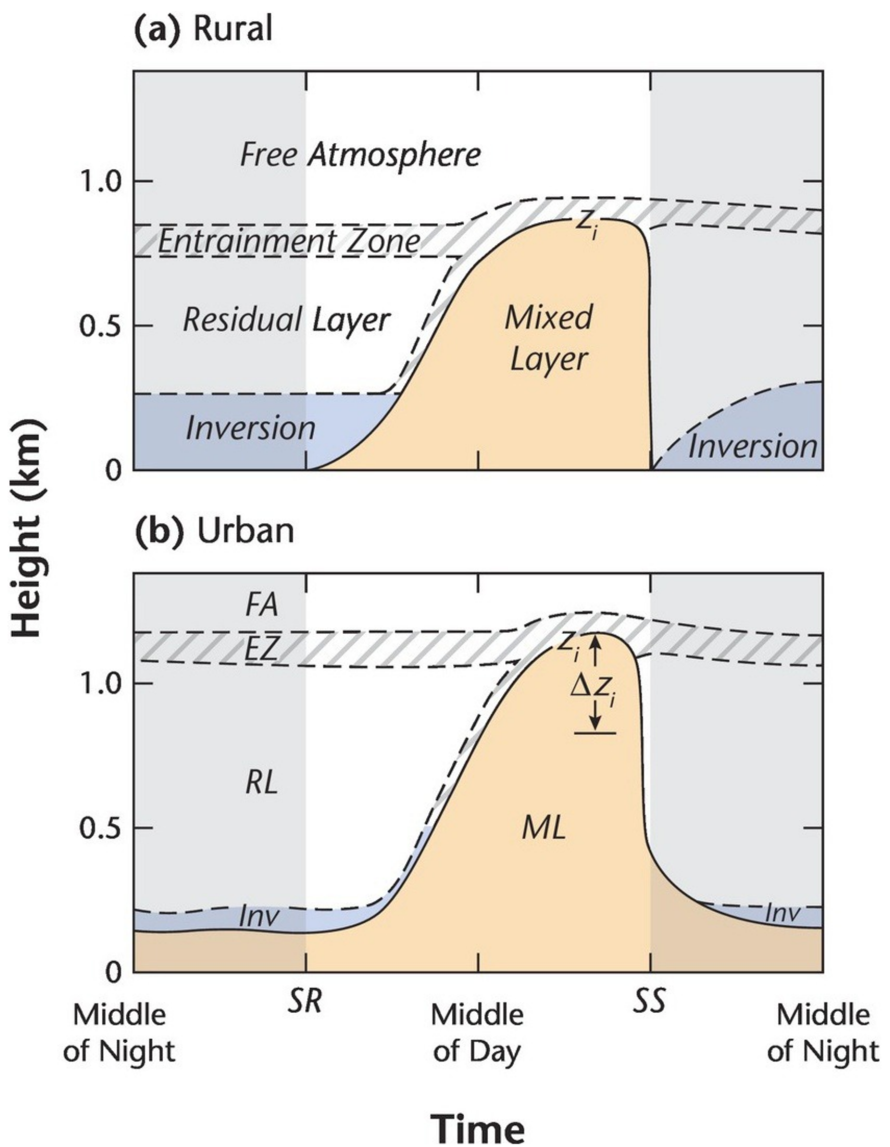
Europe: warmer climate in winter and more isolated houses than USA

