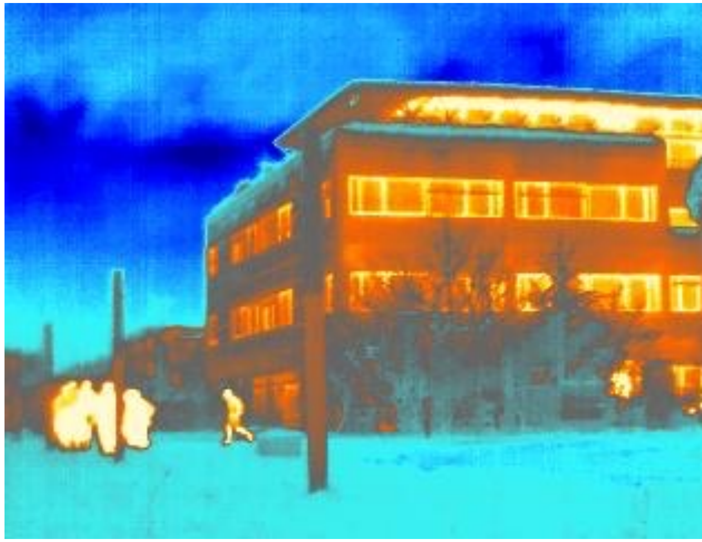
Photo: A. Black, UBC

07 Long-wave radiation.

Learning objectives

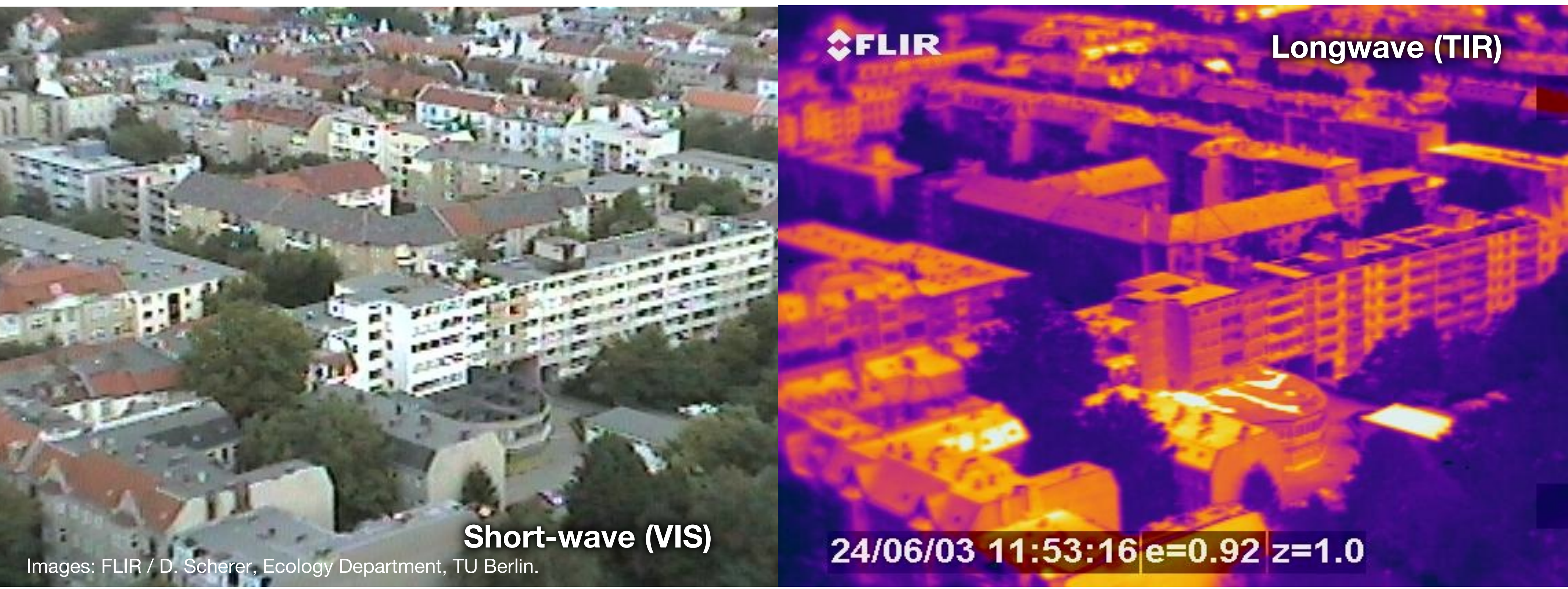
- Explain how radiation laws apply to long-wave radiative exchange as well.
- Describe how long-wave radiation can interact with surfaces.
- Explain how we can calculate longwave outgoing radiation, and how it relates to surface emissivity.
- Know how we can estimate / model longwave incoming radiation and the emissivity of the atmosphere.



People in front of a heated building in winter as seen in the thermal infrared by a thermal camera (Source: A. Christen)

What is ‘Longwave’ radiation?

- Wavelength range: 3 μm to 100 μm
- Longwave = far-infrared = thermal infrared radiation (TIR)



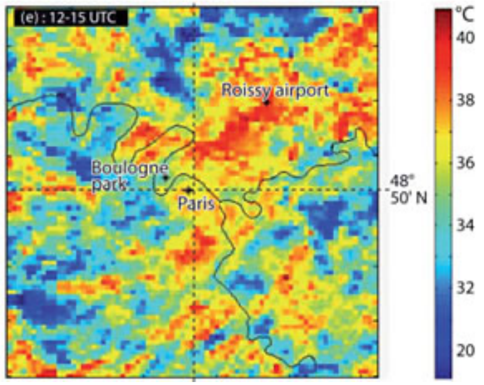
Images: FLIR / D. Scherer, Ecology Department, TU Berlin.

Measuring long-wave radiation

There is a range of instruments available that measure long-wave radiation received in a particular **field of view** (FOV), within a particular **band**, and/or with a particular spatial **resolution**.

Pyrgeometer

Thermal camera



Thermal satellite channel

Review: Stefan-Boltzmann law: grey body

Natural objects are not full radiators. The **emittance** from these objects (called grey bodies) is given by

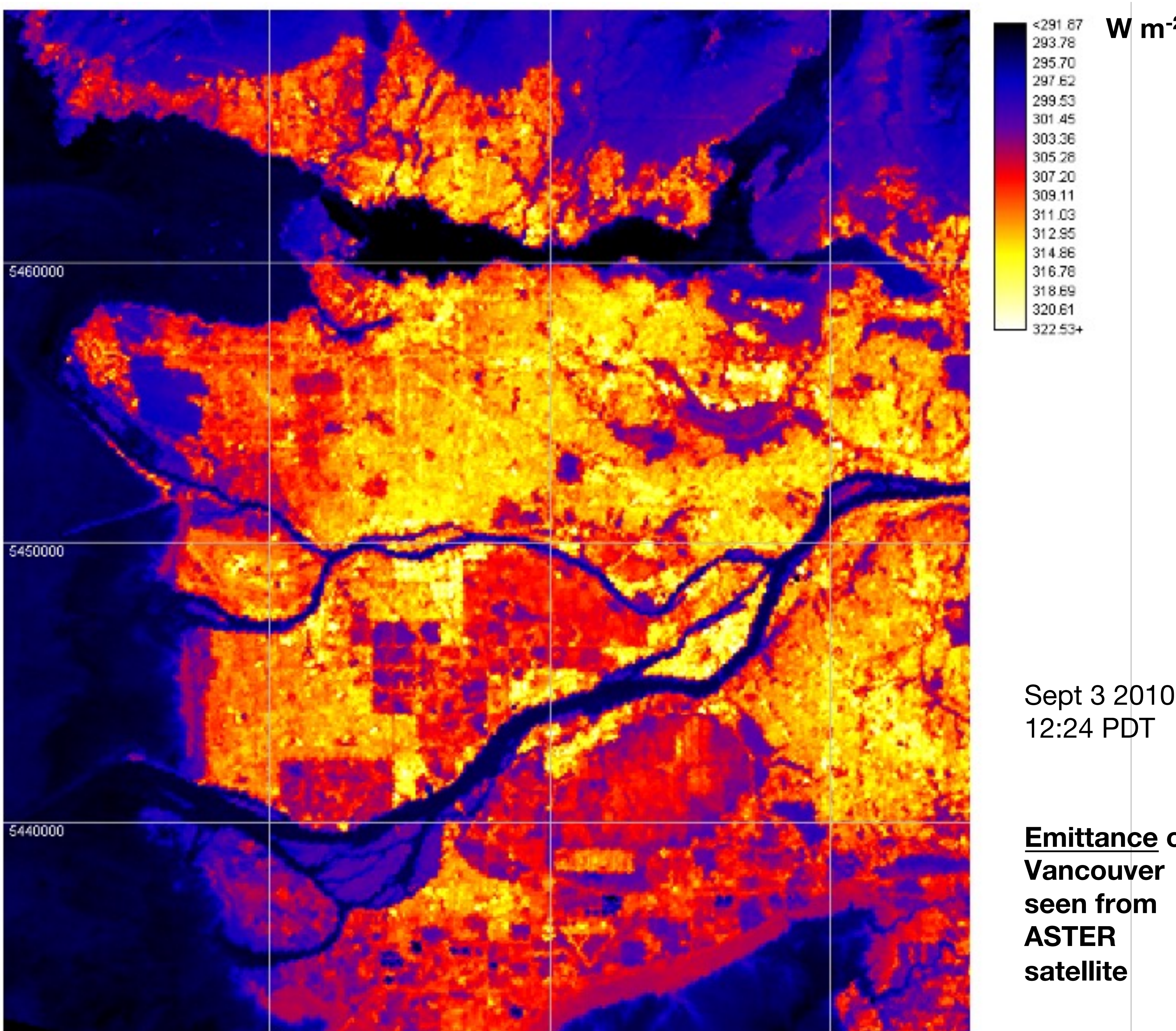
$$\boxed{E_g = \varepsilon \sigma T^4 \quad \star}$$

where ε is their **surface emissivity**. Emissivity is the ratio of the actual emission to that of a blackbody (i.e. ε = 1.0).

