A micrometeorology / energy balance site in a tidal wetland in Puget Sound, Washington (Photo: F. Anderson)

11 Soil heat transfer

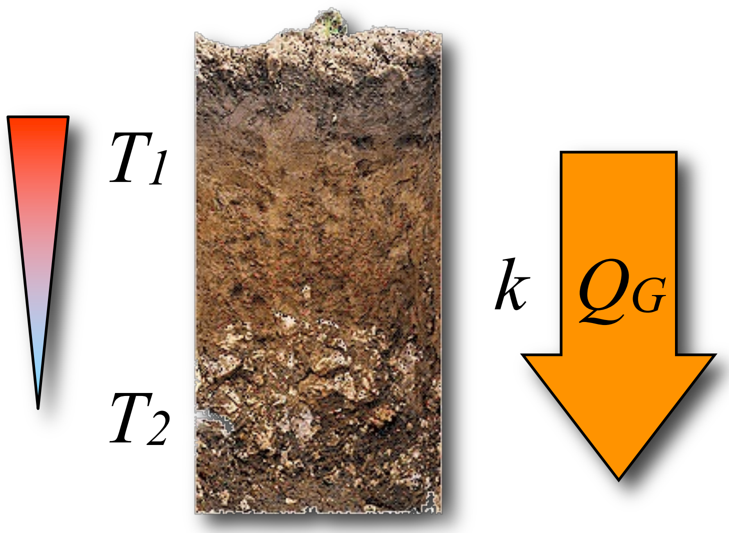
Learning objectives

- Explain how we **measure** heat storage changes, and transfer in and out of soils.
- Describe how we **predict** and model heat storage changes, and transfer in and out of soils.
- Explain how heat storage in soils changes on diurnal and annual scales.

Installation of soil sensors in an organic soil of the Low Arctic (Illisarvik)
Photo: W. Skeeter (UBC Geography)

Measurement of Q_G

Theoretically, soil heat flux could be simply measured using a vertical array of thermocouples using Fourier's law



Measuring soil heat flux using a temperature profile

Soil heat flux at a given depth in W m^{-2}

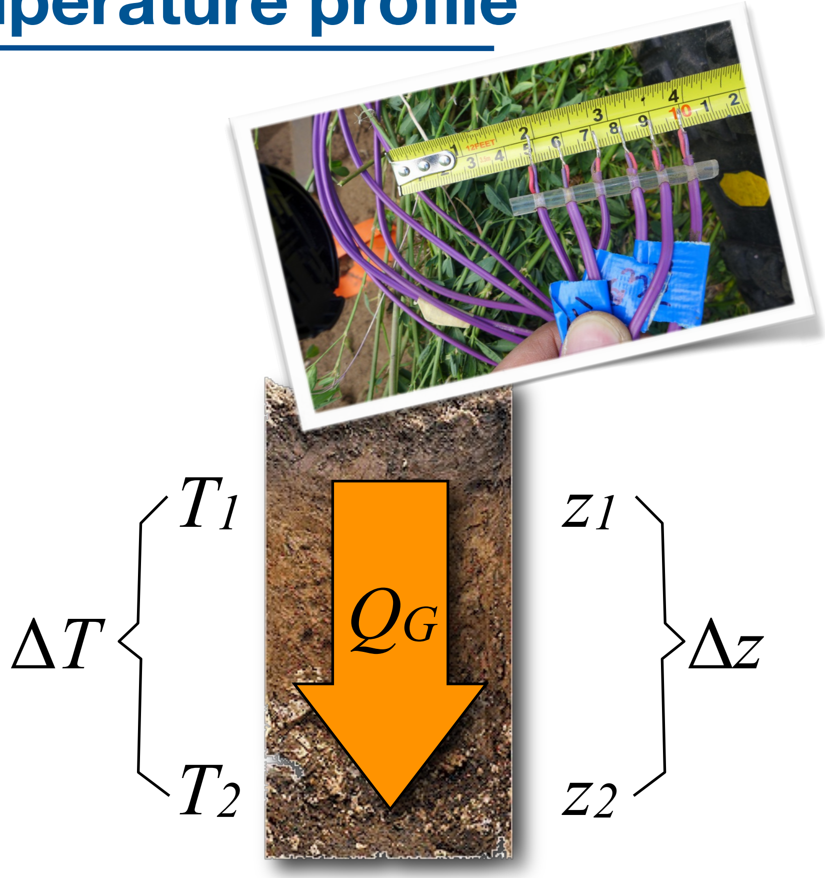
Temperature gradient (change with depth) in K m^{-1}

Difference between two temperatures...

$$Q_G = \boxed{-k}\,\boxed{\frac{\partial T}{\partial z}} \approx -k\,\frac{T_2 - T_1}{z_2 - z_1}$$

Thermal conductivity in W m^{-1} K^{-1}

... measured at two different depths



The role of soil water content

Practically, the problem is the variably ***k*** with water content. Requires continuous measurements of soil volumetric water content θ_w.