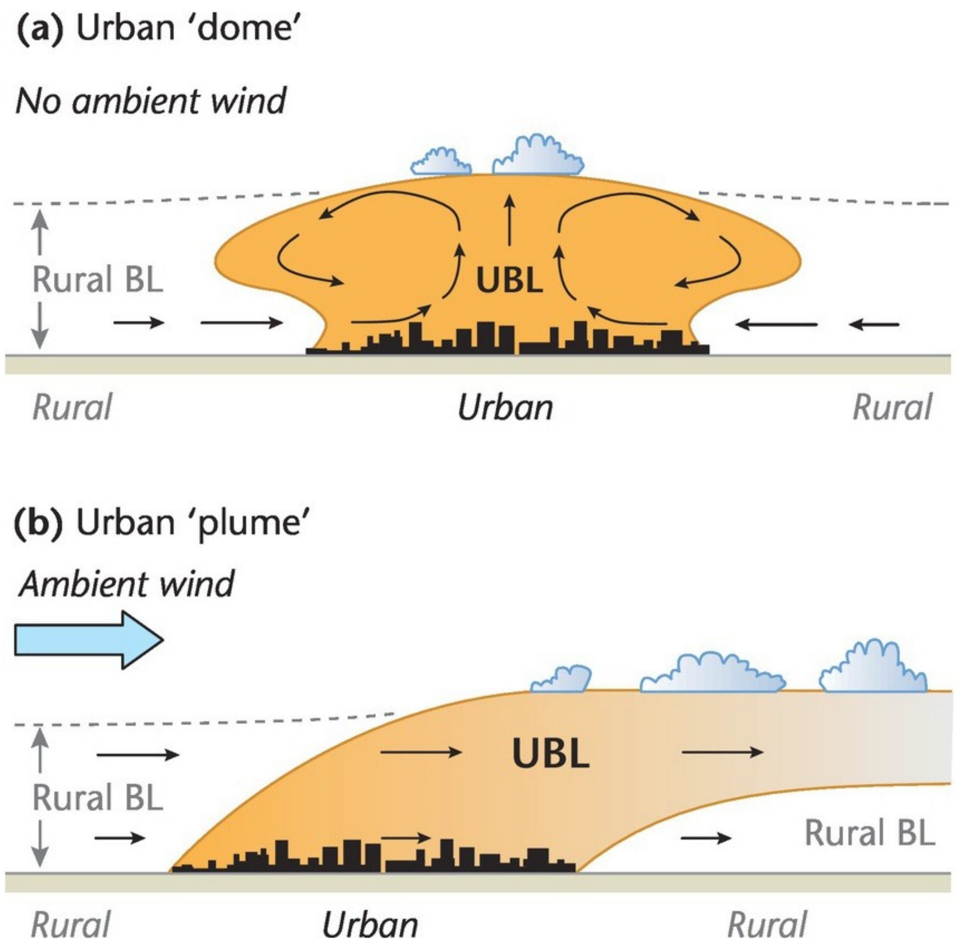
6. Urban Heat Island (UHI): temperature



Atmospheric boundary layer (ABL/PBL): thickness of atmosphere of the same behaviours (due to mixing/turbulence) which are in contact with surface

6. Urban Heat Island (UHI): temperature





Urban modification of moisture

- Relative humidity 6% lower
  - Larger decreases in summer
- Cloud cover increases 10%
  - Condensation nuclei
- Fog/smog more frequent

Urban modification of precipitation

- Up to 10% higher in city
- Up to 33% higher within 50 km of city DOWNWIND

Peak of precipitation after the city

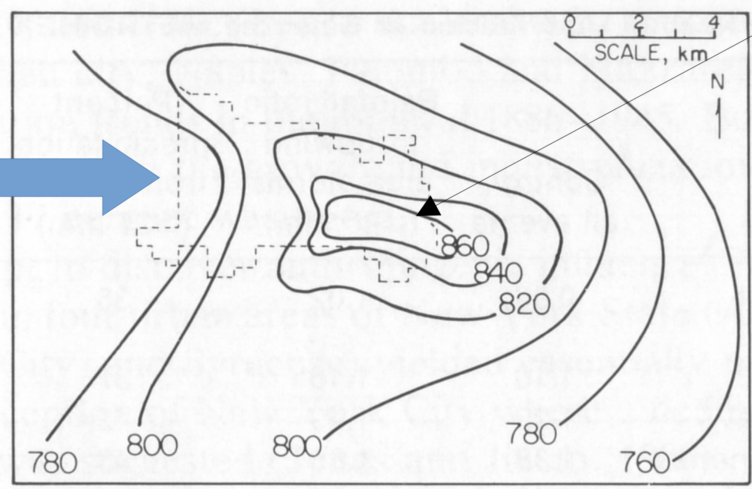


Fig. 8.4 Isohyets of annual precipitation (mm) at Urbana, Illinois (a
Changnon, 1962). Dashed line marks urban limits.