This law is the basis of remote sensing in the TIR incl. satellite sensors.

Surface	Emissivity ε*
Soil	0.90 – 0.98
Grass	0.90 – 0.95
Crops	0.90 – 0.99
Forests	0.97 – 0.99
Water	0.92 – 0.97
Iron	0.13 – 0.28

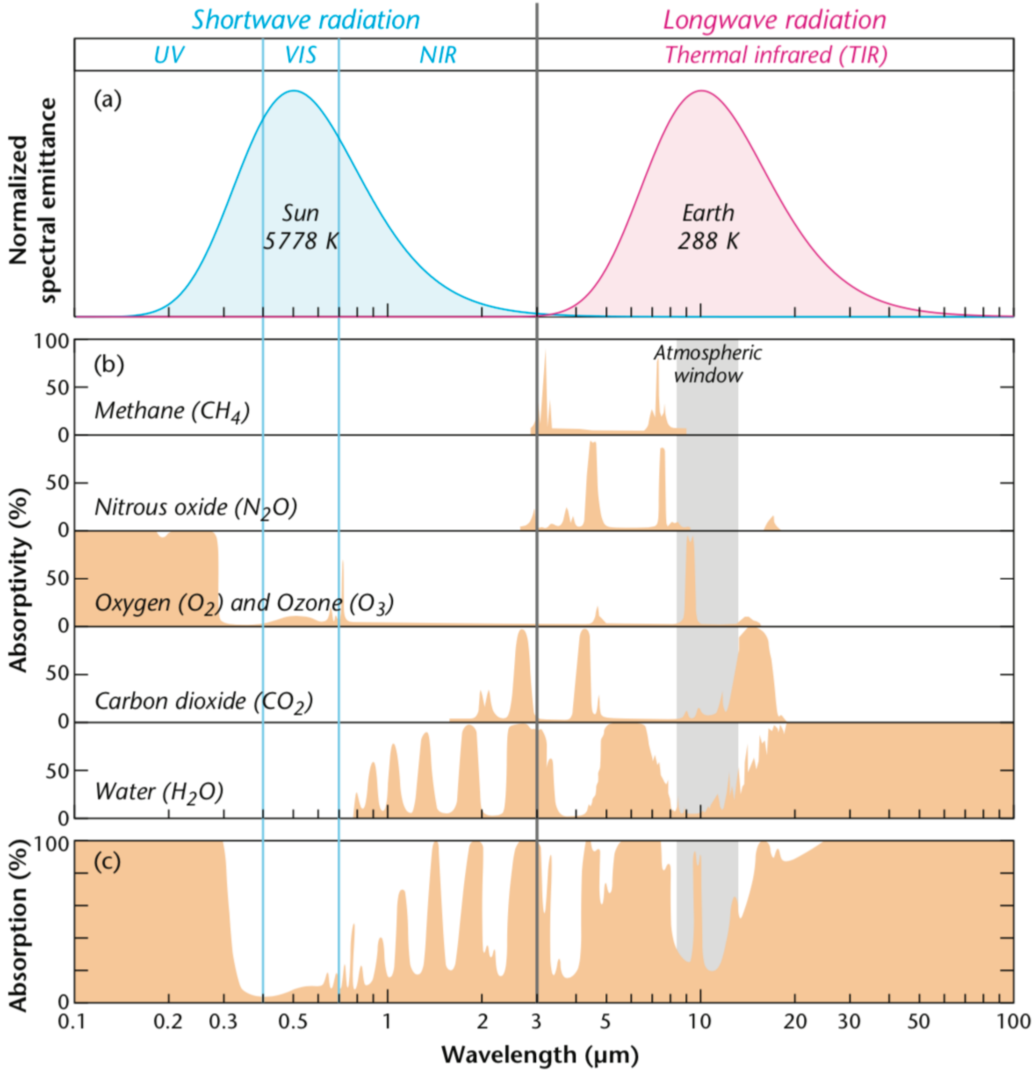
* in the long-wave range of the spectrum (relevant for microclimates)



Sept 3 2010
12:24 PDT

Emittance of Vancouver seen from ASTER satellite

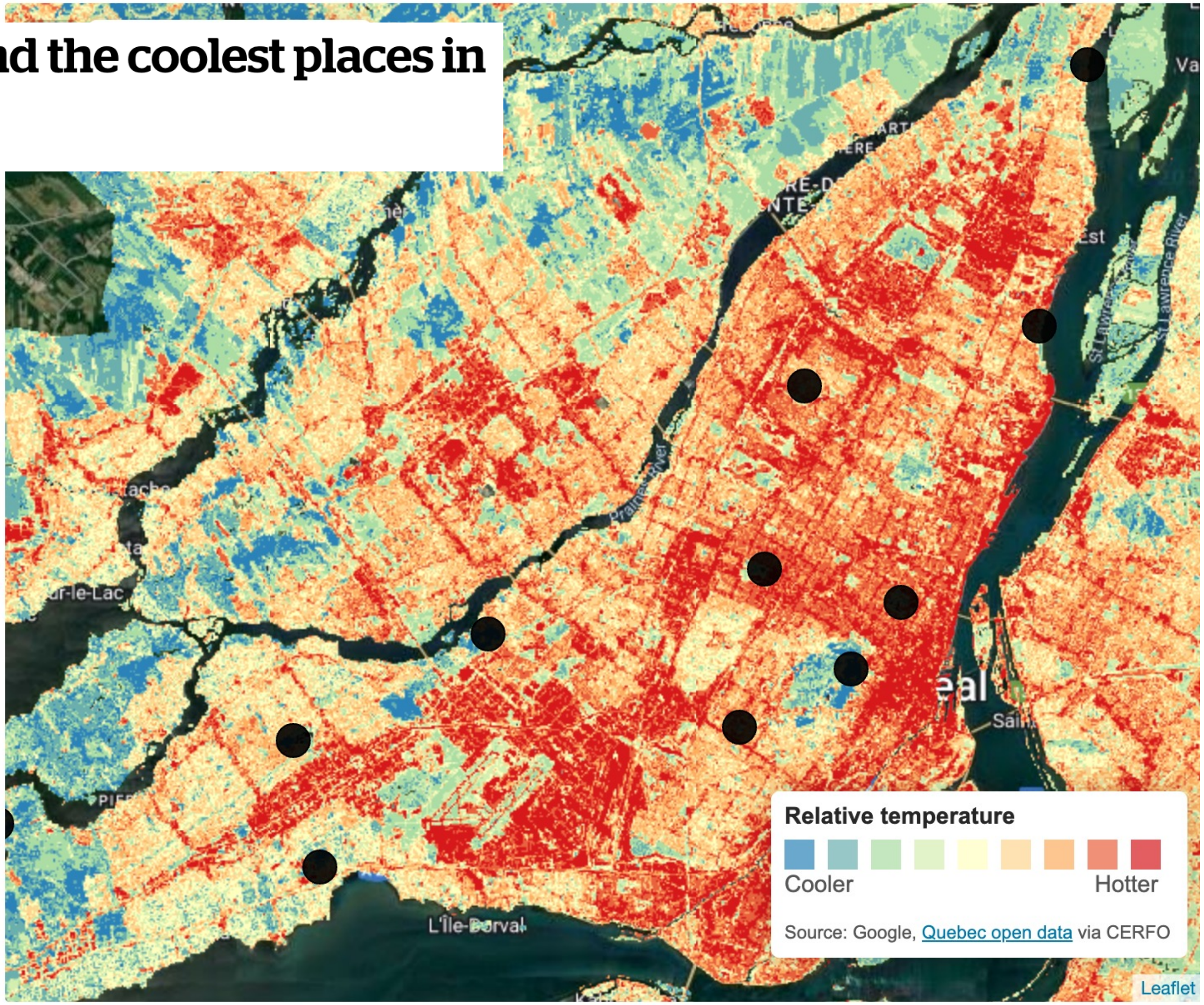
Atmospheric window & remote sensing in the TIR