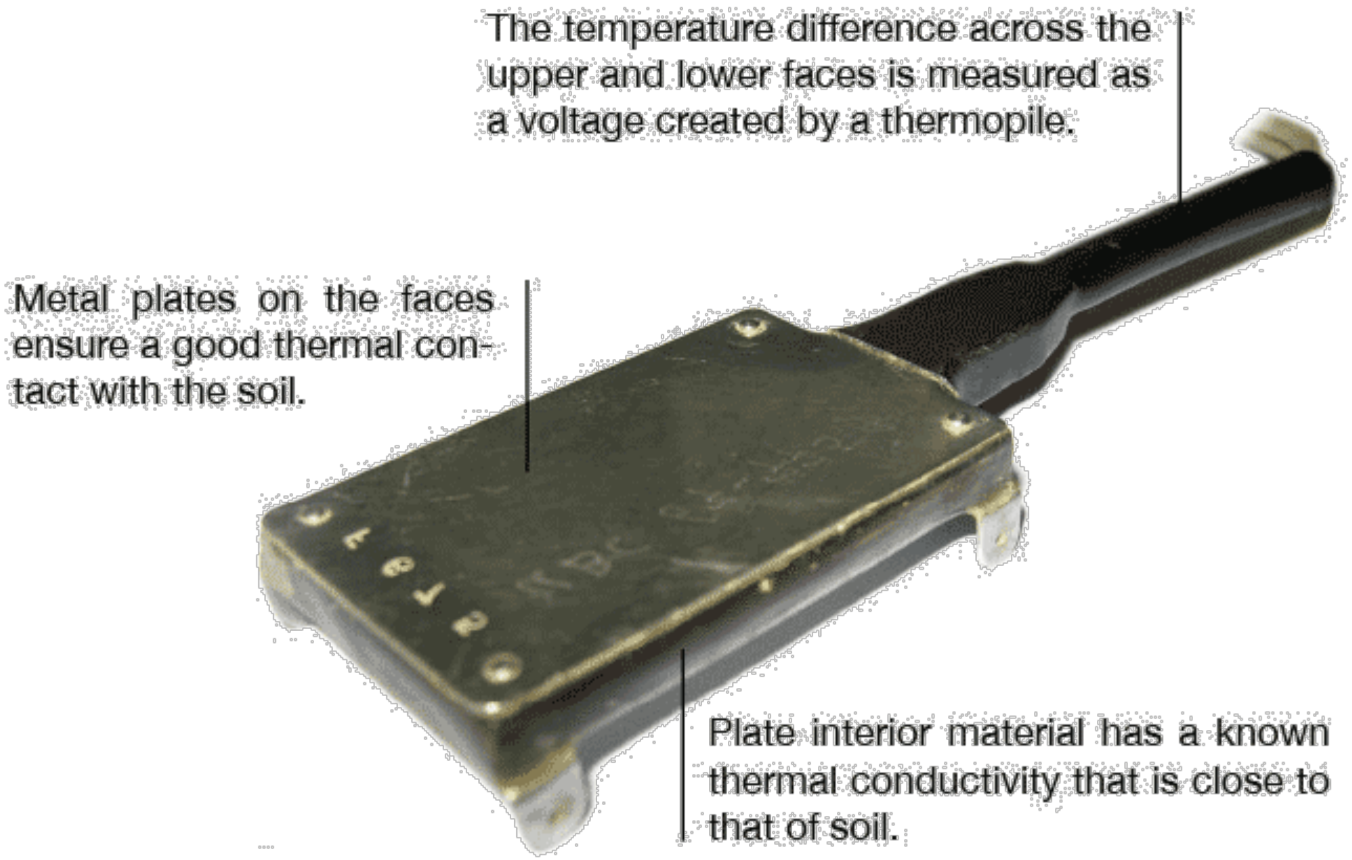
A UBC Geography undergraduate measures the volumetric water content of a soil using a TDR.

The role of soil water content

θ_w can be measured in the lab using the gravimetric method or using time domain reflectometry (TDR) (i.e. an indirect measure of soil water content based on the travel time of a high frequency electromagnetic pulse through the soil)

Soil heat flux density - direct measurement

More useful are **soil heat flux plates**:



A floating flux tower in
Burns Bog, Delta, BC
operated by UBC Geography


2

Soil heat flux plates - correction

Soil heat flux plates are inserted in undisturbed soil, not at the surface but in a few cm depth.

We have to consider the soil heat flux divergence / convergence in the topmost layer.

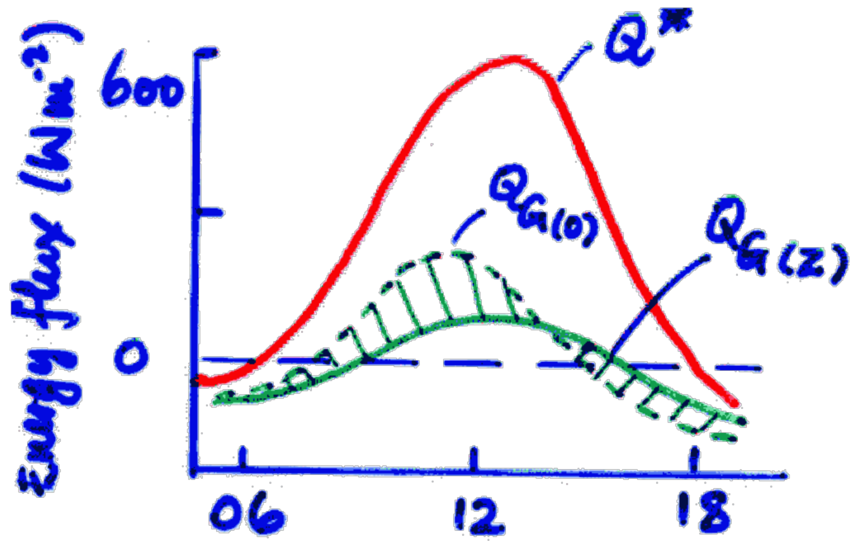
$$Q_{G(0)} = Q_{G(z)} + \Delta Q_{G(0-z)}$$