Weekly signal

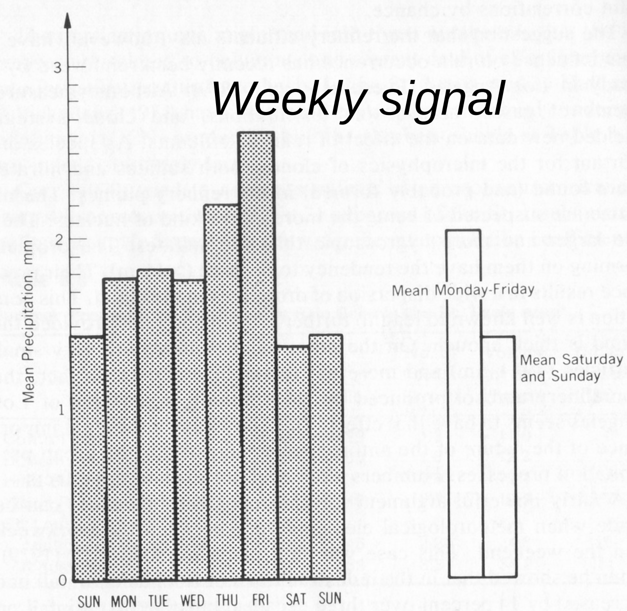