From emittance we can estimate surface temperature

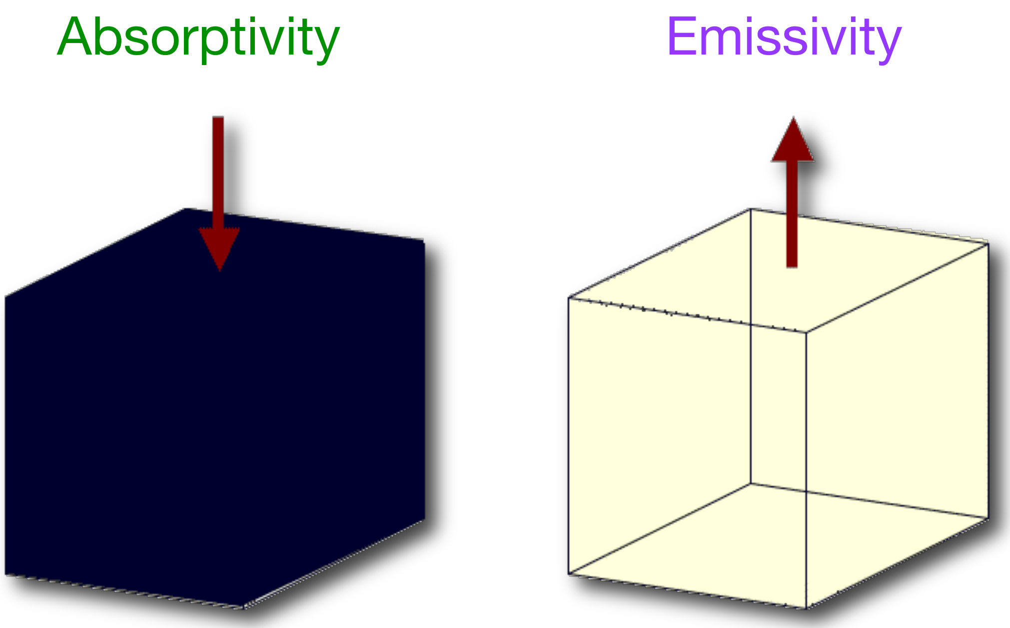
Where are the warmest and the coolest places in Montreal?

CBC



https://www.cbc.ca/news/canada/montreal/montreal-warm-cool-places-interactive-1.3669639

Review: Kirchhoff's law



$$\zeta\lambda = \varepsilon\lambda \quad \star$$

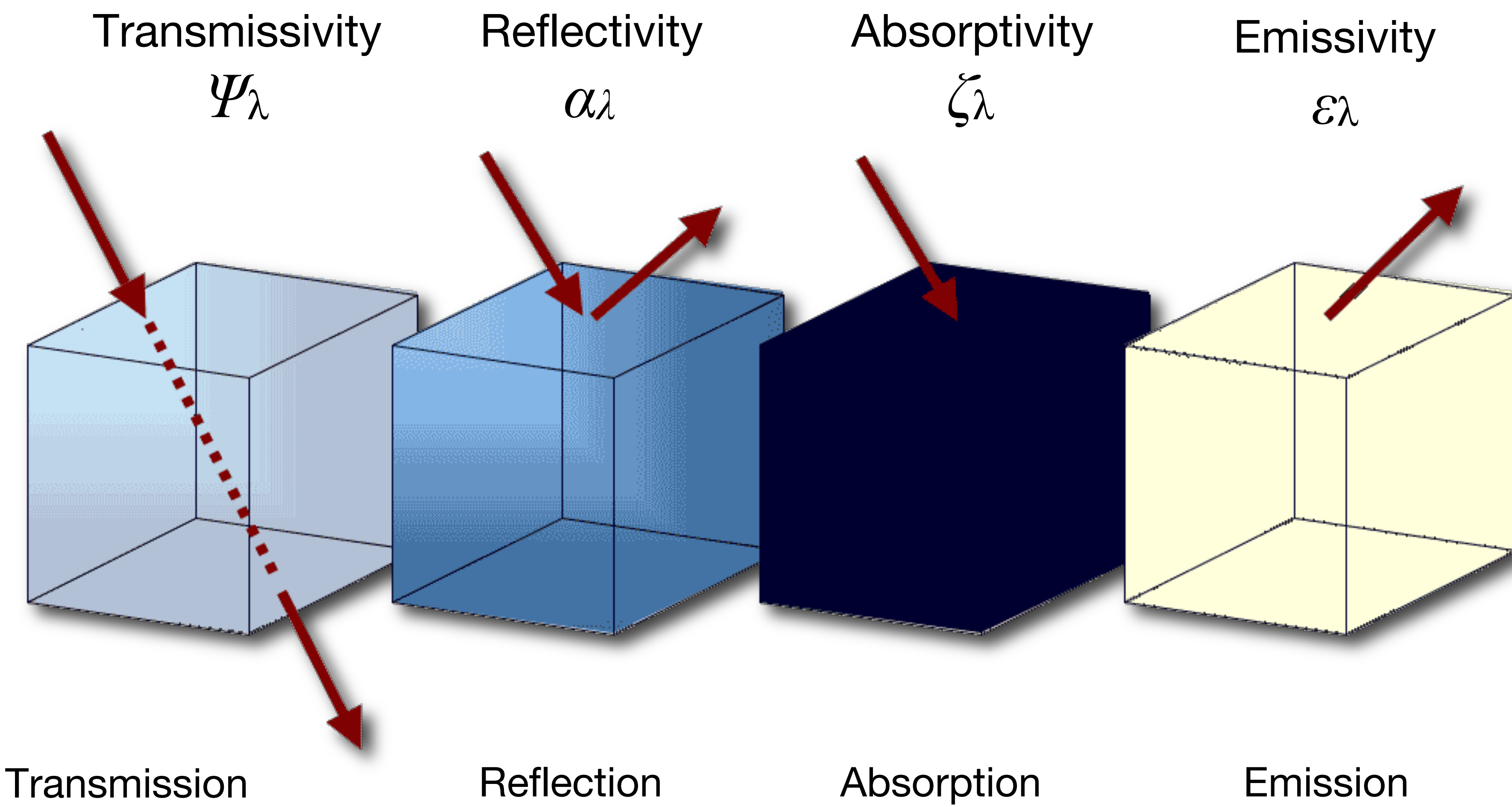
Assuming no transmission the absorptivity of a body ($\zeta\lambda$) equals its emissivity ($\varepsilon\lambda$) at a given wavelength.

A good absorber is a good emitter

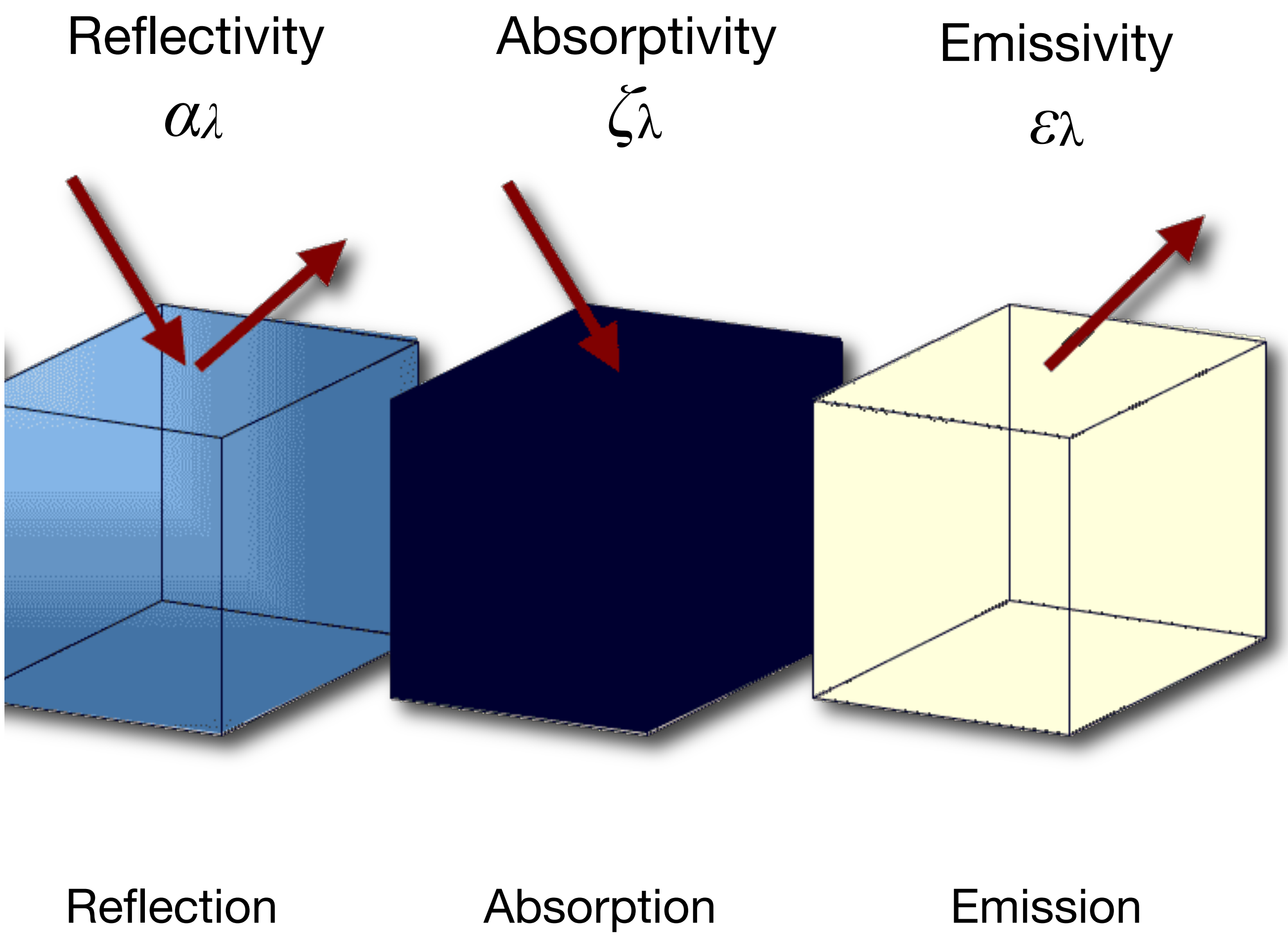
Kirchhoff’s law of thermal radiation

- Kirchhoff’s law only applies if the wavelength considered is the same – do not mix them together.
- Kirchhoff’s law only has relevance to **long-wave exchange** in climatology.
- The law does not apply to fluorescent objects, which can absorb energy at a given a wavelength and release it at another one.