Using heat capacity:

Field observations

$$\Delta Q_{G(0-z)} = C_s \left(\frac{\Delta T}{\Delta t} \right) \Delta z_{0-z}$$

Measured using a temperature profile.



Typical effect of correction during a clear-sky day for bare soil.

Test your knowledge (Class activity - Slido)

At 11:30 in the morning, we measure a soil heat flux density Q_G(5cm) of 25 W m^{-2} using a heat flux plate installed at 5 cm depth. Calculate the soil heat flux density at the surface $Q_G(0)$, if the soil's heat capacity in the layer from 0 to 5 cm depth is 2 MJ m^{-3} K^{-1} and the temperature in the same layer changed from 24.8°C at 11:00 to 25.3°C at 12:00.

$$Q_{G(0)} = Q_{G(z)} + \Delta Q_{G(0-z)}$$

$$\Delta Q_{G(0-z)} = C_s \left(\frac{\Delta T}{\Delta t} \right) \Delta z_{0-z}$$

Test your knowledge (Class activity) - Solutions

1. Assume a uniform and linear warming rate of the soil:

$$\begin{aligned}\frac{\Delta T}{\Delta t} &= \frac{25.3°\mathrm{C} - 24.8°\mathrm{C}}{1\,\mathrm{h}} \\ &= 0.5\mathrm{K\,h^{-1}} = 1.38 \times 10^{-4}\mathrm{K\,s^{-1}}\end{aligned}$$

