3/03/2011

Began class reviewing Darcy's law,

$$J_w = -K(\psi_m)\frac{d\psi_m}{dz} + gK(\psi_m)$$

using figures from the text (Fig and from Soil Science Simplified (SSS). Note mistakes in vertical axes in the SSS figure.

Guest lecture, Dr. Robert Horton

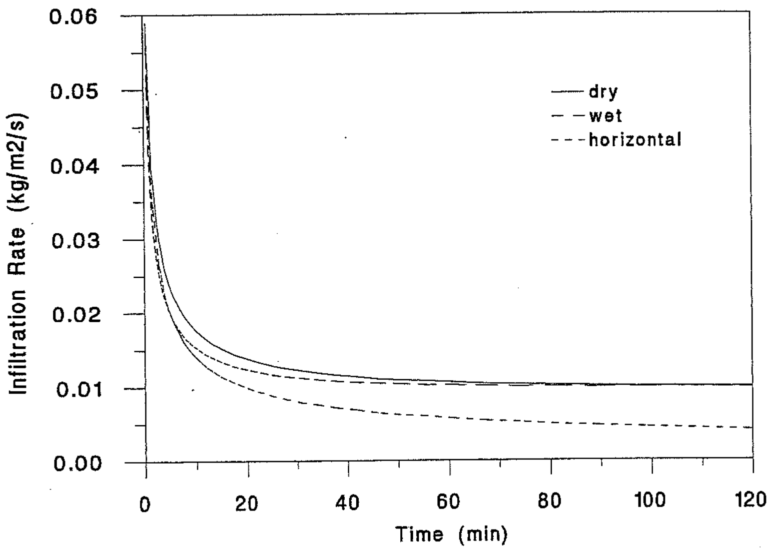
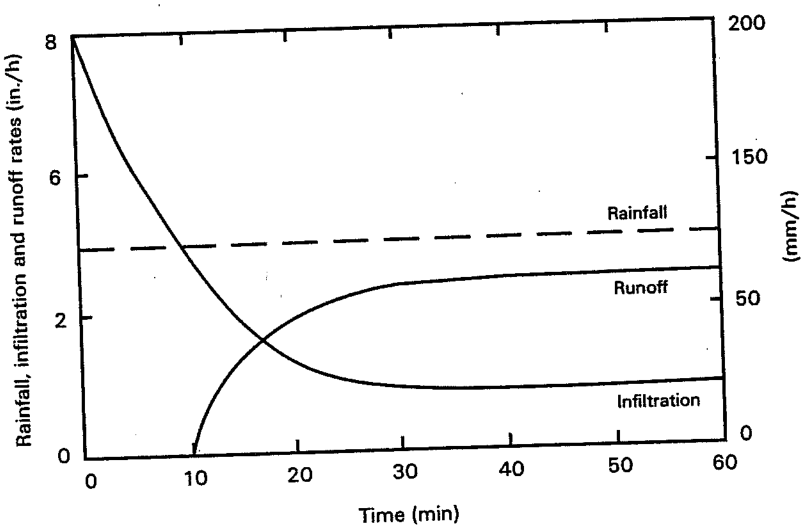
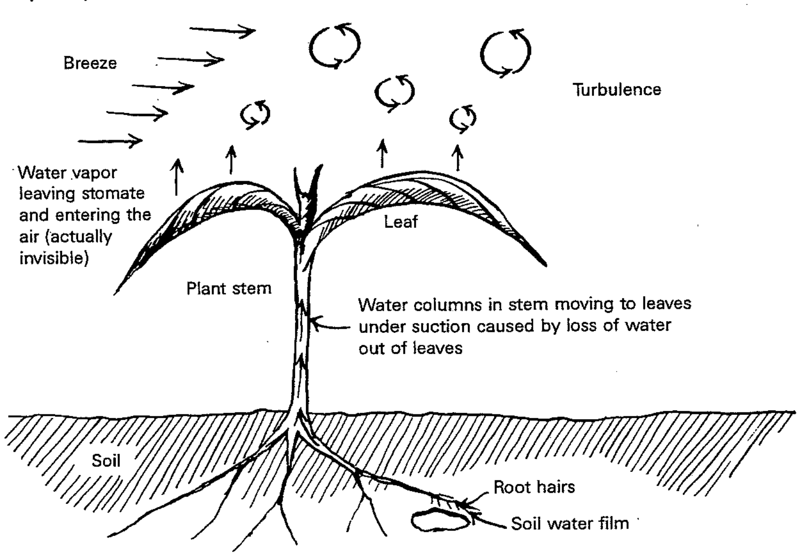


FIGURE 9.1. Vertical infiltration rate for water into initially dry and wet soil, and horizontal infiltration into dry soil.