Mass-radiation interactions



most surfaces are opaque to long-wave radiation therefore absorptivity typically = emissivity



Is long-wave reflection important?

	Transmissivity		Absorptivity		Reflectivity		
Energy conservation	$\psi_{o,LW}$	+	$\zeta_{o,LW}$	+	$\alpha_{o,LW}$	= 1.0	★

However, most surfaces are opaque to long-wave (i.e., $\psi_{o,LW}$ ~ 0) so

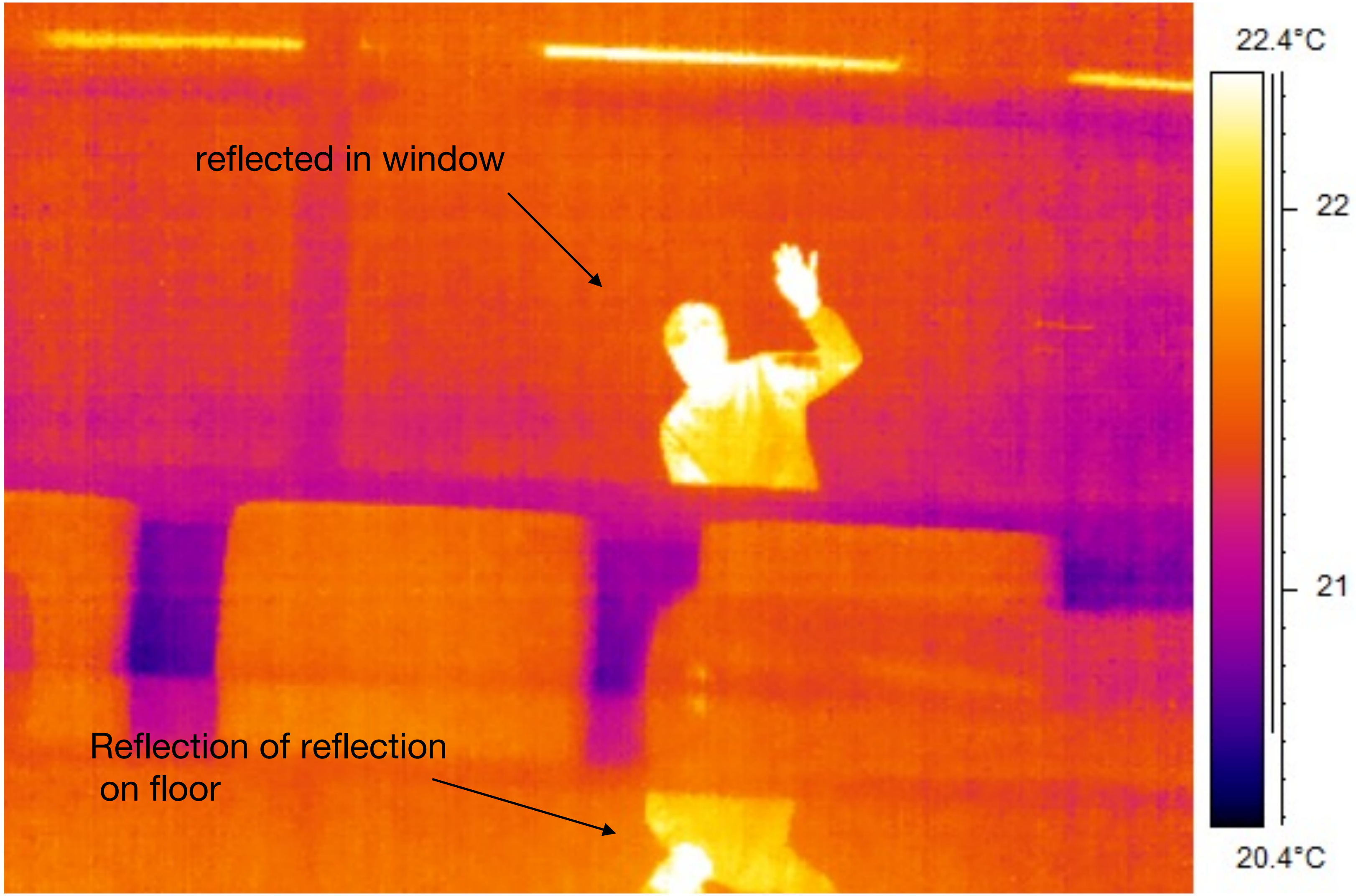
$$\zeta_{o\ LW} + \alpha_{o,LW} = 1.0$$

and since $\zeta_{o\ LW} = \varepsilon_{o,LW}$ (according to Kirchhoff's Law)

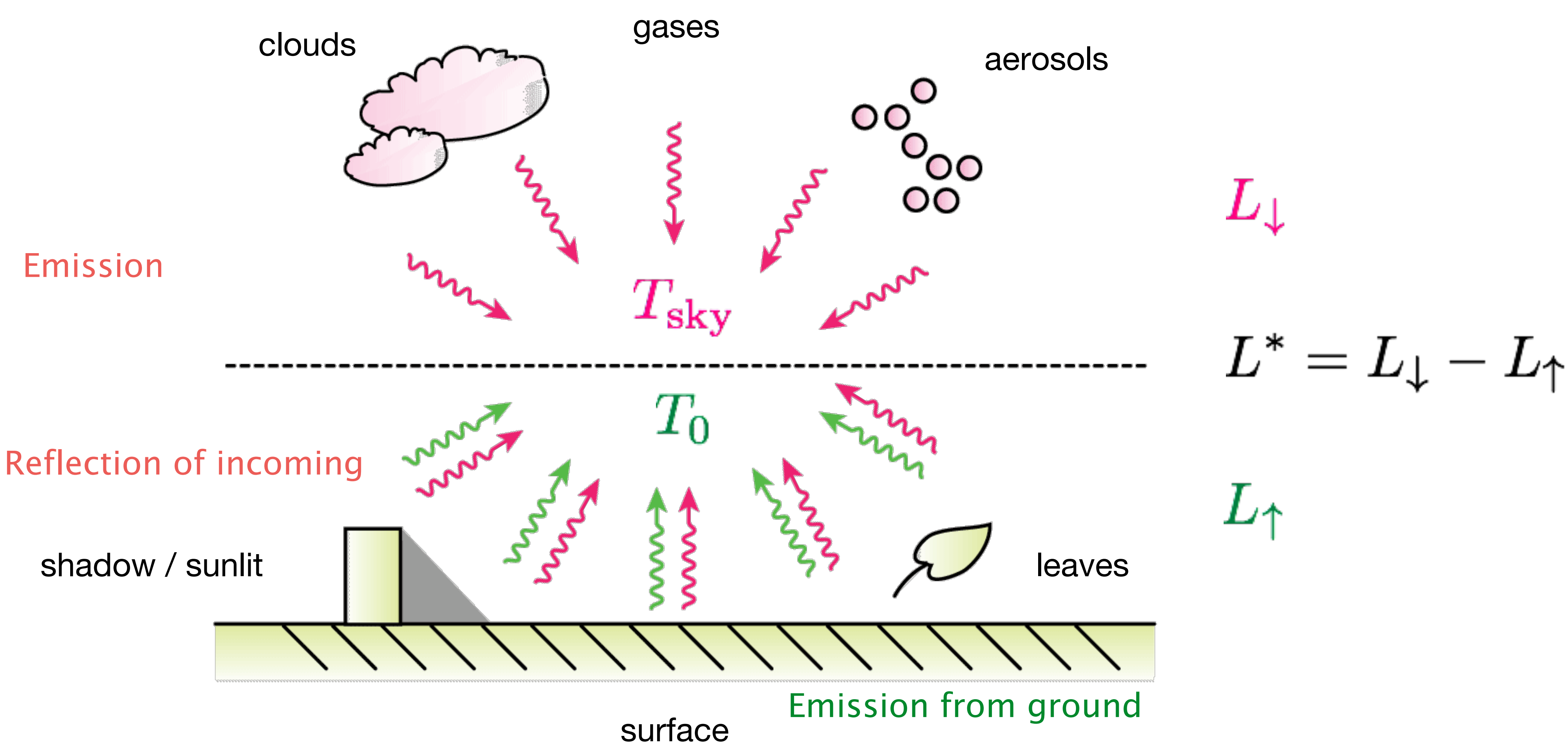
$$\alpha_{o,LW} = 1 - \varepsilon_{o,LW}$$