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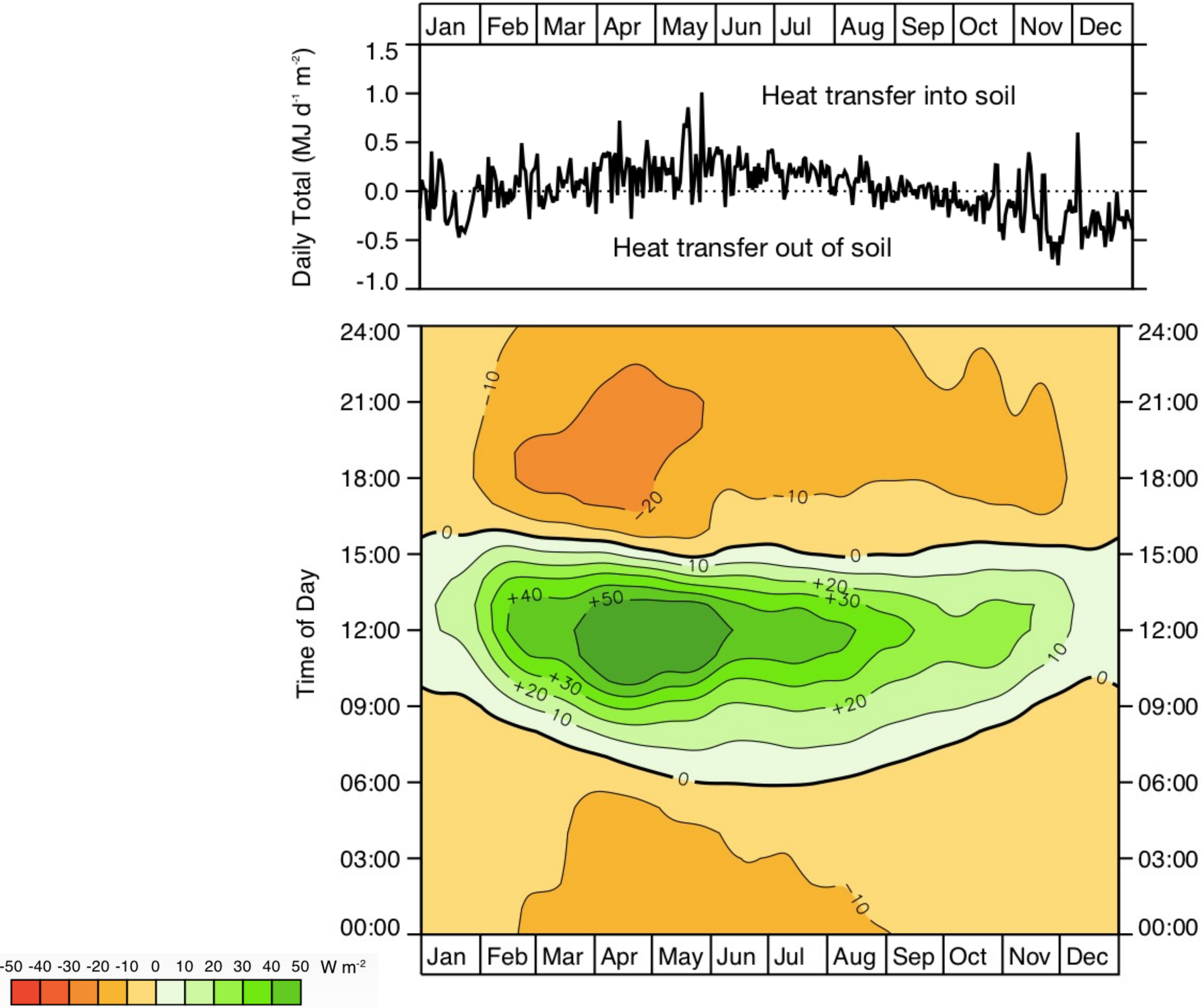
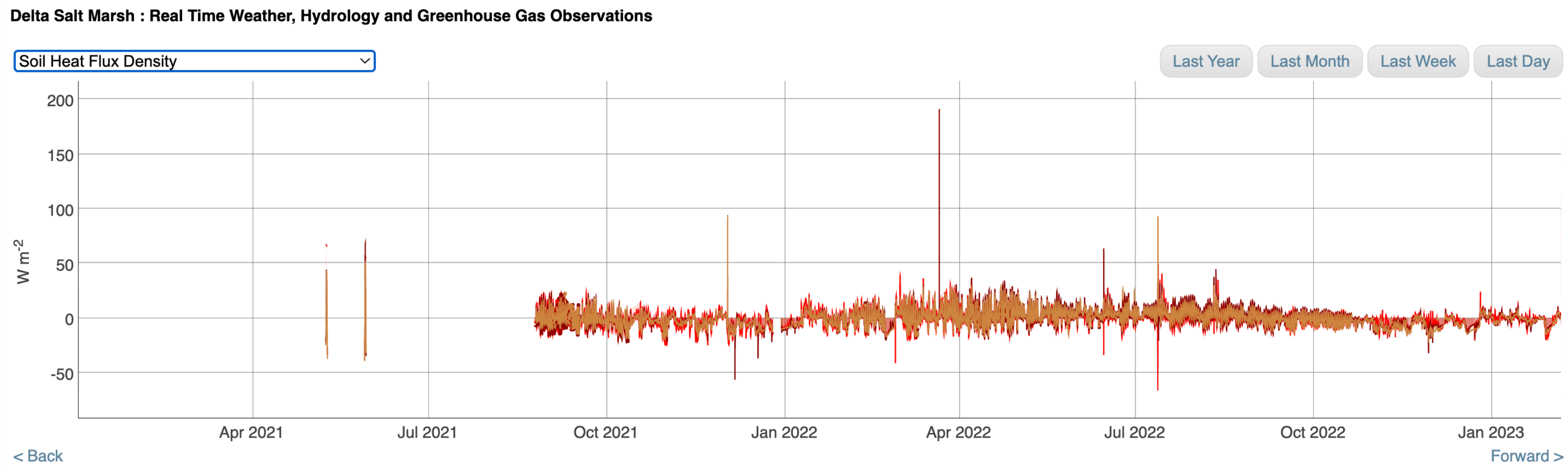
The heat flux at the surface $Q_{G(0)}$ is (Lecture 11, Slide 10):

$$\begin{aligned}Q_{G(0)} &= Q_{G(5\mathrm{cm})} + C\frac{\Delta T}{\Delta t}\Delta z \\ &= 25,\mathrm{W\,m^{-2}} + 2\,\mathrm{MJ\,m^{-3}\,K^{-1}} \times 1.38 \times 10^{-4}\mathrm{K\,s^{-1}} \times 0.05\,\mathrm{m} \\ &= 25\,\mathrm{W\,m^{-2}} + 13.8\,\mathrm{W\,m^{-2}} \\ &= \underline{38.8\,\mathrm{W\,m^{-2}}}\end{aligned}$$