6.4. Runoff and infiltration for a 1.5-inch (38-mm) rainfall in 1 hour. The infiltration rate decreases as the soil wets until runoff begins after 10 minutes. Late in the storm, the runoff and infiltration rates are steady. Runoff would have begun later and been less if the soil had a higher infiltration rate.



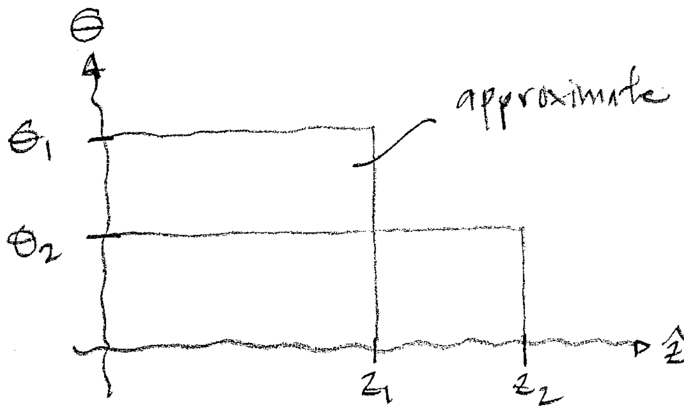
6.12. Water moves into the roots and through the plant primarily by capillary action.



Soil Water Redistribution

After a rain or irrigation event, the soil water profile continues to change as the water is redistributed.

See Fig 9.3



approximate by rectangle,

area = depth of wetted profile × average water content

$= D \times \Theta = d$

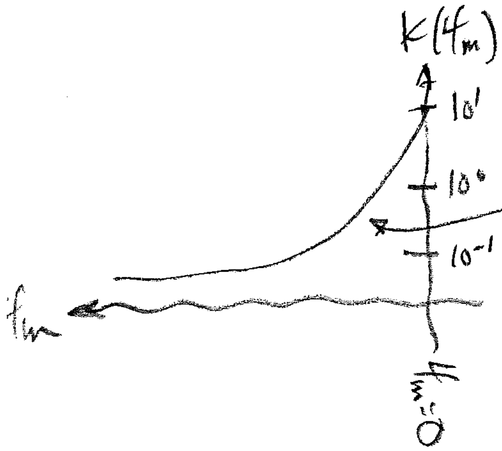
= equivalent depth of water = constant

The rectangle of wet soil changes shape but the area remains approximately constant as water is redistributed. — ask why this is an approximation

When redistribution slows down significantly, the water content/potential of the soil has reached field capacity.

$$J_w\Big|_{\psi_m=\psi_{m,fc}} = \text{water flux at field capacity} \doteq -\underbrace{K(\psi_{m,fc})}_{\text{small}}\underbrace{\frac{d\psi_m}{dz}}_{\text{small}} + g\underbrace{K(\psi_{m,fc})}_{\text{small}}$$

$\psi_{m,fc}$ = matric potential at field capacity $\simeq -30 \frac{J}{kg}$



rapid decrease in K(ψm)

Note that $J_w|_{\psi_{m,fc}} \neq 0$ since there will always be some water movement. Think of soil water as being held by a leaky bucket. Hole in bucket through which water leaks changes as soil water content changes, and gets smaller rapidly near $\psi_{m,fc}$.
— large at high water content

JJK
3/03/09

Soil Water Evaporation

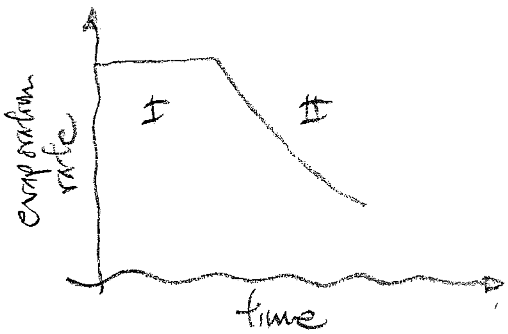
Optional 3/03/2011

vapor pressure of atmosphere compared to vapor pressure at soil surface

Evaporation occurs in 2 stages.

I) steady rate of evaporation, determined by evaporative demand of the atmosphere. Think of diffusion of water vapor at soil surface to atmosphere

II) falling rate of evaporation once the soil surface has dried, rate of evaporation now depends on ability of soil to move water to the surface.