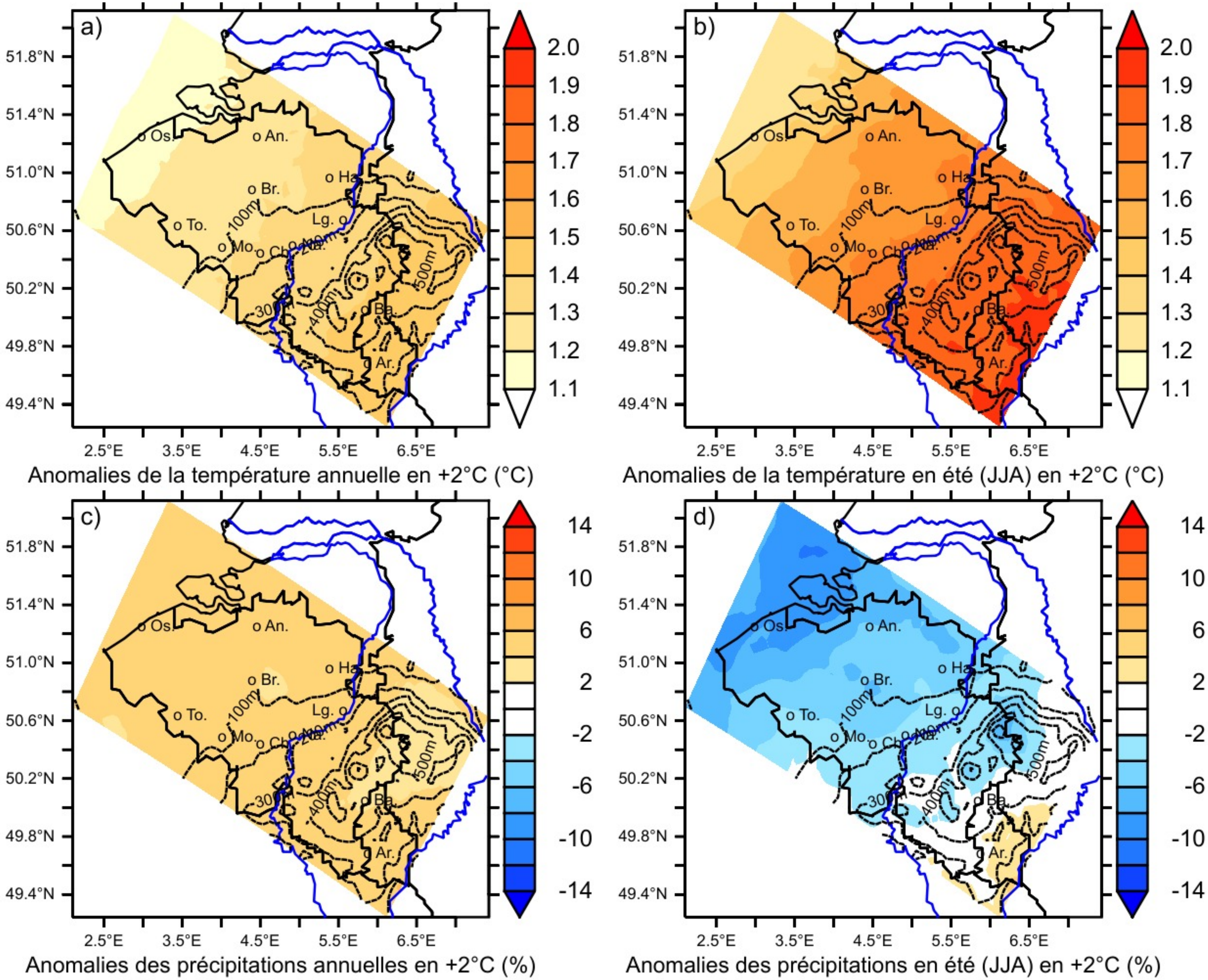
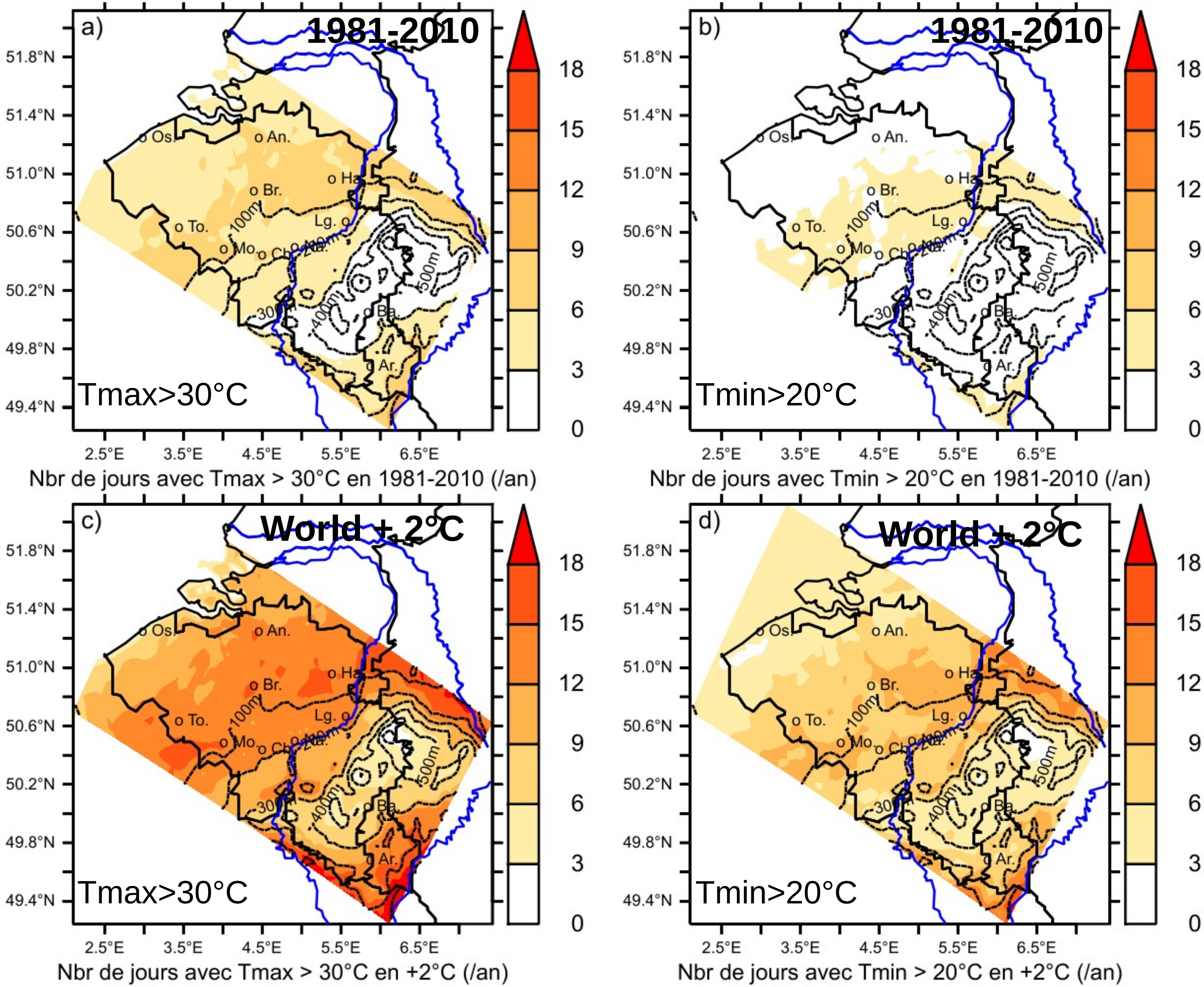


