Disturbed micrometeorological flows – example – “local advection”

Horizontal gradients of mean properties
($\overline{u}$, $\overline{T}$, $\overline{u'w'}$, $\overline{w'T'}$ etc.) in the atmospheric surface layer may be generated

- by inhomogeneity in the surface boundary conditions** – inhomogeneity in surface properties and fluxes e.g. ΔQ_{H0}, ΔQ_{E0}, Δz_0,… due to varying soil moisture, surface elevation/cover ,…

- by purely aerodynamic disturbances (windbreaks, hills, buildings,…)

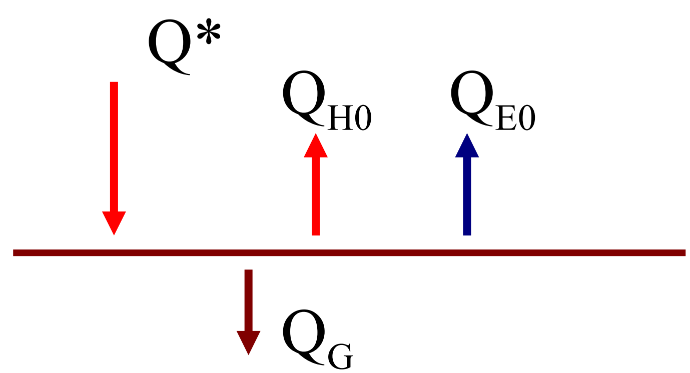
- by a combination of these types of influences

Note: the flow need not be disturbed at the boundary in order to be inhomogeneous

In this real world, irrigated fields adjoin deserts, reservoirs are of finite extent, dry lands exist beside seas, and cornfields beside close-grazed pasture. It is not surprising, then, that many important problems of micrometeorology require that we take cognizance of *advection*. This we define as the exchange of energy, moisture, or momentum due to horizontal heterogeneity. One symptom of the presence of advection is that vertical mean profiles of (potential) temperature, specific humidity, and wind speed are non-equilibrium profiles, even under conditions steady in time.

(Philip*, 1959, The Theory of Local Advection, J. Meteorol. Vol. 16)

Our subject material so far has addressed flows that are horizontally-homogeneous. We now progress to consider examples of “disturbed flows,” with emphasis on their modelling – which, to the extent that it is accurate, is indicative of our ability to generalize from specific instances of disturbed flows…



**The surface energy budget

* J.R. Philip was chief of CSIRO’s “Pye Lab” (Canberra), and provided ingenious analytical solutions to the mass conservation equation applied to soil moisture and soil solute flows – solutions vitally useful in the pre-computer era

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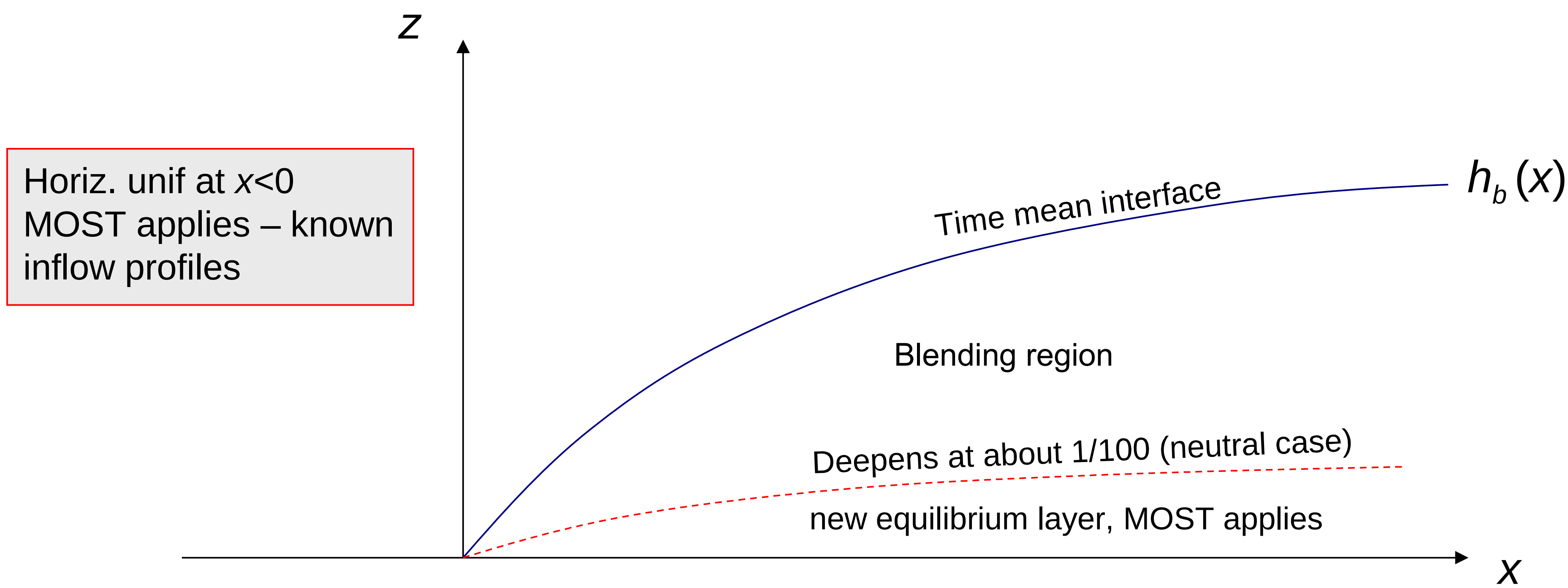
JDW vers. 20 Nov. 2012