The subscript '0' refers to values of the land surface (in contrast to atmosphere with subscript *a*). In the upcoming slides we will implicitly assume values are in the long-wave band, and omit the subscript 'LW'

Long-wave reflection



Long-wave radiation exchange of a surface

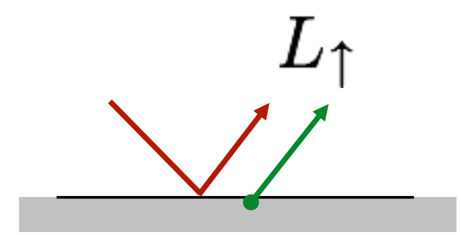


Calculating L↑

$$\varepsilon = \frac{\text{radiative flux density emitted by a body}}{\text{radiative flux density emitted by a blackbody}}$$

defines blackbody as ε = 1.0, and grey bodies as ε < 1.0.

$$L_{\uparrow} = \underbrace{\varepsilon_o \sigma T_o^4}_{\text{emission}} + \underbrace{\underbrace{(1-\zeta_o)}_{\alpha_o} L_{\downarrow}}_{\text{reflection in the long-wave}}$$



where ζ_o - absorptivity of surface in the long-wave, α_0 reflectivity of surface in the long-wave and from Kirchhoff's law ($\zeta_\lambda = \varepsilon_\lambda$):

$$L_{\uparrow} = \varepsilon_o \sigma T_o^4 + (1-\varepsilon_o) L_{\downarrow} \quad \star$$

where subscript '0' indicates surface value.

Net long-wave radiation flux density L^*

The long-wave net radiation L^* at the surface is the difference between the input from above $L_\downarrow$ and the output from emission and reflected $L_\downarrow$:

$$L^* = L_\downarrow - L_\uparrow$$

$$L^* = L_\downarrow - (\varepsilon_o \sigma T_o^4 + \overbrace{(1-\zeta_o)}^{\alpha_o} L_\downarrow)$$

$$L^* = L_\downarrow - (\varepsilon_o \sigma T_o^4 + (1-\varepsilon_o) L_\downarrow)$$

emitted

reflected

Kirchhoff's Law

Question for discussion (Slido)

What influences $L_\downarrow$?

Calculating L↓ (if you don't have a direct measurement of L↓)

Screen level observations are typically available from regular climate stations for air temperature (T_a) and sometimes vapour pressure (e_a) but not $L_\downarrow$

If we are interested in estimating $L_\downarrow$ across full long-wave range (3 to 100 µm) we need a bulk value of ε_a for both cloudless and cloudy cases.

Several equations are available to estimate ε_a from screen-level T_a and e_a.

The subscript 'a' refers to values the entire atmosphere

A standard Stevenson-screen sheltering sensors to measure air temperature and humidity (Photo: A. Christen)

Equations for apparent atmospheric emissivity (ε_a) with cloudless skies

1. *Brunt (1932)*

$$\varepsilon_a = a + b.e_a^{1/2}$$ a ~ 0.61, b ~ 0.05 (vary geog.)

2. *Swinbank (1963)*

$$\varepsilon_a = a.T_a^{2}$$ a = 9.2 x 10^{-6} (only T_a > freezing)

3. *Brutsaert (1975)*

$$\varepsilon_a = a\ (e_a/T_a)^{1/7}$$ a = 1.24

4. *Idso and Jackson (1969)*

$$\varepsilon_a = 1 - a.\exp\{\ b(273 - T_a)^2\}$$ a = 0.261, b = -7.77 x 10^{-4}

4. *Idso (1981)*

$$\varepsilon_a = a + b.e_a.\exp\{1500/T_a\}$$ a = 0.70, b = 5.95 x 10^{-5}

5. *Berdahl and Martin (1984)*

$$\varepsilon_a = a + b(T_d/100) + c(T_d/100)^2$$ a = 0.711, b = 0.56, c = 0.73

6. *Prata (1996)*

$$\varepsilon_a = [1-(1+\zeta)\ \exp\{-(a + b\zeta)^{1/2}\}\ ]$$ a = 1.2, b = 3.0, $\zeta = 46.5(e_a/T_a)$

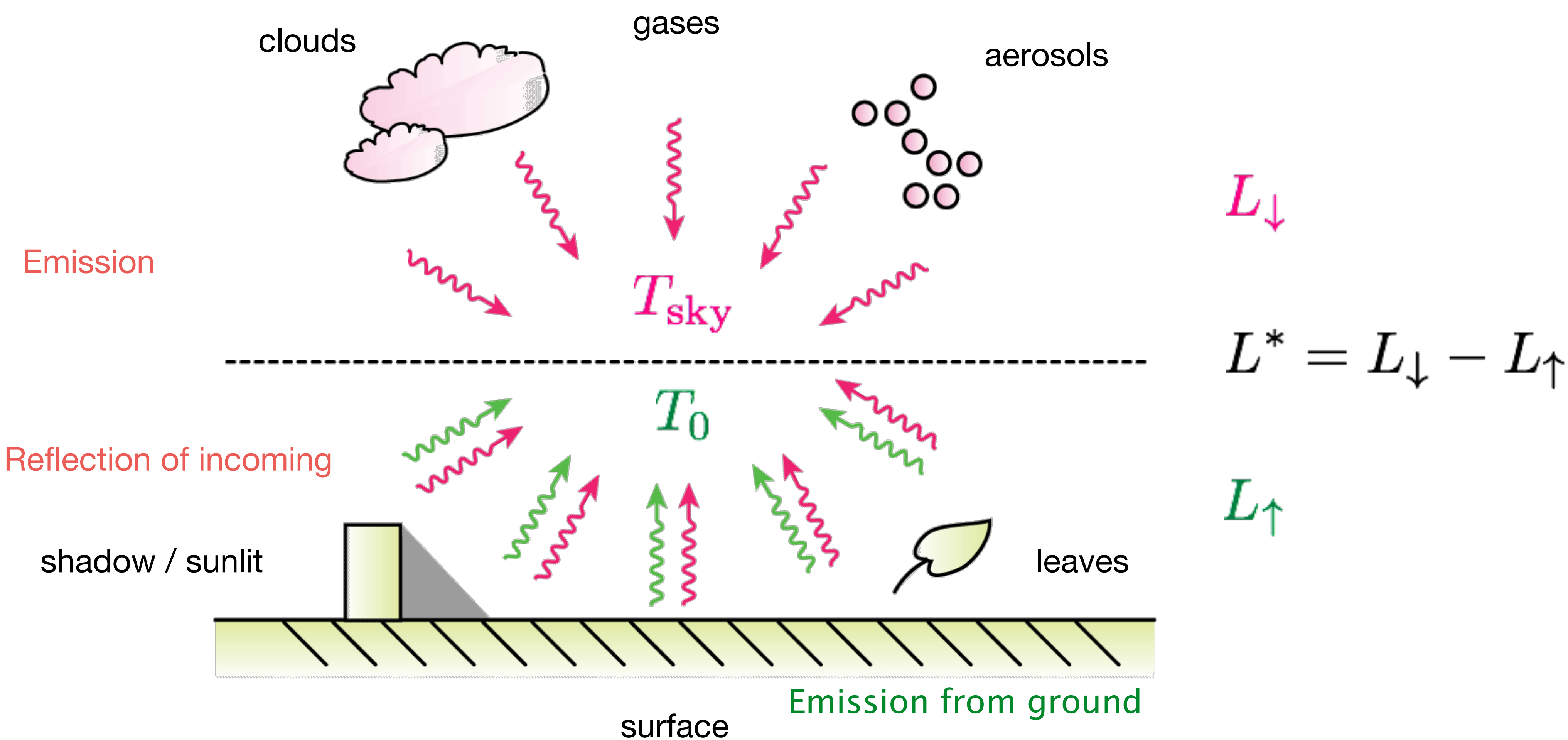
[T_a - air temperature (K); e_a - vapour pressure (mb); T_d - dew-point temperature (°C)]