Fig. 8.6 Precipitation in Paris, France, averaged by day of week.

6. Urban Heat Island (UHI): future projections

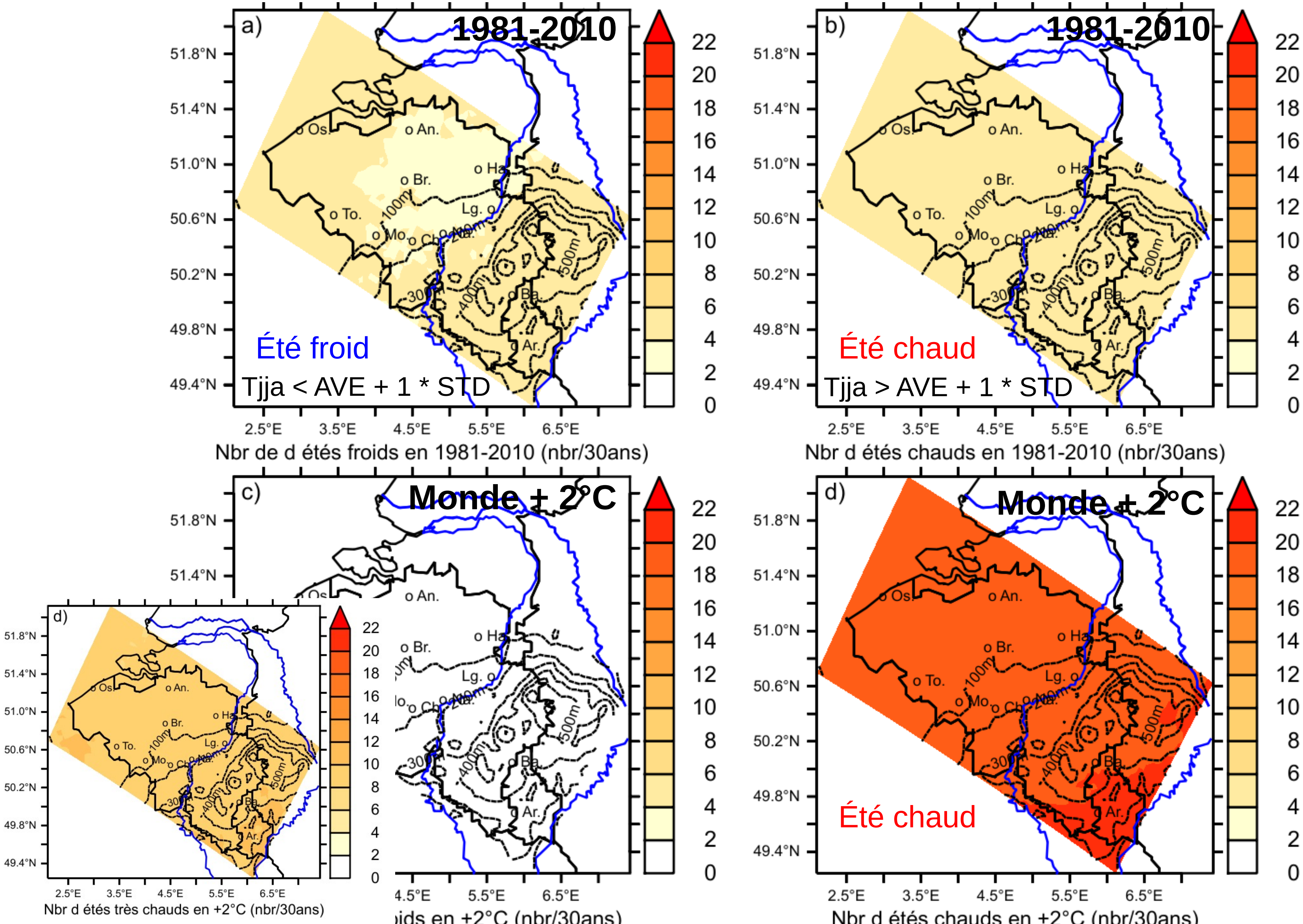


6. Urban Heat Island (UHI): future projections