The Paradigm of the Internal Boundary Layer



Paradigm:

$$\frac{\partial h_b}{\partial x} \propto \frac{\sigma_w}{\bar{u}(\bar{z})} = \frac{\sigma_w}{\bar{u}(\alpha h_b(x))} \xrightarrow{\text{Neutral case}} \frac{h_b}{z_0}\left[\ln\frac{h_b}{z_0} - 1\right] + 1 = A\,\frac{x}{z_0}$$

Weakness: this approach *neglects disturbance to pressure* and considers the disturbance propogates like a passive tracer gas

$$\frac{1}{\rho_R}\nabla^2\bar{p} = -\frac{\partial \bar{u}_i}{\partial x_j}\frac{\partial \bar{u}_j}{\partial x_i} - \overline{\frac{\partial u_i'}{\partial x_j}\frac{\partial u_j'}{\partial x_i}} - \frac{g}{T_R}\frac{\partial \bar{T}}{\partial z}$$

This Poisson eqn easily derivable from the Reynolds eqns. Solution for mean pressure at point **P** responds to r.h.s. over *all* positions **r**, weighted as $|\mathbf{P} - \mathbf{r}|^{-2}$. This implies a disturbance has upstream influence

Useful reading: Garratt pp104 -108
(Sec. 4.5 up to eq. 4.30)

An early expt. & test of Philip's analytical theory of local advection – wind blows off tarmac onto short grass

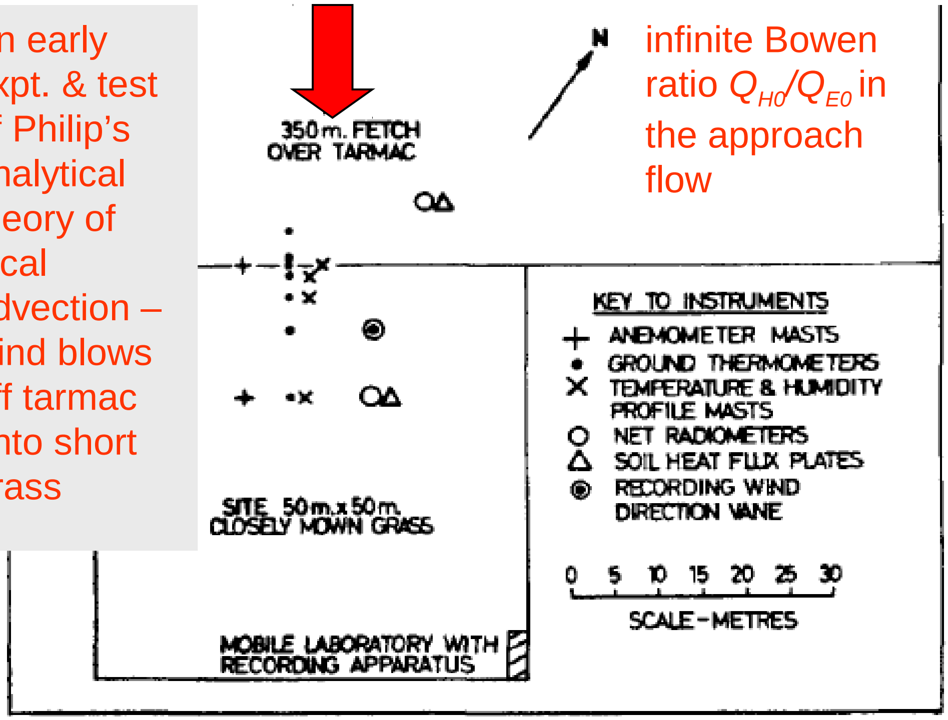
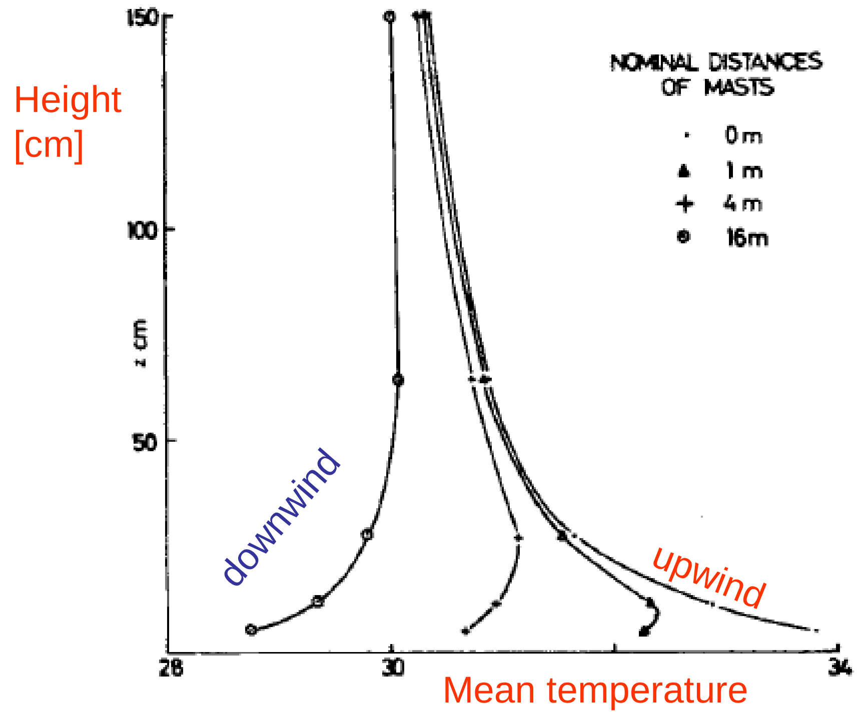