Westham Island
Delta, BC, 2008

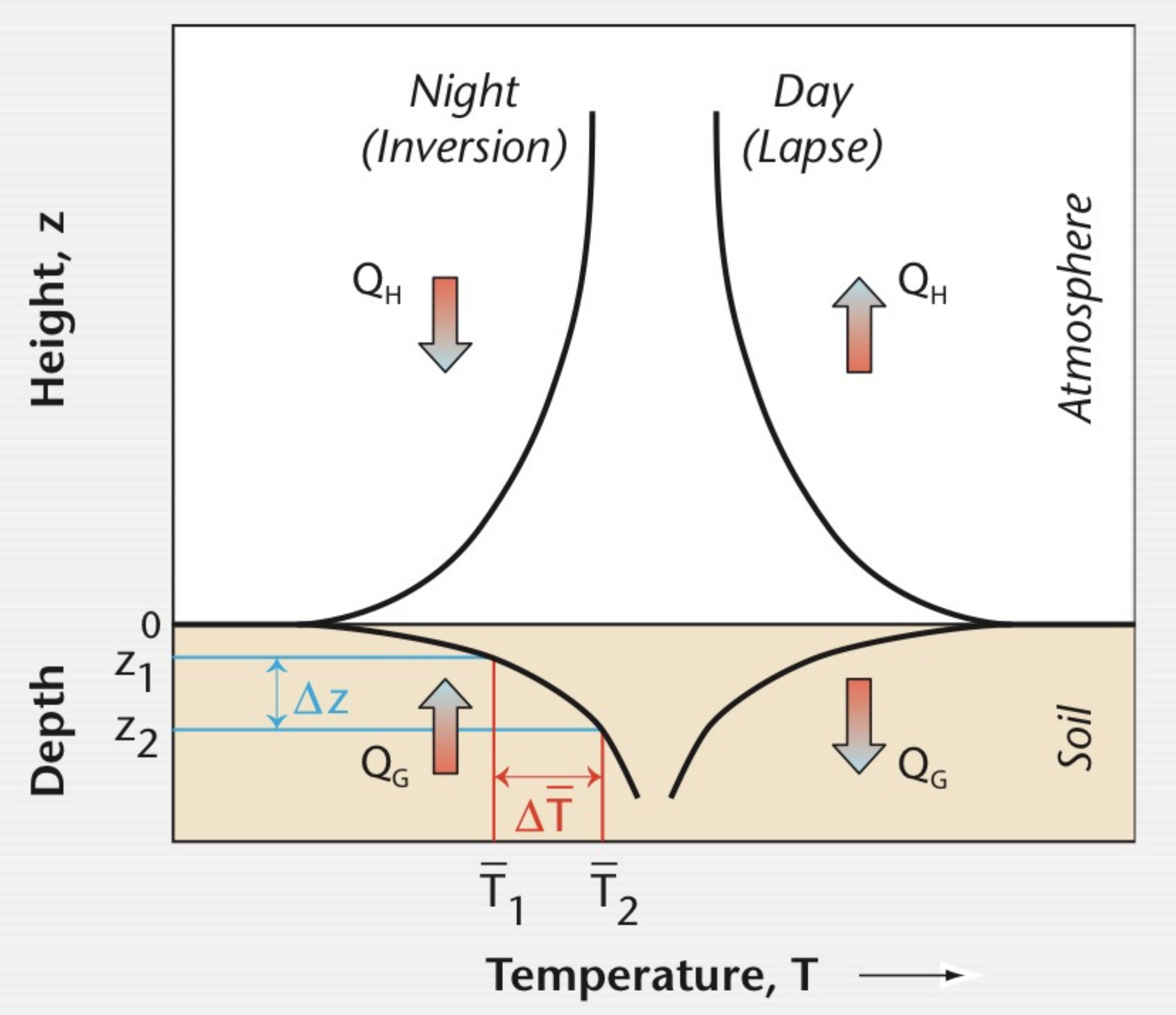
Data from a tidal marsh in Delta – questions for discussion



- How does G vary between daytime and nighttime?
- Can you say anything about the spatial variability at the site?

https://ibis.geog.ubc.ca/~micromet/data/DeltaSaltMarsh.html#

Soil temperatures and heat flux profiles



The heat conduction equation.

$$\frac{\partial Q_G}{\partial z} = C_s \frac{\partial T}{\partial t}$$

Second law of thermodynamics (Eq. 1)

$$Q_G = -k \frac{\partial T}{\partial z}$$

Fourier's law of heat conduction (Eq. 2)

$$\frac{\partial T}{\partial t} = -\frac{k}{C_s} \frac{\partial}{\partial z}\left(\frac{\partial T}{\partial z}\right) = -\kappa_s \frac{\partial^2 T}{\partial z^2}$$

Solving the heat conduction equation.

$$\boxed{\frac{\partial T}{\partial t}} = \boxed{-\kappa_s} \frac{\partial^2 T}{\partial z^2}$$

This is a **partial differential equation**, for which **no general analytic solution** exists. There are solutions for selected boundary conditions (see Lecture 12).

Often we approximate the equation with **multi-level** soil models (e.g. in numerical weather forecasting or climate models).

Alternatively one can neglect the variation with depth and look at **heat sharing** between air and soil.

Thermal admittance

The **thermal admittance** μ is the ability of a surface or system to accept or release heat following a change in (soil) heat flux $\partial Q_G/\partial t$.

$$\boxed{\mu} = \sqrt{\boxed{k}\boxed{C}} \quad \star$$

Units are: J m^{-2} K^{-1} s$^{-1/2}$.

Thermal admittance

For a given change in ΔQ_G the change in surface temperature ΔT_s is inversely proportional to μ.

$$\frac{\partial T_s}{\partial t} \propto \frac{1}{\mu} \frac{\partial Q_G}{\partial t}$$

Test your knowledge

Surface temperature fluctuations for soils with **low thermal** admittance are:

a) Small

b) Large

$$\frac{\partial T_s}{\partial t} \propto \frac{1}{\mu} \frac{\partial Q_G}{\partial t}$$

Thermal admittance of soils

High μ environments - relatively small T range (wet areas, clay, bare rock)