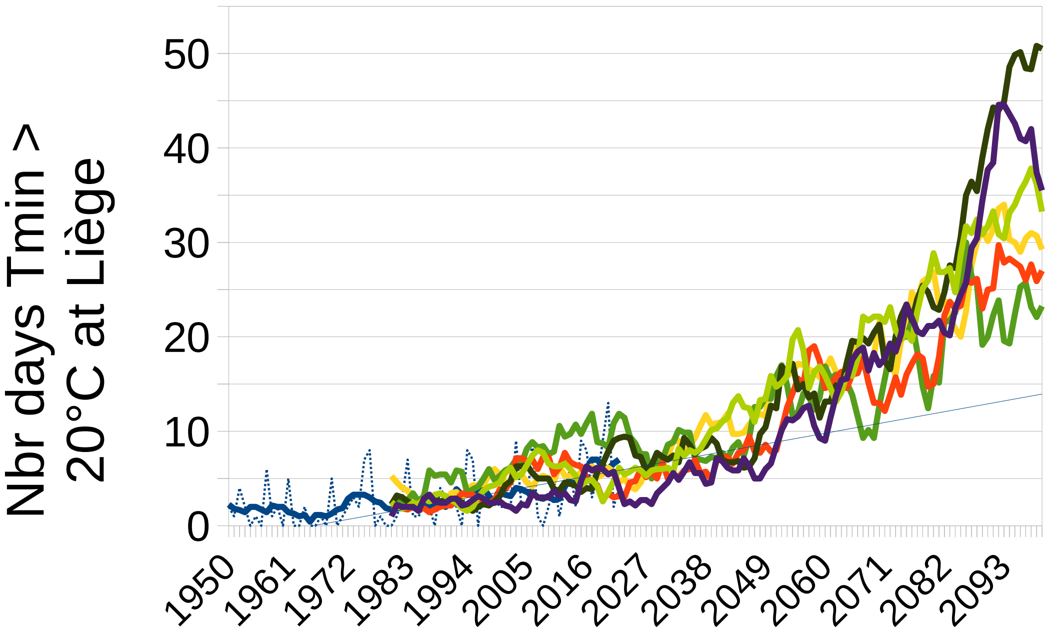
+300 %

6. Urban Heat Island (UHI): future projections

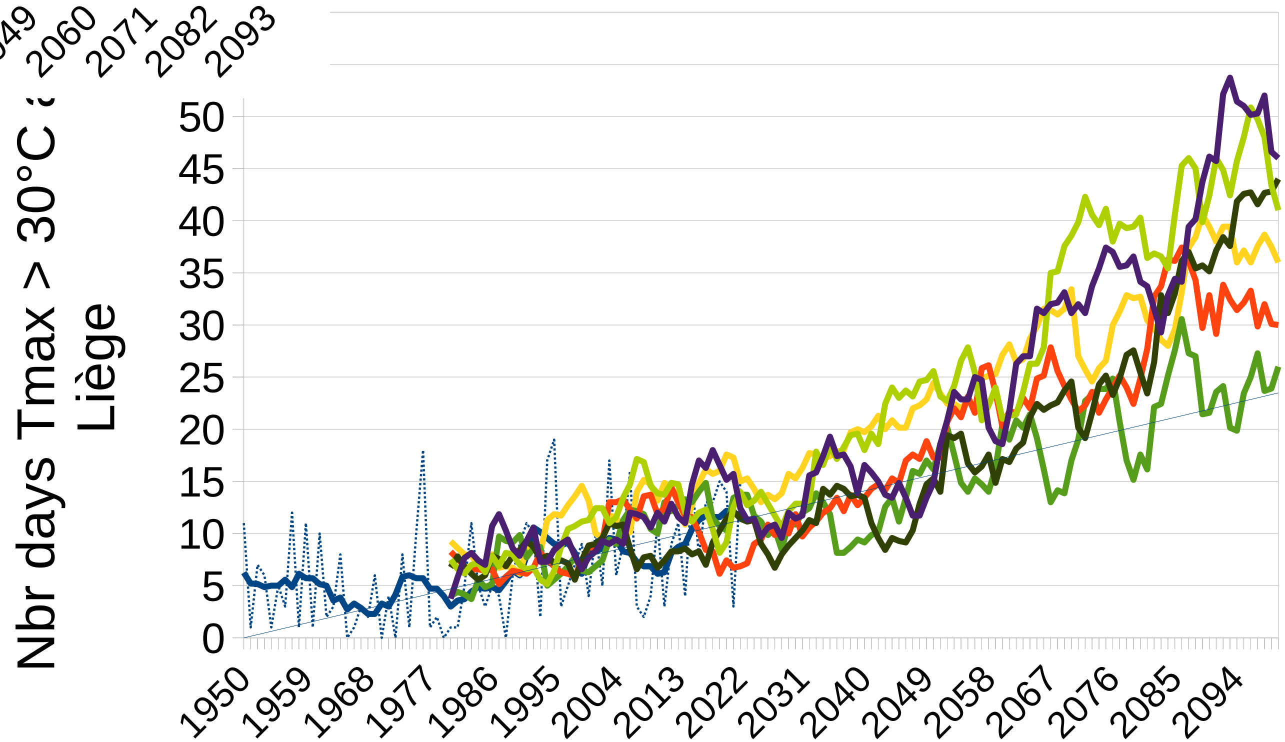


Étés caniculaires Tjja > AVE + 2 * STD