Figure 2. The experimental site, together with the disposition of instruments.



The horizontal transport of heat and moisture – a micrometeorological study

By N. E. RIDER*, J. R. PHILIP and E. F. BRADLEY
C.S.I.R.O., Division of Plant Industry, Canberra, Australia

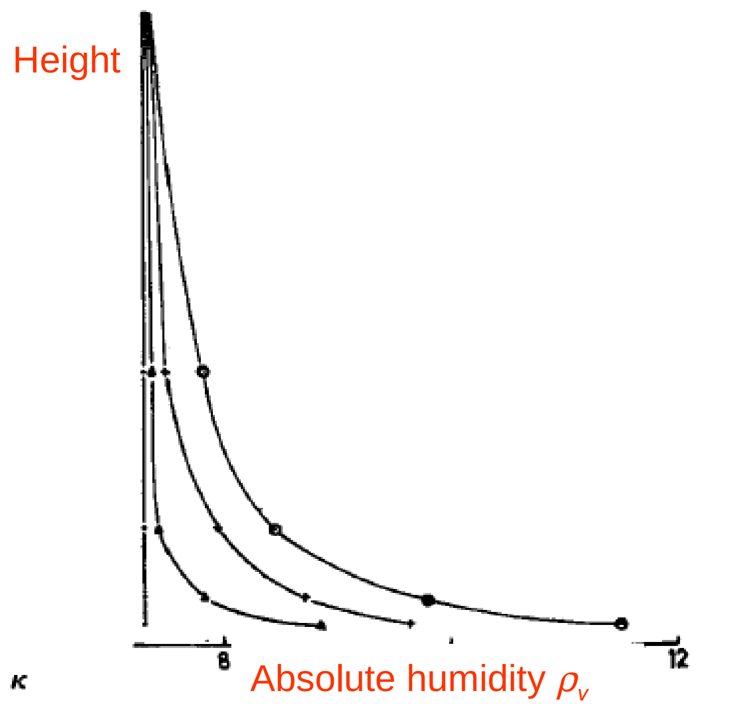
QJRMS Vol. 89, 1963

$$u\frac{\partial T}{\partial x} = \frac{\partial}{\partial z}\left[K_T \frac{\partial T}{\partial z}\right] \quad (1)$$

$$u\frac{\partial e}{\partial x} = \frac{\partial}{\partial z}\left[K_e \frac{\partial e}{\partial z}\right]; \quad (2)$$

Interpretive paradigm = eddy diffusion; steady state; divergences of sensible and latent heat flux vectors vanish; imposed (unvarying) power law profiles of wind and diffusivity

$u = u_1 z^m$; $K_T = K_e = \kappa z^n$. (m and n independent constants with $n < 1$, κ a constant)



The boundary conditions clearly must include the profiles of temperature and humidity at the leading edge of the area of interest. That is, we have the conditions :

$$x = 0,\ z \geqslant 0; \quad T = T(0, z), \quad e = e(0, z). \qquad (3)$$

The energy balance at the surface, $z = 0$, $x \geqslant 0$,

$$(1 - r)\, R_s + R_a - \epsilon\sigma T_0^4 + Q = A + L\rho_w E, \qquad (4)$$

Incoming terrestrial (longwave) radiation

Outgoing terrestrial (longwave) radiation

invariably provides a further condition. Here R_s and R_a are the flux densities of atmospheric and short wave radiation at the surface; r is the reflection coefficient of the surface for short wave radiation; ϵ, the surface emissivity; σ, the Stefan-Boltzmann constant; T_0, the surface temperature; Q, the soil heat flux at the surface; A, the sensible heat exchange between the surface and air; L, the latent heat of evaporation of water; ρ_w, the density of liquid water; and E, the rate of evaporation. We notice that in Eq. (4)

$$A = -c\rho\,(K_T\, \partial T/\partial z)_0; \quad \rho_w E = -(K_e\, \partial e/\partial z)_0. \qquad (5)$$

where c and ρ are the specific heat of air at constant pressure and the air density respectively.