Calculating L↓ - How to estimate ε_a

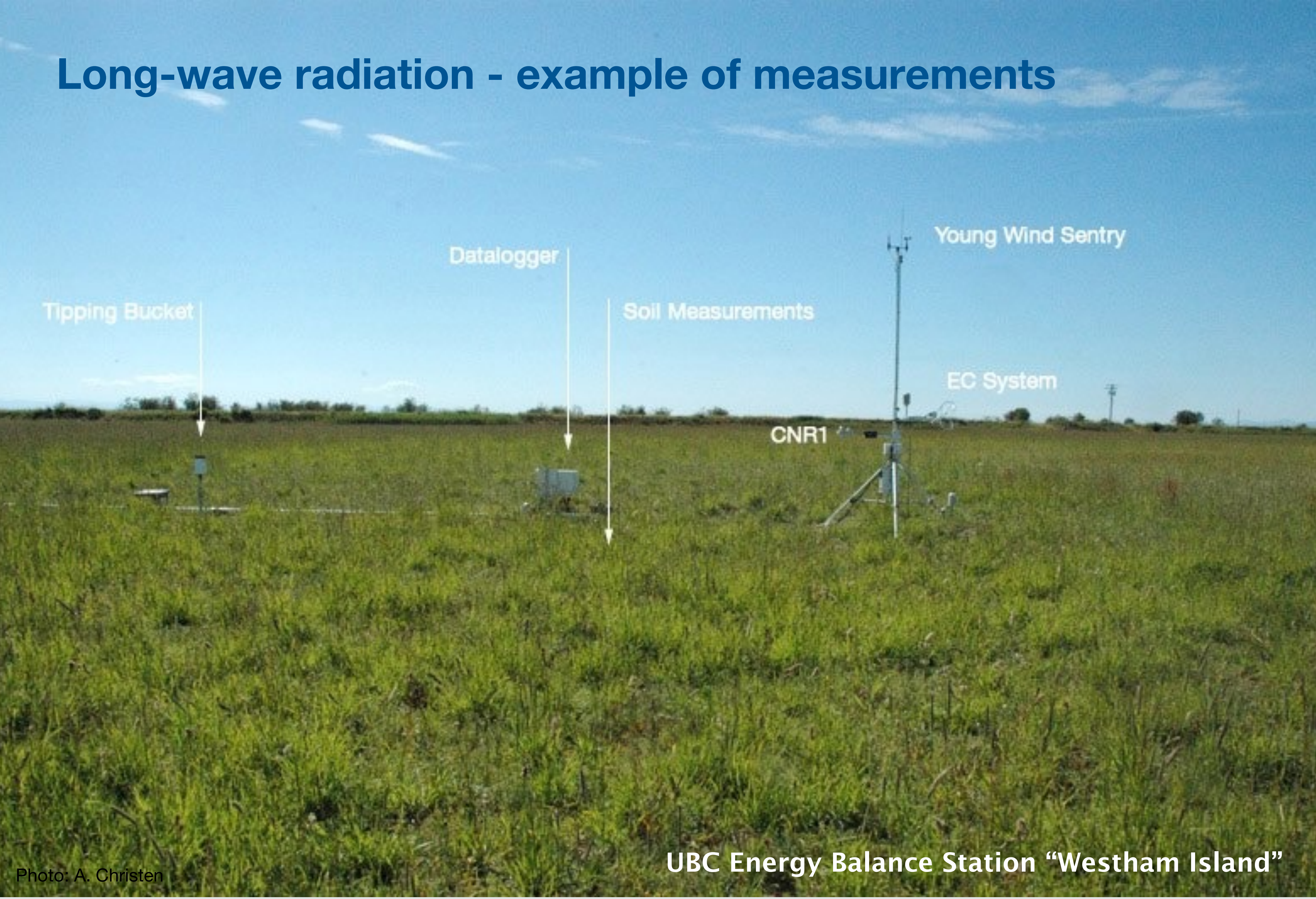
Using T_a and e_a to estimate ε_a works because:

- T_a and e_a are the main controls that change in the atmosphere
- Variations in CO_2, O_3 and other greenhouse gases are small.
- T_a and e_a are largest and show their greatest variation near the ground. Approximately 50% of $L_\downarrow$ originates from 0 to 100m.

Long-wave radiation exchange of a surface



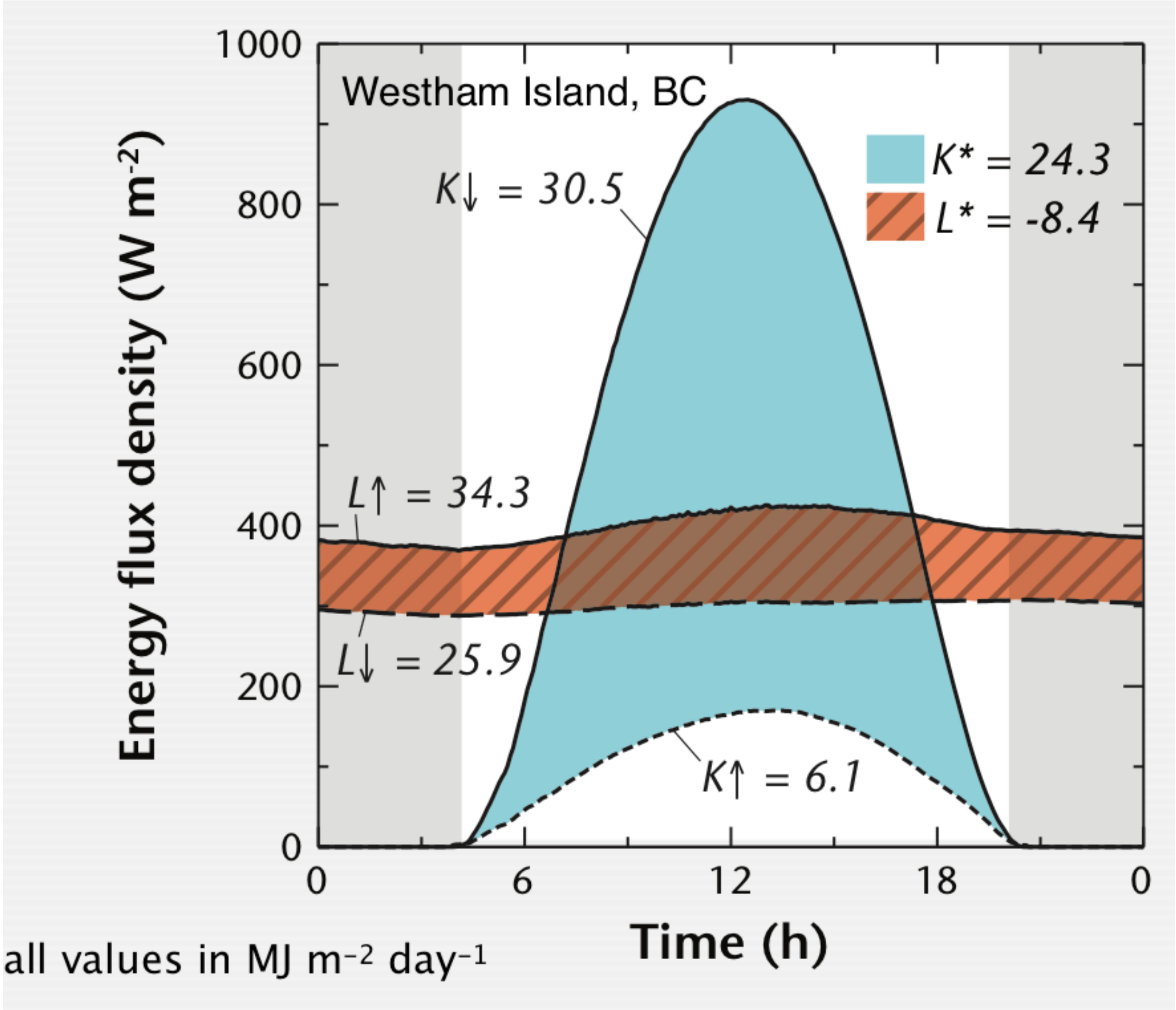
Long-wave radiation - example of measurements