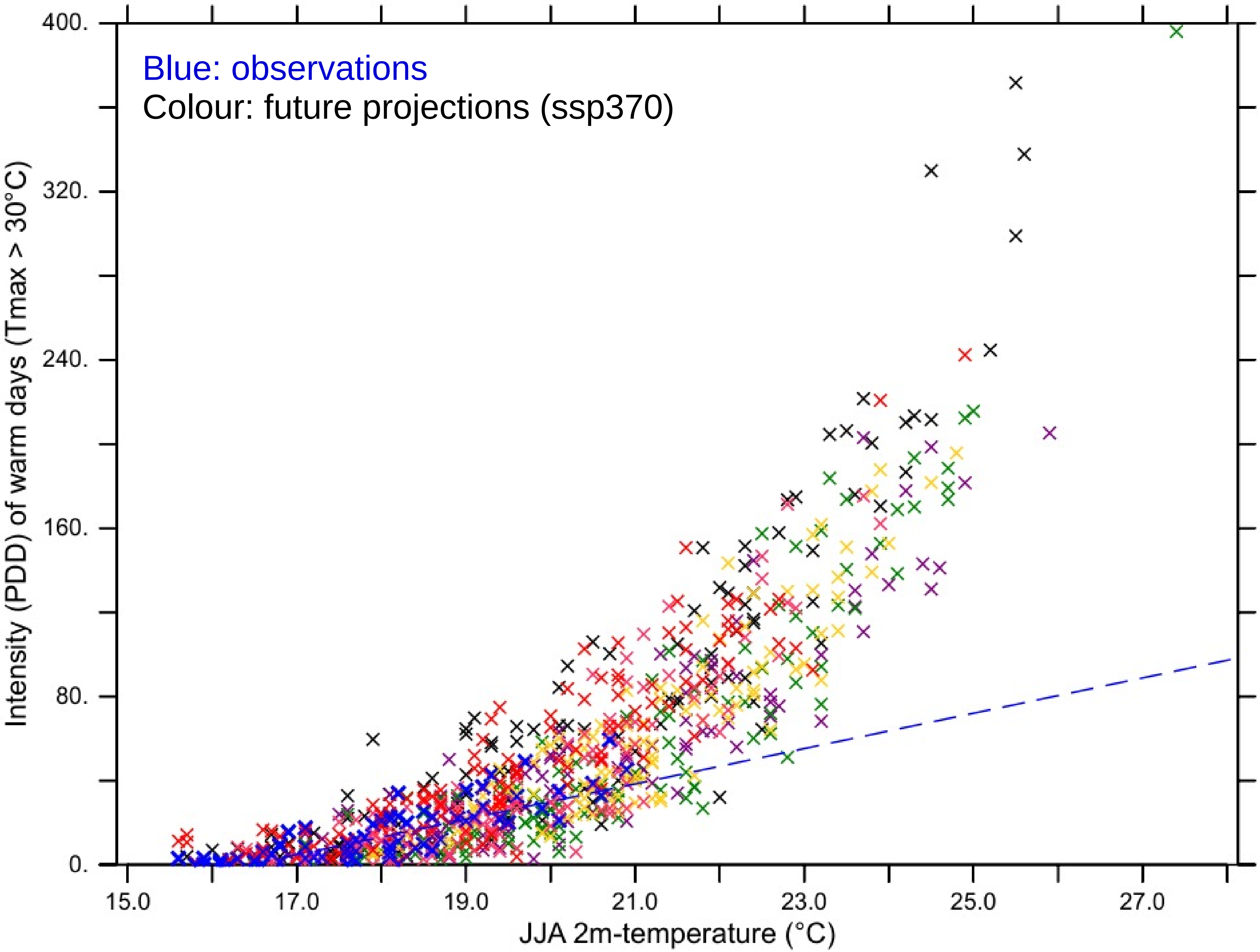
6. Urban Heat Island (UHI): future projections



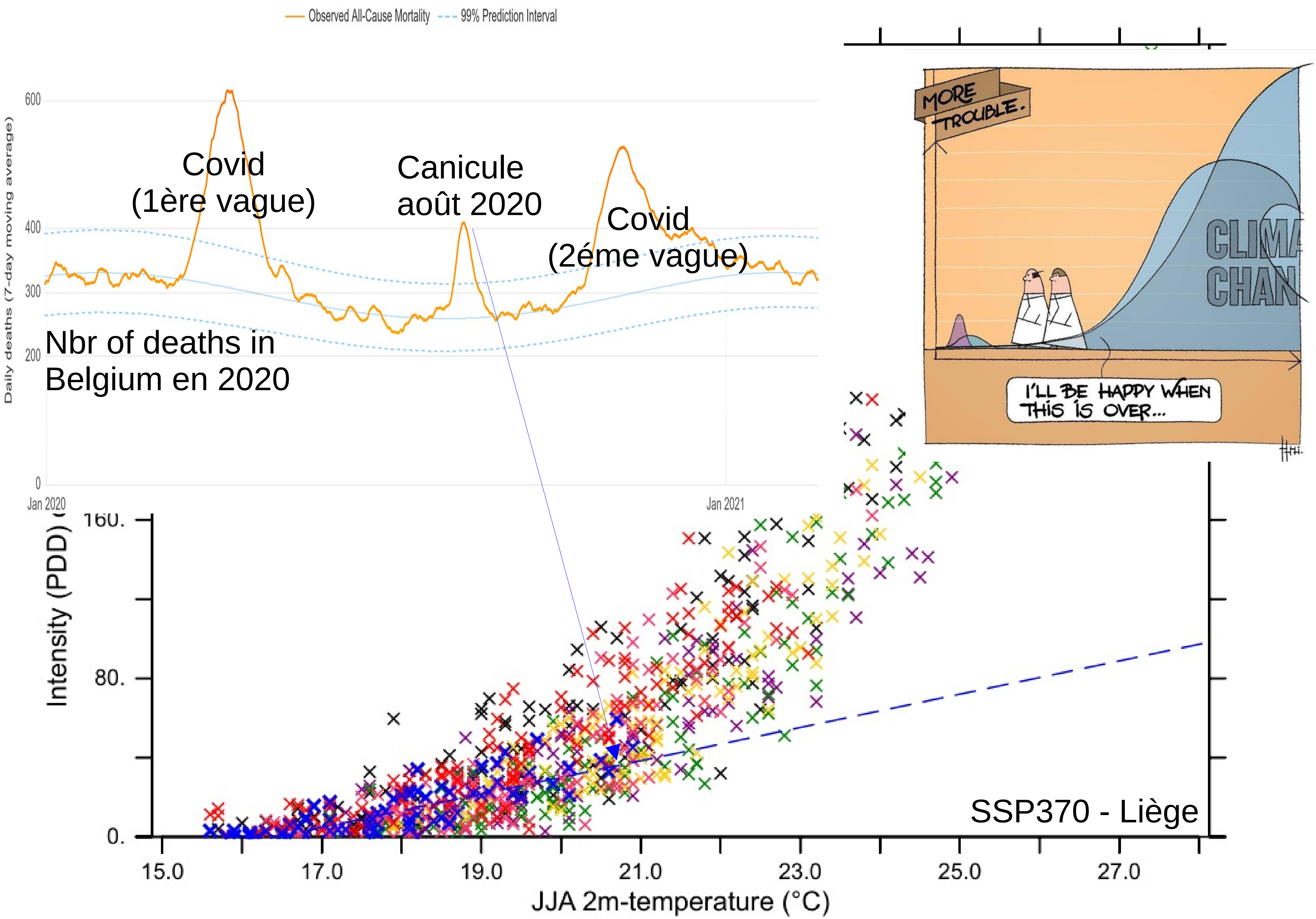
Significant increase of anticyclonic conditions in summer.