One other condition at the surface is needed to complete the system, and this is provided by the availability of water for evaporation at the surface. When water is freely available there (the case we are mostly concerned with here) the condition takes the form :

$$z = 0,\ x > 0; \quad e_0 = e_s(T_0). \qquad (6)$$

Later authors** refined the treatment of the lower boundary condition; useful to reframe in terms of equivalent temperature and saturation deficit ($\rho c_p T_{eq}$ is total thermodyn. energy).

$$\overline{T}_{eq} = \overline{T} + \frac{\overline{e}}{\gamma}$$

γ the psychrometric constant – temperature if all latent heat converted to sensible

$$\overline{D} = e_{sat}(\overline{T}) - \overline{e}$$

e is vapour pressure

$$Q_H + Q_E = -\rho_R c_p K \frac{\partial \overline{T}_{eq}}{\partial z}$$

LHS constrained at gnd by sfc energy balance

**Raupach (1991; Vegetatio, Vol. 91) preceded by McNaughton

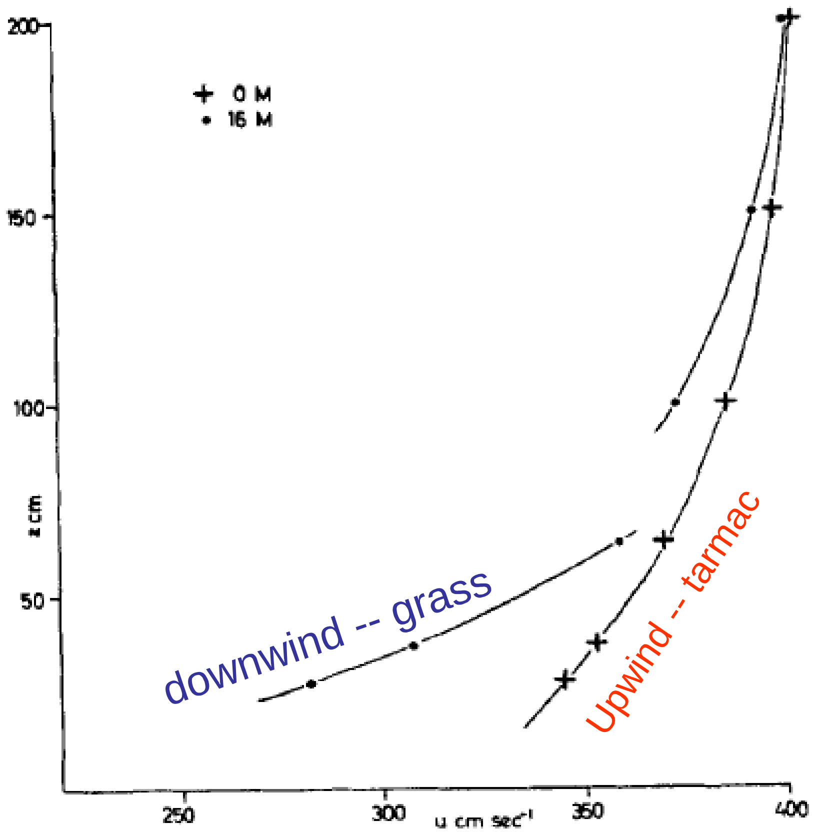


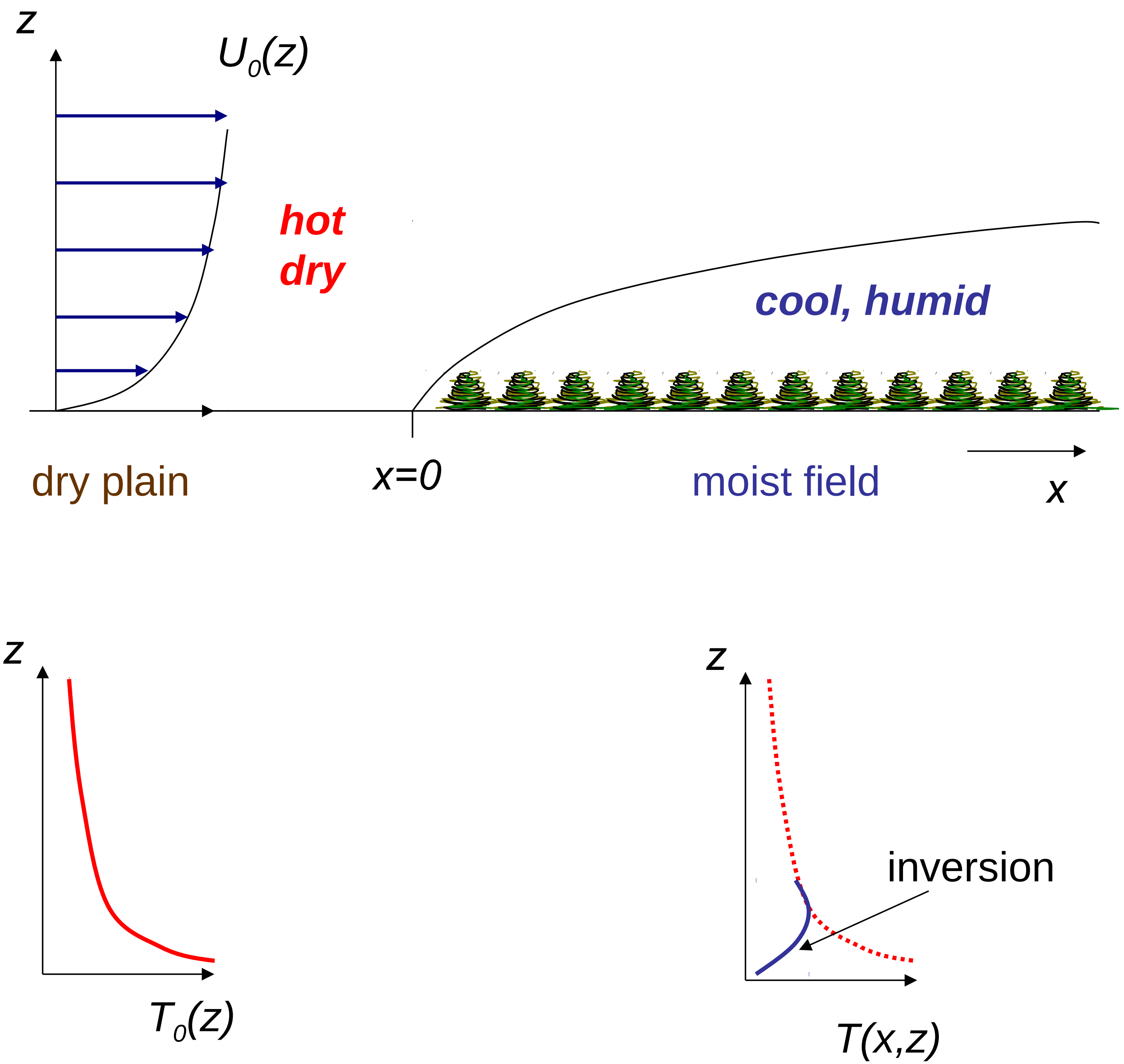
Figure 5. Typical example of the vertical profiles of wind speed at $x = 0$ and $x = 1600$ (observation No. 11).

Observed variation of the profile of the (advecting) mean wind speed implies one should account for conservation of momentum as well as heat and latent heat

Rider, Philip & Bradley had treated net radiation less soil heat flux as invariant with *x*, implying