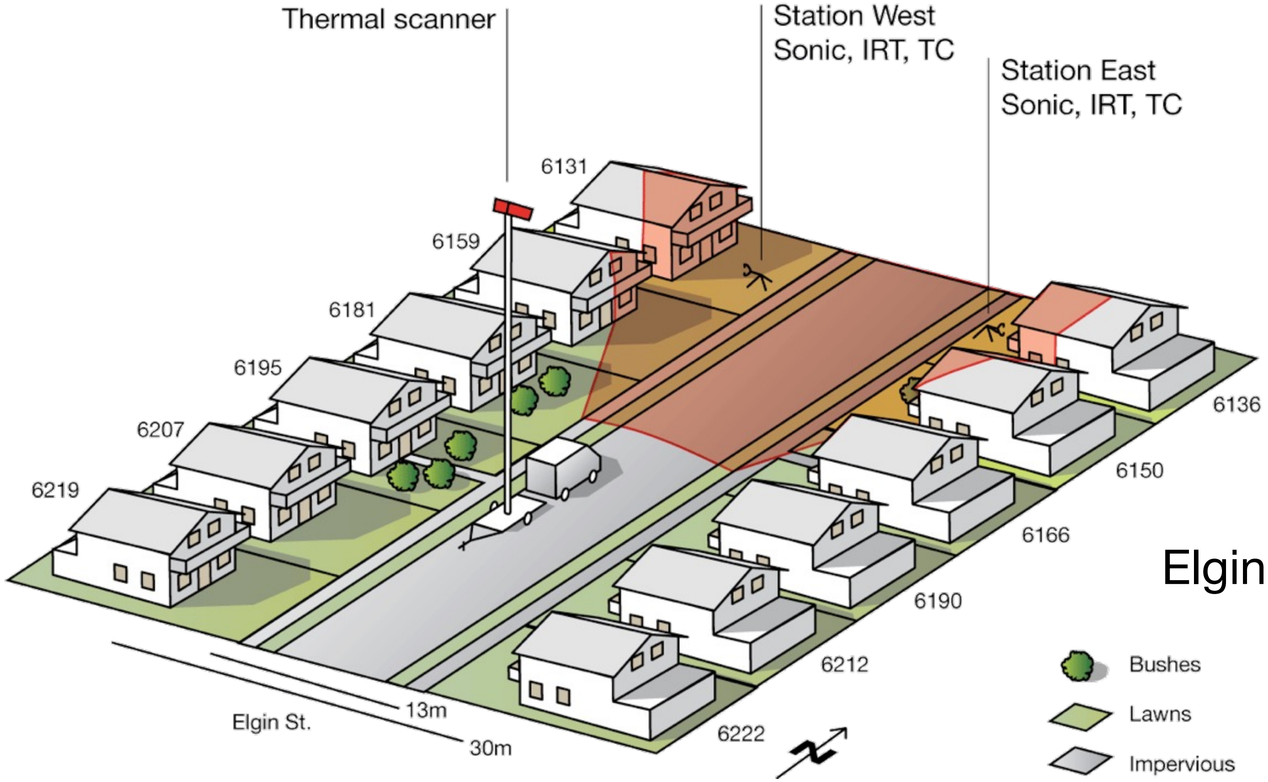
Low μ environments - large daily T range (dry areas, sandy, organic soils, vegetation cover, snow)

All forcing being equal soil with high μ will have a smaller surface temperature wave. Low μ produces a larger surface temperature wave.

Snow melt in the Peace River area - soil saturation results in high soil heat capacity and slow soil warming (Photo: A. Black, UBC)

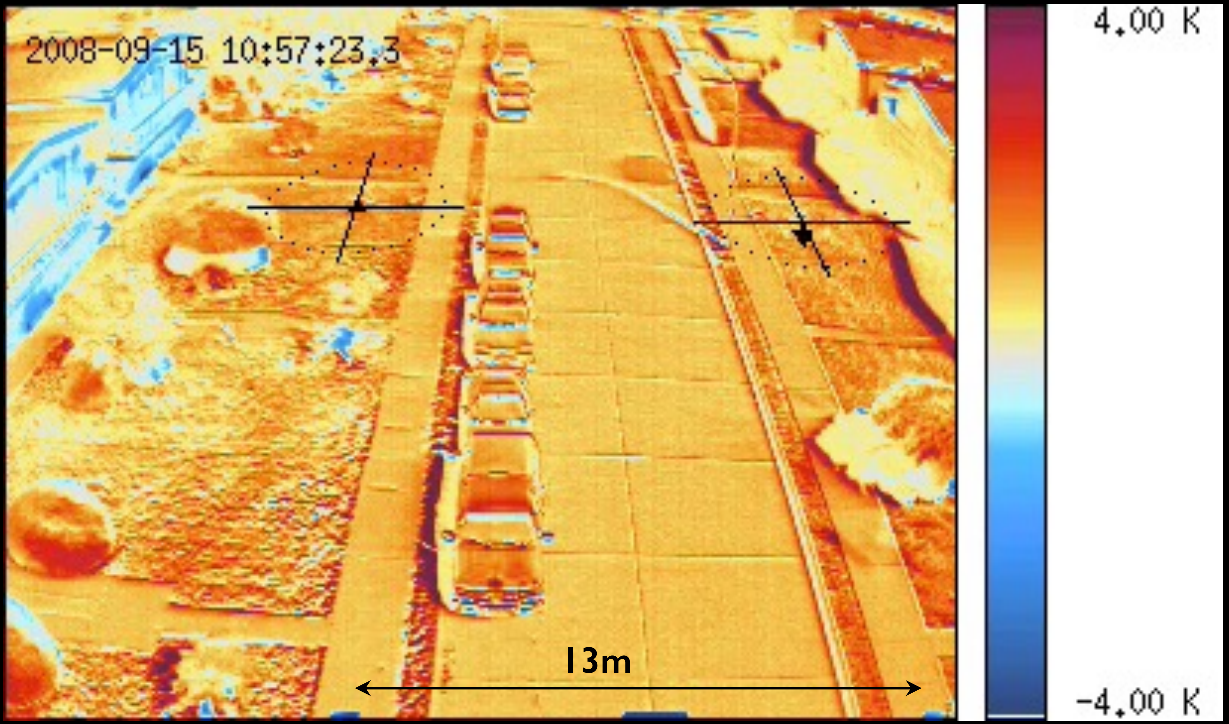
Visualizing the effect of thermal admittance