6. Urban Heat Island (UHI): future projections



6. Urban Heat Island (UHI): future projections



6. Urban Heat Island (UHI): mitigation

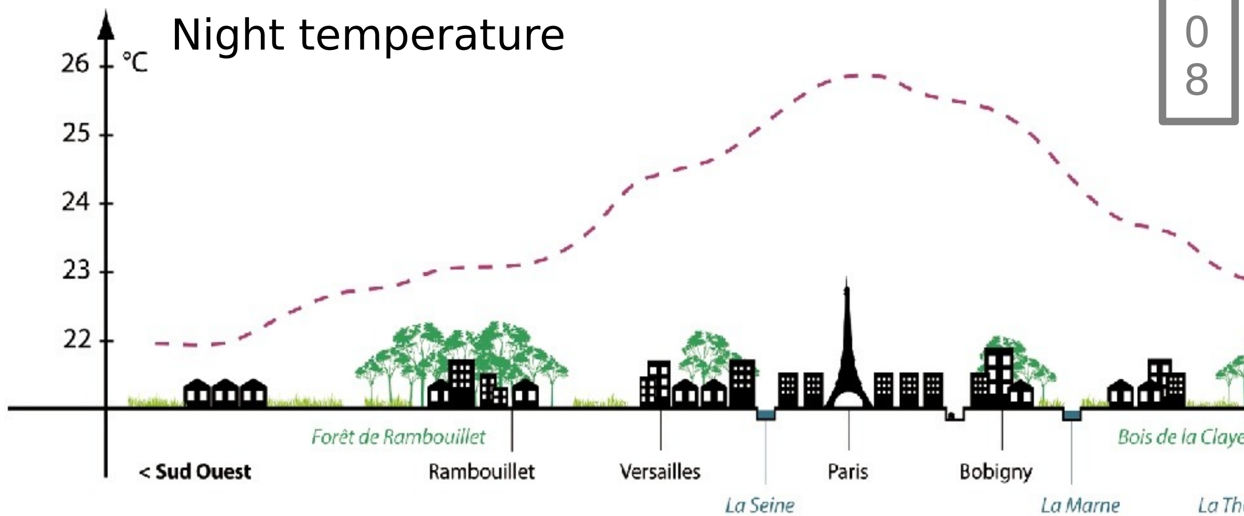
- Increase surface albedo of cities (solar radiation --)
  - Highly reflective roofs
  - Highly reflective roads
- Green and Blue infrastructures (latent heat flux ++)
  - Vegetated roofs
  - Vegetated facade
  - Add some vegetated parks in cities
  - Add some water bodies, ponds, lake ...
- Make the soils permeable (flood –- and LHF ++)
  - Green parking lots

PLAN CANOPÉE
Liège

6. Urban Heat Island (UHI): mitigation

The great deal of the Parisian agglomeration
Le Grand « Pari » de l'agglomération parisienne

Night temperature



2008

FORETS (A. Peter)
+
MATERIALS MORE REFLECTIVE (Transsolar) + Lakes

Decrease estimated between 2 and 3°C

2030