$$Q_{H0} + Q_{E0} = \text{const.} = -\rho_R c_p \left[K \frac{\partial \overline{T}_{eq}}{\partial z} \right]_0$$

Local advection experiment (La Crau Valley, France; N.J. Bink, 1996. Ph.D. thesis, Wageningen Agric. Univ.)



Rao-Wyngaard-Coté 2nd-order closure model of local advection:

17 equations in 17 unknowns (symmetry along y-axis, ie. 2d implementation):

$$U,\ W,\ P,\ T,\ Q,\ \overline{u'^2},\ \overline{v'^2},\ \overline{w'^2},\ \overline{u'w'},\ \overline{u'T'},\ \overline{u'Q'},\ \overline{w'T'},\ \overline{w'Q'},$$

$$\overline{T'^2},\ \overline{q'^2},\ \overline{q'T'},\ \varepsilon$$

Very similar to other 2nd-order closures, e.g. Launder,Reece & Rodi

U-mtm:

$$\frac{\partial}{\partial x}\left(UU+\sigma_u^2\right)+\frac{\partial}{\partial z}\left(UW+\overline{u'w'}\right)=-\frac{1}{\rho_0}\frac{\partial P}{\partial x}$$

hor.flx. **vrt.flx.**

pressure disturbance (not allowed for in original RWC treatment)

hor.flx.

adv. **diff.**

σ_u^2:

$$\frac{\partial}{\partial x}\left(U\overline{u'^2}-a_t\tau\overline{u'^2}\frac{\partial\overline{u'^2}}{\partial x}\right)+\frac{\partial}{\partial z}\left(W\overline{u'^2}-a_t\tau\overline{w'^2}\frac{\partial\overline{u'^2}}{\partial z}\right)=-2\overline{u'^2}\frac{\partial U}{\partial x}-2\overline{u'w'}\frac{\partial U}{\partial z}-\frac{2}{3}\varepsilon-\frac{c_{11}}{\tau}\left(\overline{u'^2}-\frac{2}{3}k\right)$$