Long-wave radiation - example of measurements

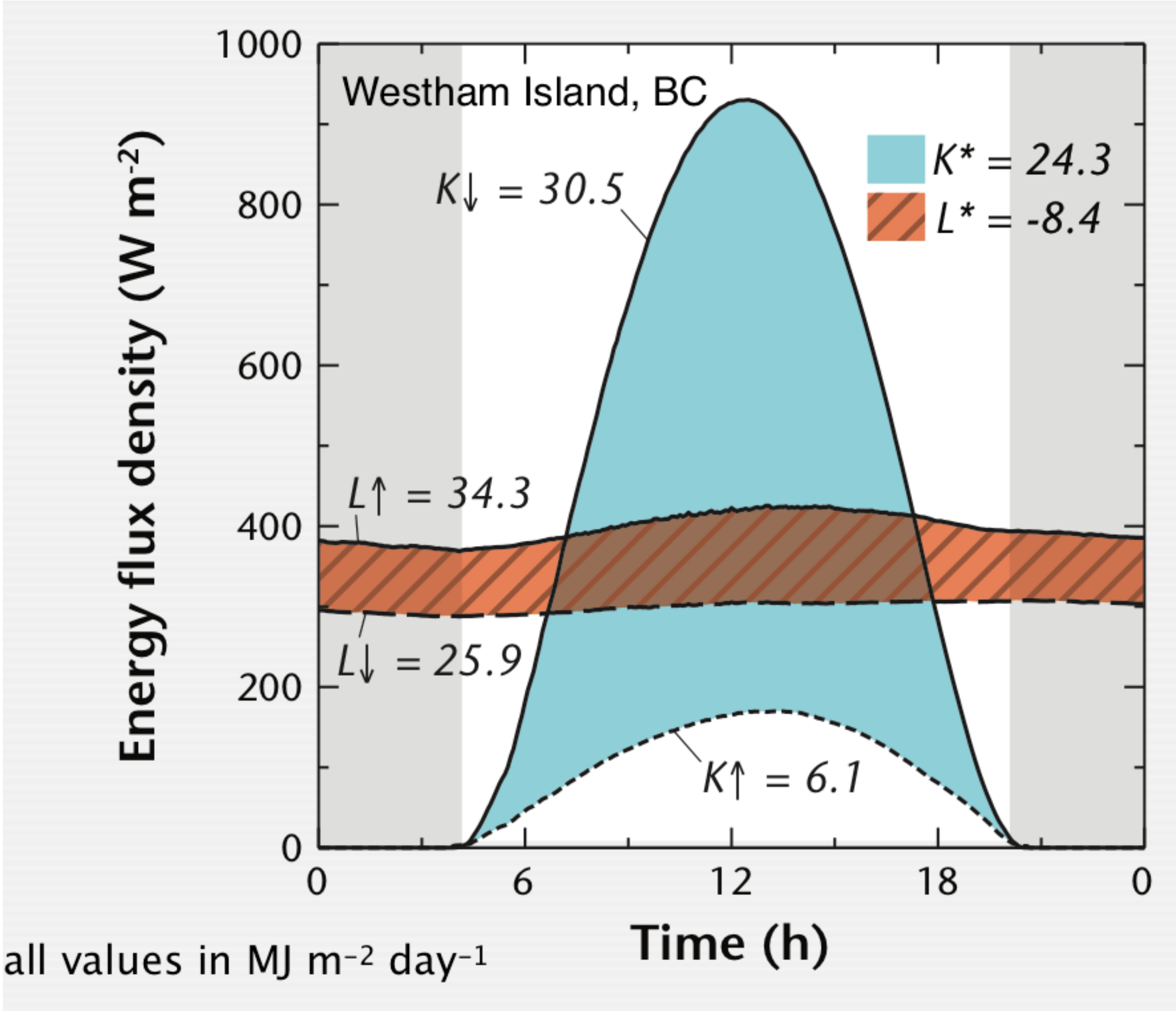
Photo: J. Verfaillie

Diurnal course of radiation - clear sky day



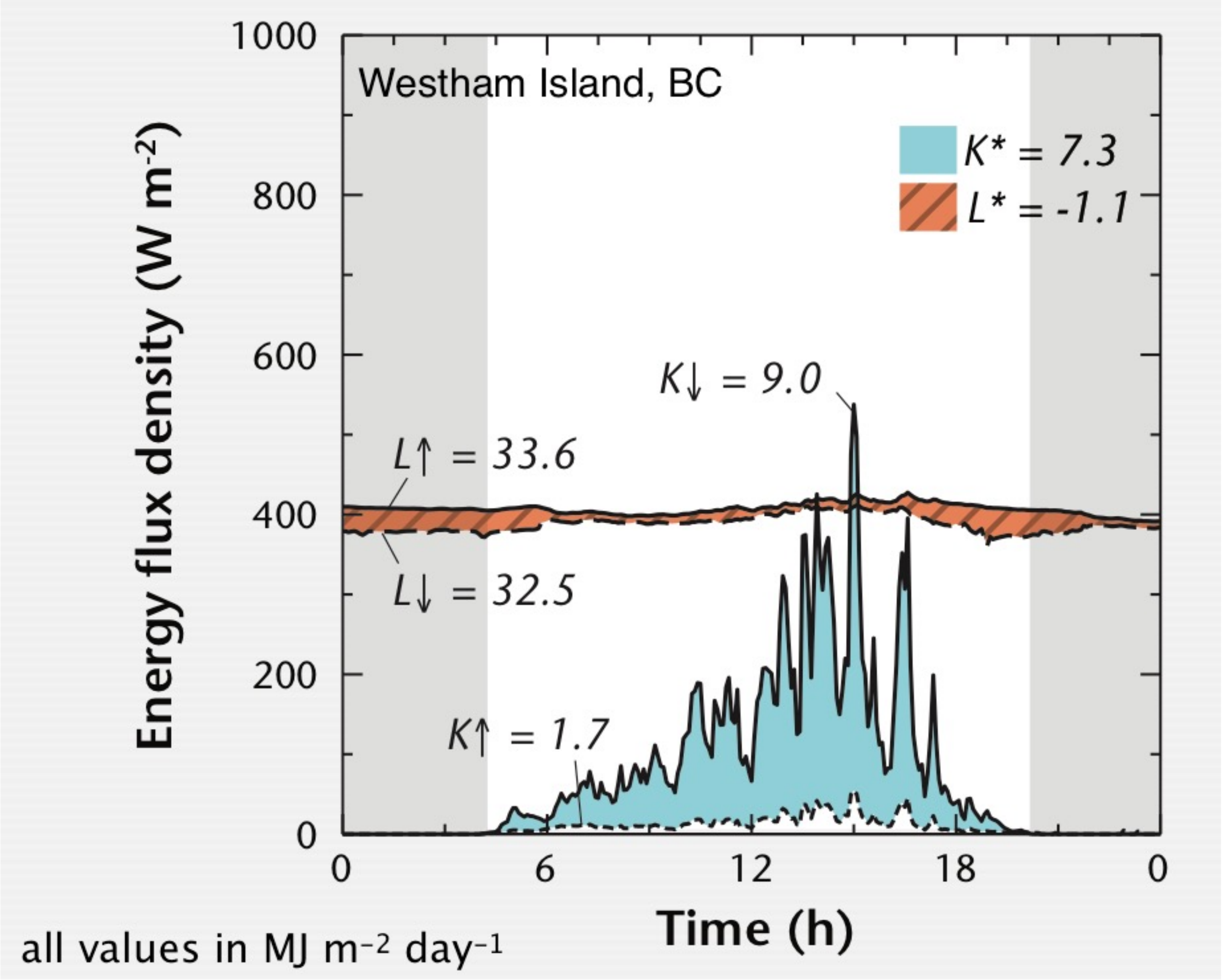
Images from Katkam in Vancouver

Class activity - Why is $L_{up} > L_{down}$? (Slido)



Images from Katkam in Vancouver

Diurnal course of radiation - overcast / broken sky day

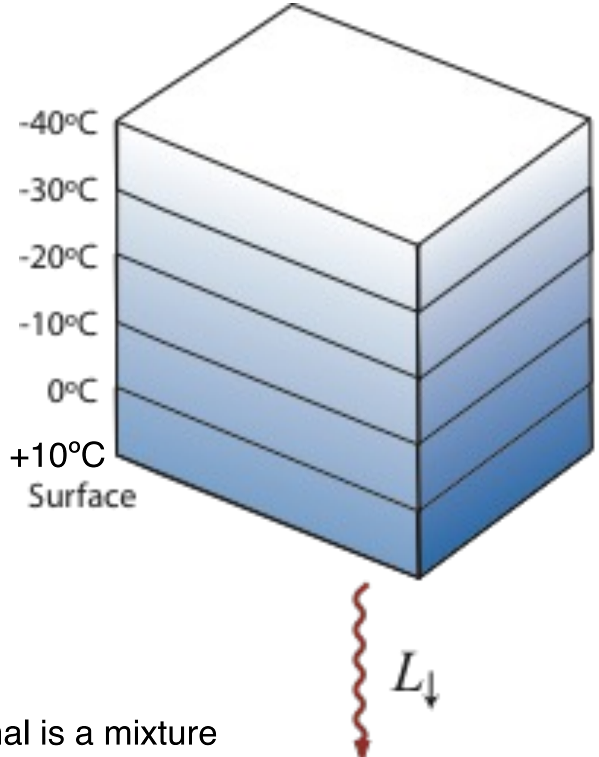


Images from Katkam in Vancouver

Effect of clouds on longwave irradiance

With a clear-sky, L↓ originates from **all layers** of the atmosphere, because the atmosphere is **partly transparent** ('atmospheric window' open)

With clouds L↓ originates from **the cloud base** and the atmosphere below the cloud, because the cloud is **opaque to long wave** ('atmospheric window' closed)