Thermal admittance controls how quickly the surface accepts heat, i.e. how much temperature fluctuation a surface is experiencing.



Elgin St. in Vancouver

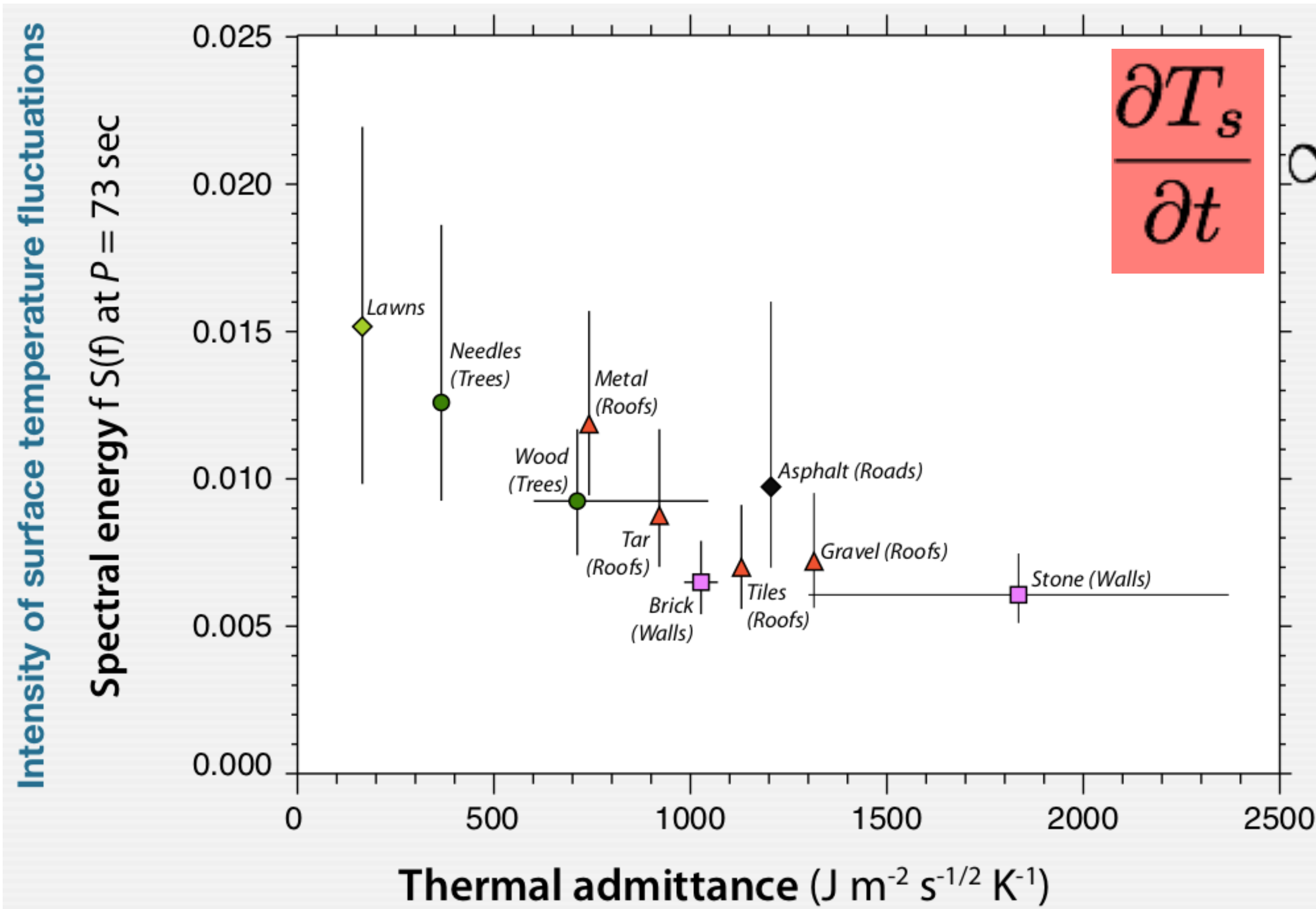
Fast thermal infrared thermography of temperature fluctuations at the surface with wind vectors overlaid



Approximate visible field of view

A.. Christen, J. A., Voogt (2010): 'Inferring turbulent exchange processes in an urban street canyon from high-frequency thermography', *19th Symposium on Boundary Layers and Turbulence*, Keystone CO, USA.