shear prod. **redistrib.**

$$+\frac{\partial}{\partial x}\left(a_t\tau\overline{u'w'}\frac{\partial\overline{u'^2}}{\partial z}\right)+\frac{\partial}{\partial z}\left(a_t\tau\overline{u'w'}\frac{\partial\overline{u'^2}}{\partial x}\right)$$

effective diffusivity

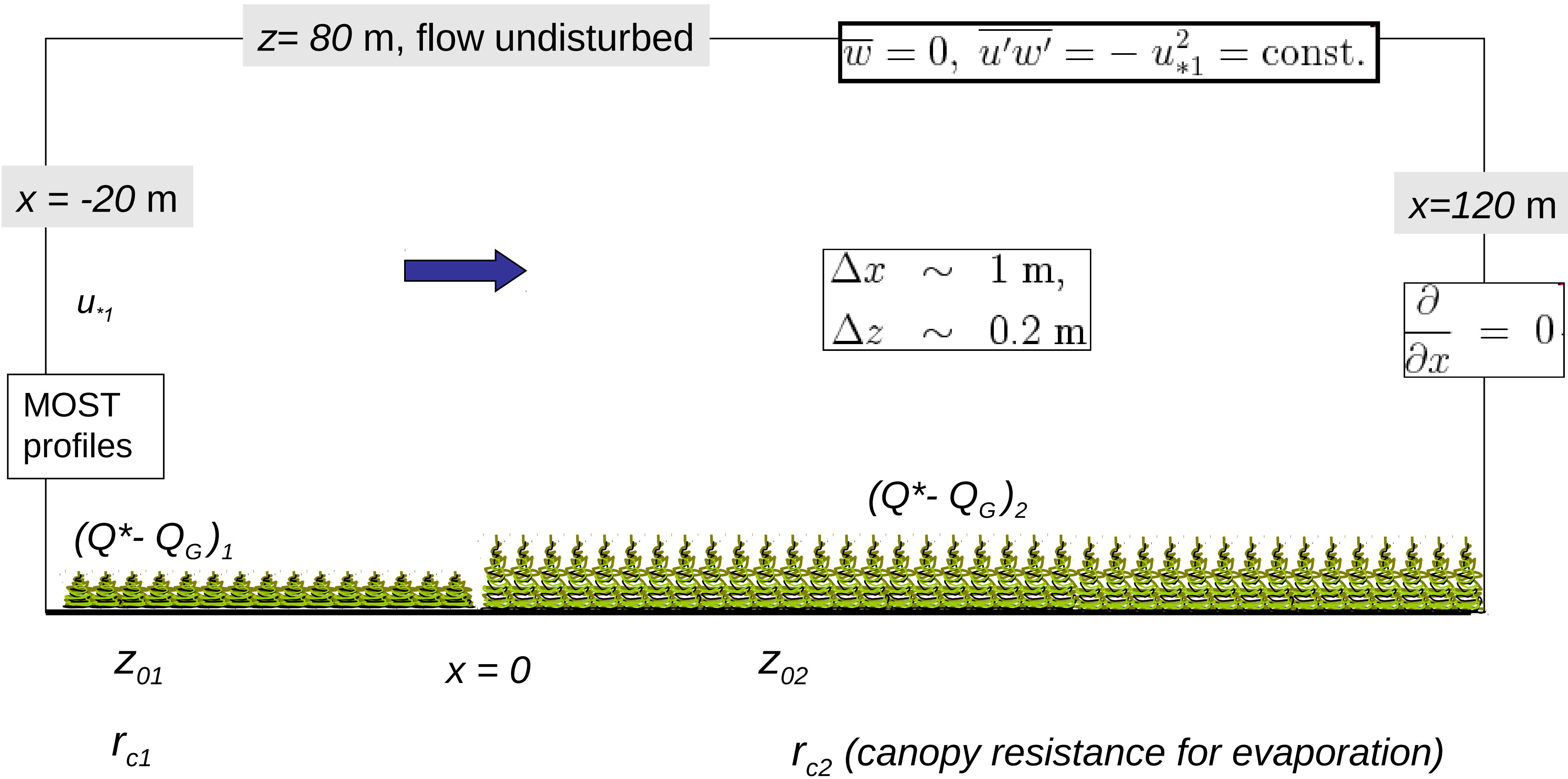
(here a_t is a closure constant)

$$\tau=\frac{2k}{\epsilon}$$

a turbulence time scale

The closure constants are **not free** – they are constrained by forcing the model to reduce to an exact model of the ideal NSL

Computational domain and boundary-conditions for application** of Rao-Wyngaard-Cote 2nd-order closure model to La Crau experiment:



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Micro-meteorological methods for estimating surface exchange with a disturbed windflow