$E_{cum II}$ = cumulative evaporation in stage II

$$= C\sqrt{t - t_I}$$

$C = f(\text{soil type})$, t_I = time stage I ends

recall $I_w \sim \sqrt{t}$

Transpiration and Plant Water Uptake

Water moves from the soil to plant roots, through the xylem, to the leaves, and then evaporates from the cavities of the stomata.

$$U_p^* = \frac{\text{dimensionless}}{\text{quantity}} = \frac{U_p}{E_{p,max}} = \frac{\text{max potential water uptake by plant}}{\text{max possible transpiration}} = 1 - (1 + 1.3 A_w)^{-b}$$

determined mostly by atmosphere

$$A_w = \begin{matrix}\text{available}\\\text{water}\\\text{fraction}\end{matrix} = \frac{\theta - \theta_{pwp}}{\theta_{fc} - \theta_{pwp}}$$

θ_{pwp} = vol water content @ permanent wilting point

θ_{fc} = vol water content @ field capacity

$[U_p] = [E_{p,max}] = \text{mm} \cdot \text{day}^{-1}$

If ambient water vapor pressure is such that a plant is transpiring at $E_{p,max}$, then

$$U_p = \left[1 - (1 + 1.3 A_w)^{-b}\right] E_{p,max}$$

see Fig 9.9

spatial and temporal variability

Heat Pulse Measurements to determine:

soil thermal properties
soil water content
infiltrating liquid water flux
sensible heat flux in soil
latent heat flux (vaporization or fusion)
upward liquid water flux

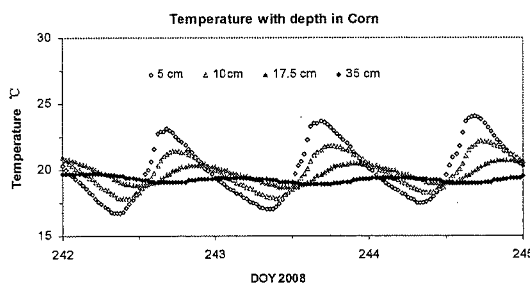
Agron 405/505

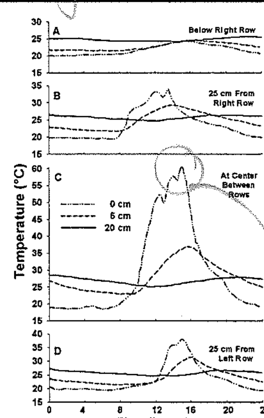
Soil heat and water dynamics

- Impact biological, chemical, and physical, processes
- Modeling coupled heat and water dynamics is difficult and requires many hard to measure parameters
- Measuring in situ coupled heat and water dynamics has improved recently

Research: improving models of soil heat and water flow
Needs: good measurements!

Use measurements to improve our understanding of physical processes

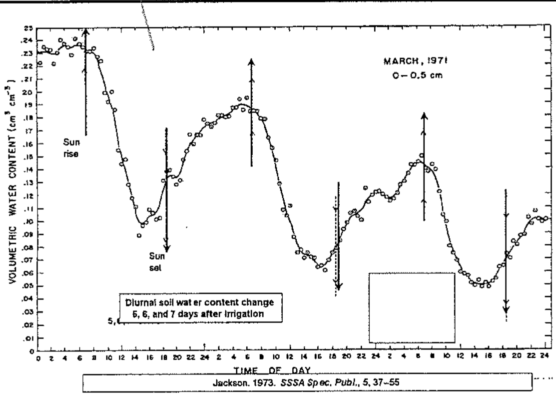




from Dr. Horton's thesis, New Mexico

hot!

Ray Jackson data from Arizona



occur simultaneously

Coupled Heat and Water Transfer

Thermal gradients cause water to move in unsaturated soil.

When water moves in soil, it carries heat.

Because heat transfer and water movement affect one another they are coupled.

Determining of soil thermal properties by heat pulse sensor

Soil thermal diffusivity α (J/m³°C)

$$\alpha = \frac{r^2}{4}\left[\frac{1}{(t_m - t_0)} - \frac{1}{t_m}\right] \Big/ \ln\left[\frac{t_m}{(t_m - t_0)}\right]$$

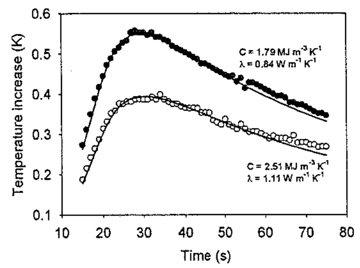
Soil heat capacity C (J/m³°C):

$$C = \rho c = \frac{t_0 q'}{e \pi r^2 \Delta T_m}$$

Soil thermal conductivity λ (W/m°C):

$$\lambda = \alpha C$$

Example of heat pulse data



By fitting a heat transfer model to the heat pulse data we determine the soil thermal properties.