Signal is a mixture from -40°C to +10°C

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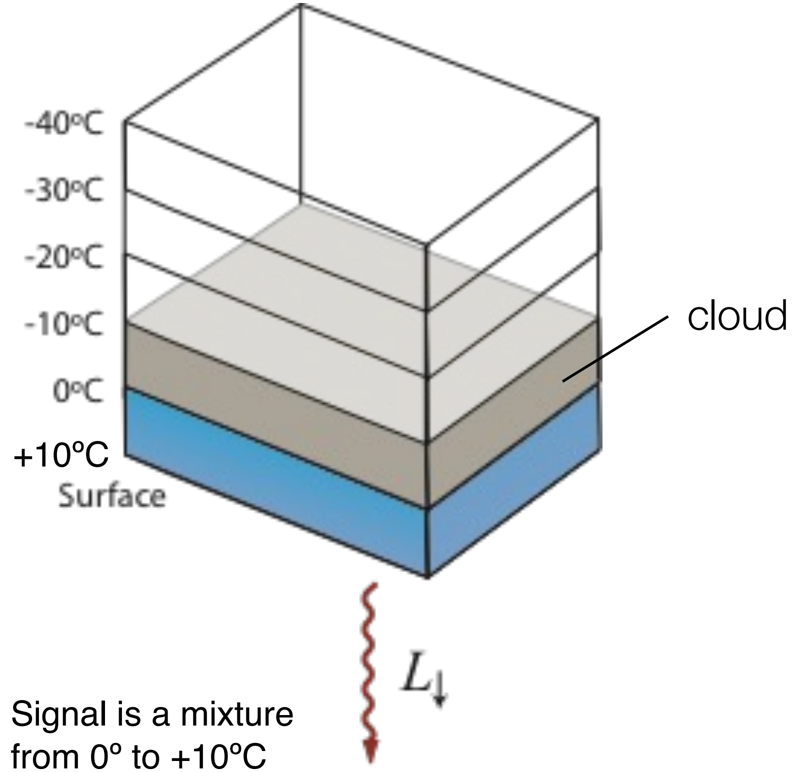
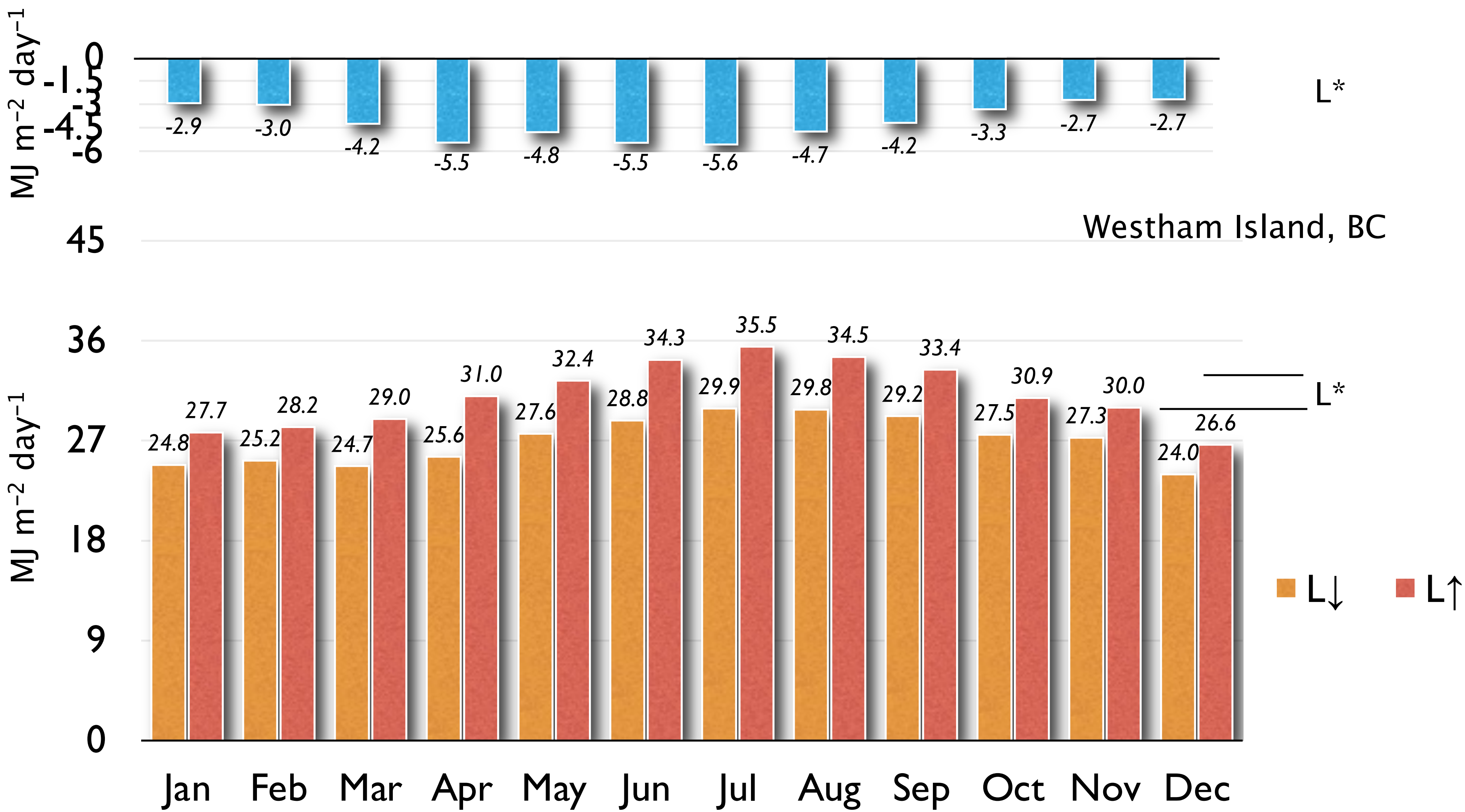
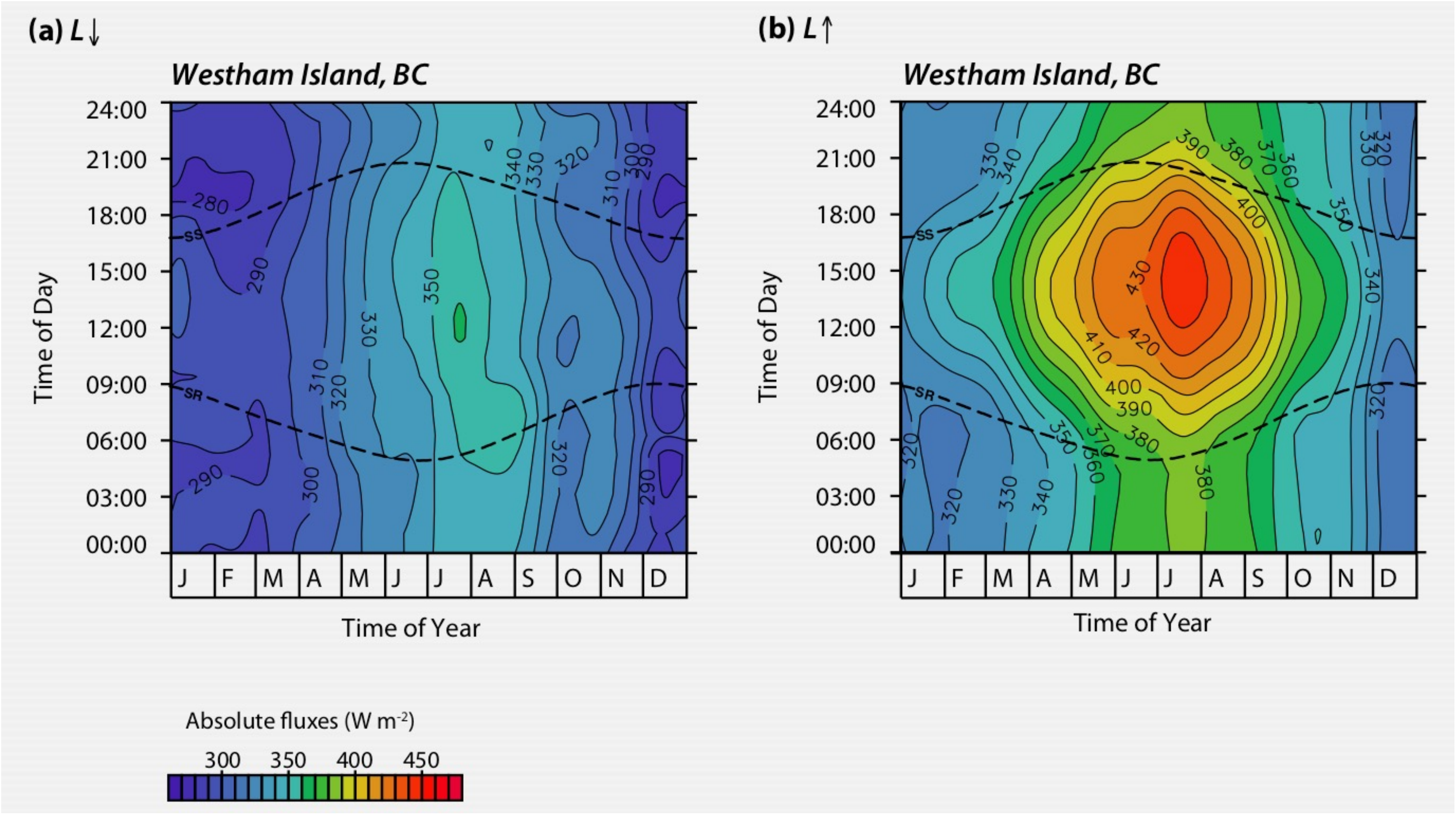


Signal is a mixture from 0° to +10°C

Monthly totals of L↓, L↑ and L* in Vancouver



‘Fingerprint’ of L↓ and L↑ measured in Vancouver



Take home points

- Radiation laws apply the same way to the longwave part of the spectrum - but **Kirchhoff's law** and the concept of **emissivity** become relevant.
- The net-long wave radiation is driven by the difference in apparent sky and surface temperatures and hence **clouds and thermal surface properties** are controlling radiative exchange in the long-wave.