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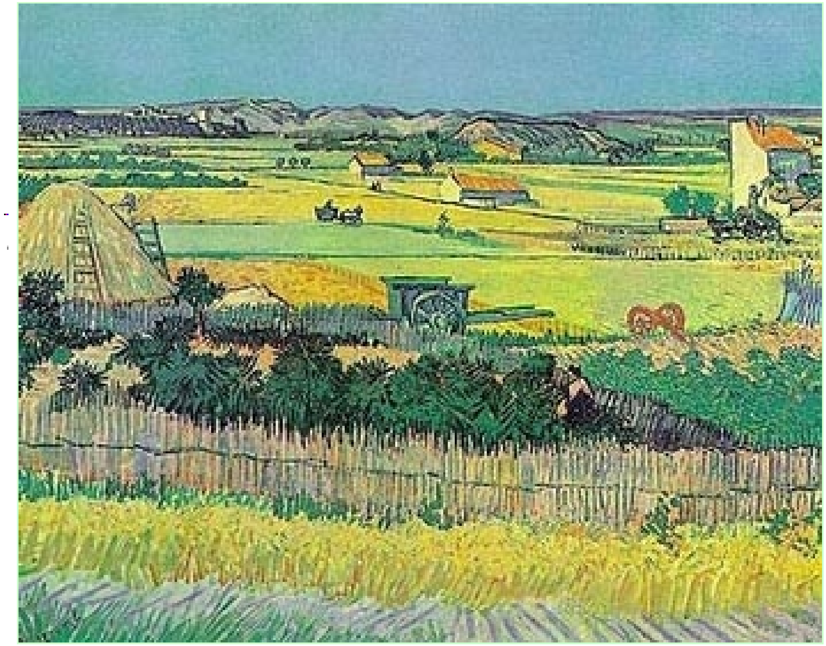
La Crau run 42 – specification of controlling boundary conditions



Vincent Van Gogh: "Harvest at La Crau"

$$T_{*1} \sim \frac{-362}{1 \times 1000 \times 0.63} \sim -0.6^{\circ} K$$

$$\begin{aligned}
T_1(3.05\text{m}) &= 24.08\ \text{K} \\
Q_1(3.05\text{m}) &= 6.6\ \text{g}\,\text{kg}^{-1} \\
u_{*1} &= 0.63\ \text{m}\,\text{s}^{-1} \\
z_{01} &= 0.01\ \text{m} \\
Q_{*1} - Q_{G1} &= 434\ \text{W}\,\text{m}^{-2} \\
Q_{H1} &= 362\ \text{W}\,\text{m}^{-2} \\
z_{02} &= 0.07\ \text{m} \\
Q_{*2} - Q_{G2} &= 500\ \text{W}\,\text{m}^{-2} \\
r_{c2} &= 47\ \text{s}\,\text{m}^{-1}
\end{aligned}$$

Surface treated as a "big leaf" and coupled to model atmosphere's lowest plane of gridpoints (at $z = z_0 + \Delta z \sim 0.2$ m) using the Penman-Monteith evapotranspiration eqn

$$Q_{E0} \equiv \lambda E_0 = \frac{\epsilon_{sa}}{\epsilon_{sa} + r_v / r_h} \left[Q^* - Q_G \right] + \frac{\rho \lambda D_a / r_h}{\epsilon_{sa} + r_v / r_h}$$

"Canopy resistance" r_c is the excess resistance for vapour loss, such that $r_v = r_h + r_c$

• λ the latent heat of vapourization; ε_{sa} ratio of the slope of the sat'n vapour pressure curve to the psychrometric constant; D_a the saturation deficit at the surface, varying with x

Aside on bulk transfer resistances

e.g. let r_h be the transfer resistance for heat between levels $z=z_0$ to $z=h$, defined by