$C \rightarrow$ volumetric heat capacity ($\rho_s C_s$)
$\lambda \rightarrow$ thermal conductivity (k)
$\alpha \rightarrow$ thermal diffusivity (K)

Thermal properties

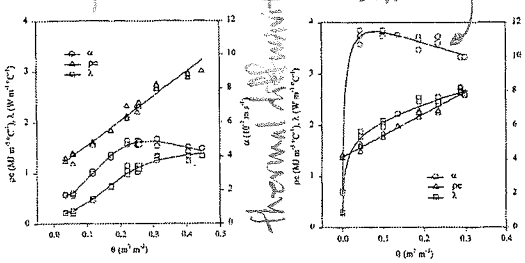
Volumetric heat capacity

clay loam

thermal diffusivity

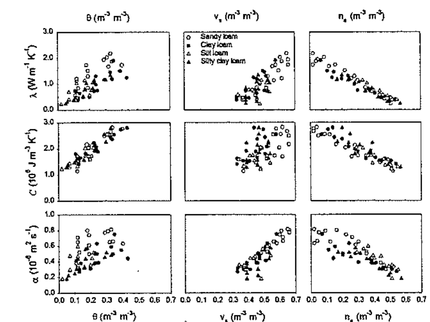
sand

because of quartz



water content

Soil thermal properties



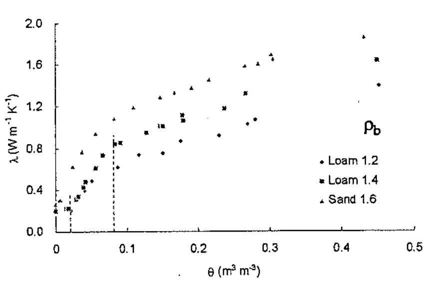
volume fraction of solids

volume fraction of air

relate thermal properties to air content

bulk density, kg/m³

Influences of soil texture, ρ_b and θ on λ



Calculation of Volumetric Heat Capacity

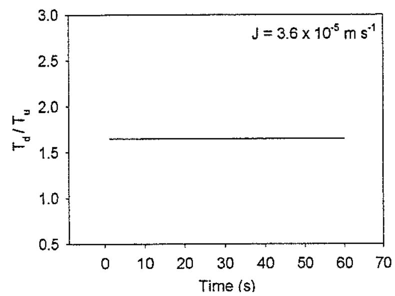
$$\rho c = \rho_b c_s + \theta \rho_w c_w$$

(ignore gas phase)

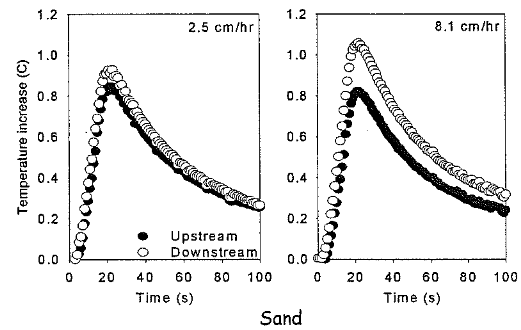
This equation can be used to estimate soil ρ_b or θ with the heat-pulse technique.

If ρ_b doesn't change (soil doesn't shrink or swell) θ can be inferred from $\rho_s C_s$.

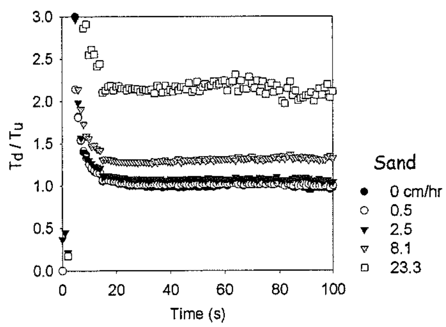
Temperature ratio is constant



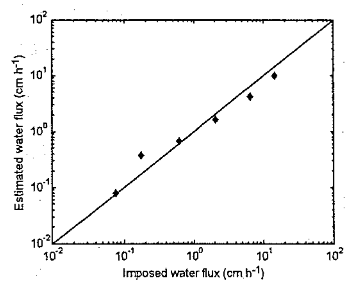
Measured heat pulse signals



Heat pulse signals converted to T_d/T_u



Heat pulse flux estimates versus imposed unsaturated fluxes



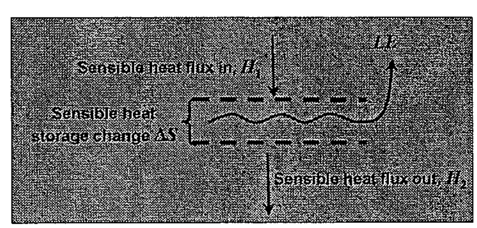
A Heat Pulse Technique for Estimating Soil Water Evaporation

$$H = -k \frac{dT}{dz}$$

Basic theory of HP method:

Sensible heat balance provides a means to determine latent heat (*LE*) used for evaporation.

$$LE = (H_1 - H_2) - \Delta S \quad \begin{cases} =0 & \text{no evaporation} \\ >0 & \text{evaporation} \\ <0 & \text{condensation} \end{cases}$$



Latent Heat in Soil Heat Flux Measurements

Which temperature should we be using to estimate the evaporation? Surface? Subsurface?

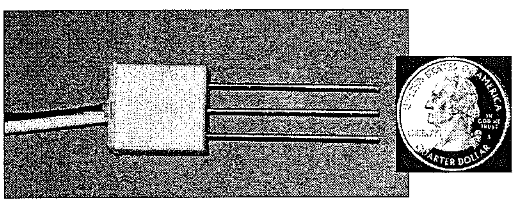
Better Energy Balance Closure

When the latent heat flux (*LE*) includes evaporation from soil, the depth at which we measure soil heat flux (*G*) is critical to accurately representing the surface energy balance.

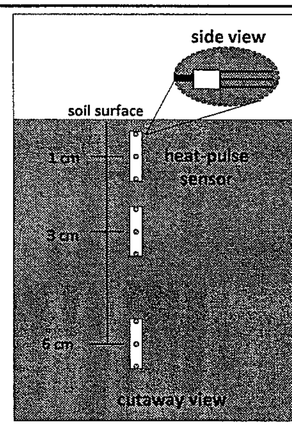
Objective: Characterize variations in *G* with depth near the soil surface.

Materials and Methods

Soil heat flux (*G*) measured via heat-pulse sensors installed at 3 depths: 1, 3, and 6 cm

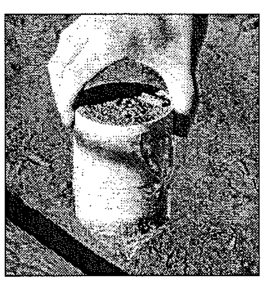


$$G = -\lambda(\Delta T/\Delta z)$$

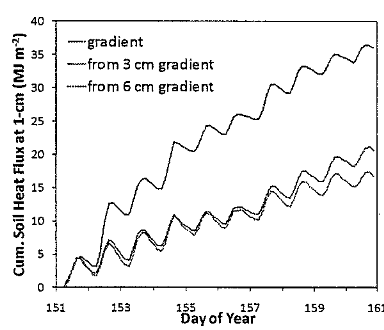


Materials and Methods

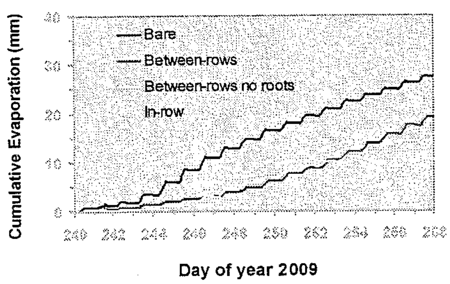
Evaporation (*LE*) determined via microlysimeters (per 24 h)

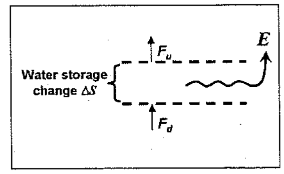


Cumulative Soil Heat Flux at 1-cm Depth



Cumulative soil water evaporation at 3-mm soil depth

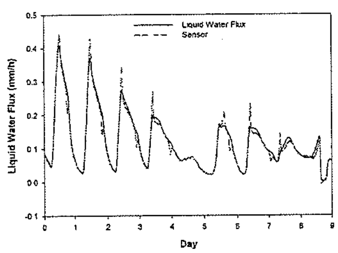




$$\Delta E = (F_d - F_u) + \Delta S$$

For a soil layer, Δ E is the evaporation rate (cm/h), F_t and F_b are the liquid water flux (cm/h) at top and bottom boundaries, and Δ S is the change in water storage (cm/h).

Liquid water flux at the 7.5 mm soil depth from the model simulation.



Conclusions

The heat pulse method isable to provide a wide range of soil heat and water measurements.

This is an important time period to advance coupled heat and water experiments and models.

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