Surface temperature fluctuations correlate with thermal admittance.



$$\frac{\partial T_s}{\partial t} \propto \frac{1}{\mu}\frac{\partial Q_G}{\partial t}$$

Christen A., Meier, F. Scherer D. 'High-frequency fluctuations of surface temperatures in an urban environment', *Theoretical and Applied Climatology* (2012)

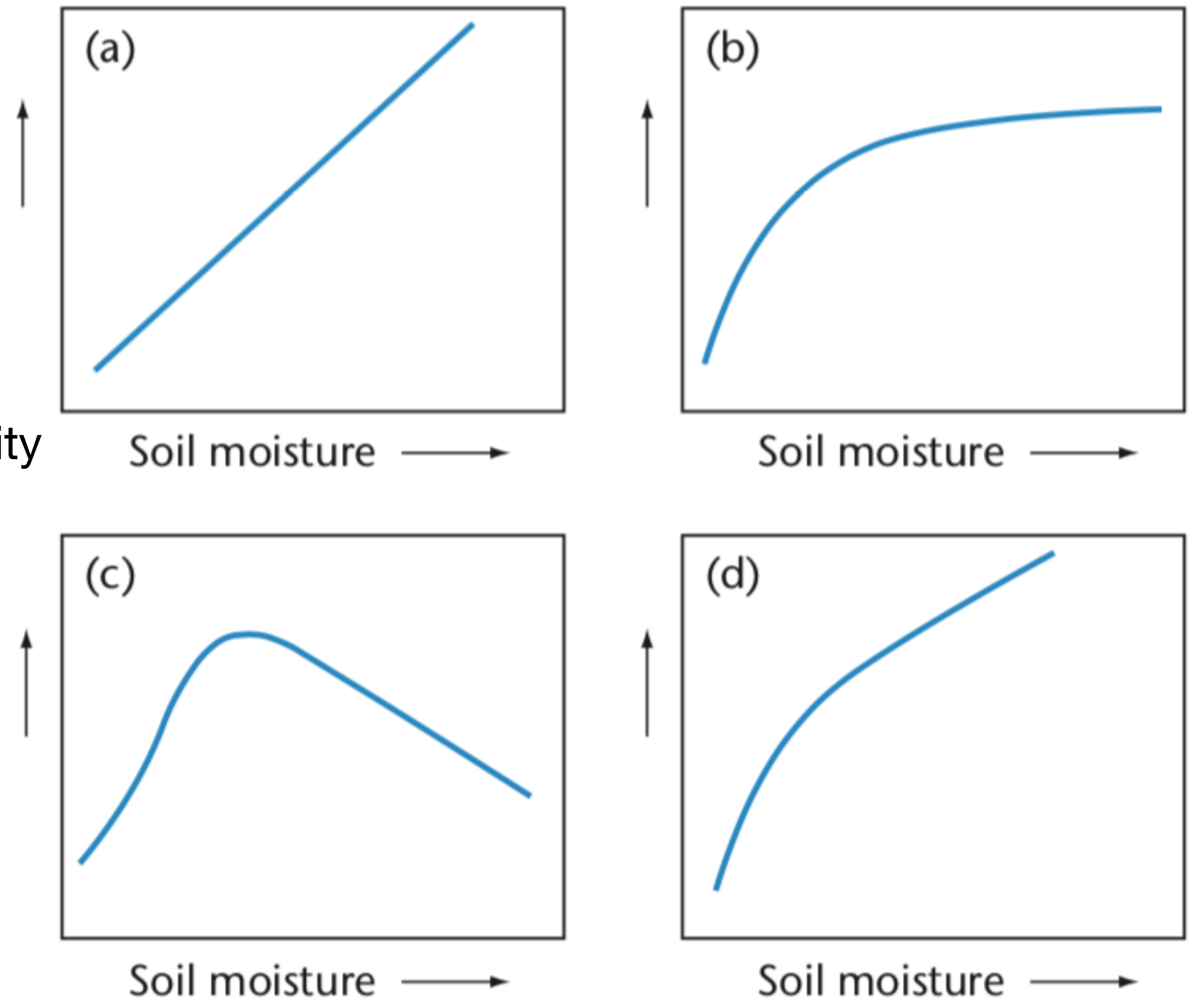
Effect of soil water content on thermal properties (top hat activity)

Thermal admittance

Heat capacity

Thermal diffusivity

Thermal conductivity



Heat sharing

Refers to partitioning of total sensible heat flux (heat for warming $Q^* - Q_E$) between soil (Q_G) and atmosphere (Q_H).

Both share in accepting heat during daytime (warming) and share in releasing heat at night (cooling):

$$\frac{Q_G}{Q_H} = \frac{\mu_s}{\mu_a}$$

where μ_a is the atmospheric thermal admittance, which increases with wind speed and free convection as result of surface heating.

Take home points

- Soils heat fluxes can be measured using **heat flux plates** inserted into some depth below the surface. Need to be corrected for divergence of heat above measurement level.
- The general **equation for heat conduction** describes how soil temperatures change over time and with depth. It has no general analytical solution, but we can approximate it or solve for selected boundary conditions.
- **Thermal admittance** μ can be used to describe the heat sharing between soil and atmosphere.