$$Q_H = \rho\, c_p\, \frac{T_0 - T_h}{r_h}$$

If the flux is height-independent it is easy to prove that

$$r_h = \int_{z_0}^{h} \frac{dz}{K(z)}\, dz$$

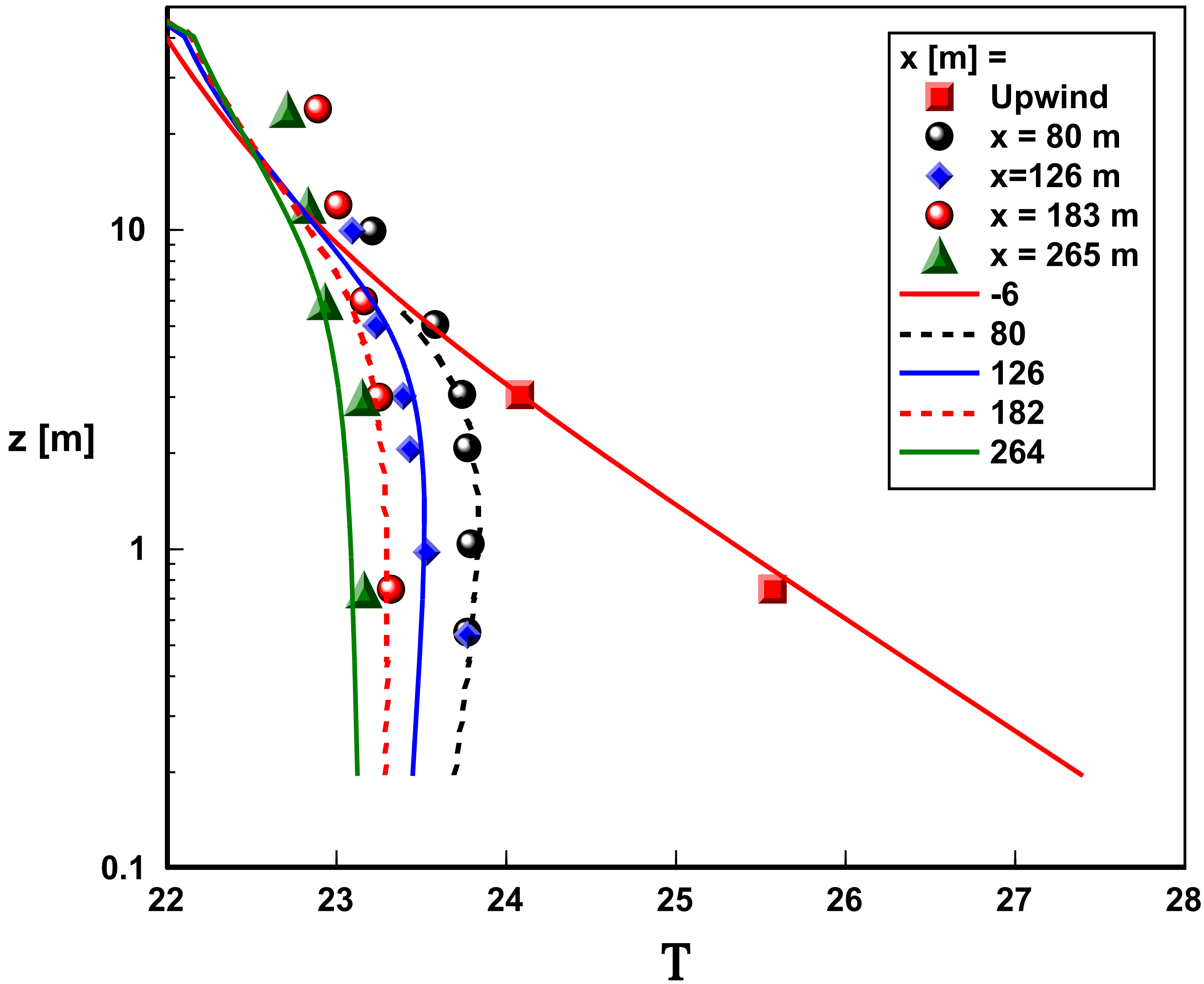
We can use MOST (entailing the assumption of height-independent flux) to calibrate the resistance:

$$\frac{k_v\, z}{(-\overline{w'T'})/u_*}\, \frac{\partial \overline{T}}{\partial z} = \phi_h\left(\frac{z}{L}\right)$$

$$\int_{z_0}^{h} \frac{\partial \overline{T}}{\partial z}\, dz = \overline{T}(h) - \overline{T}(z_0) = \frac{-Q_H}{\rho\, c_p} \int_{z_0}^{h} \frac{\phi_h(z/L)}{k_v\, u_*\, z}\, dz$$

r_h

Observations at La Crau Valley (France) versus numerical solution of conservation equations using RWC 2nd-order closure – modification of the mean temperature profile

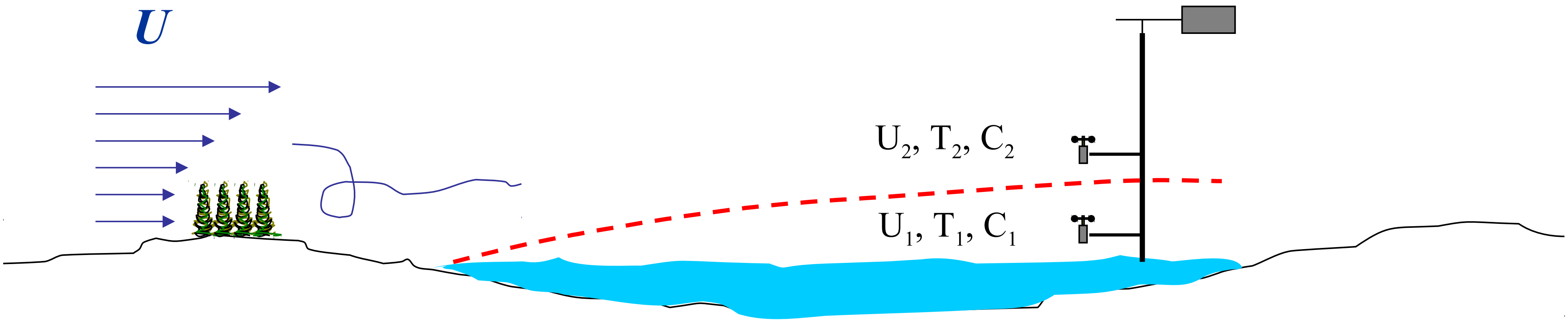


An application of the RWC local advection model…

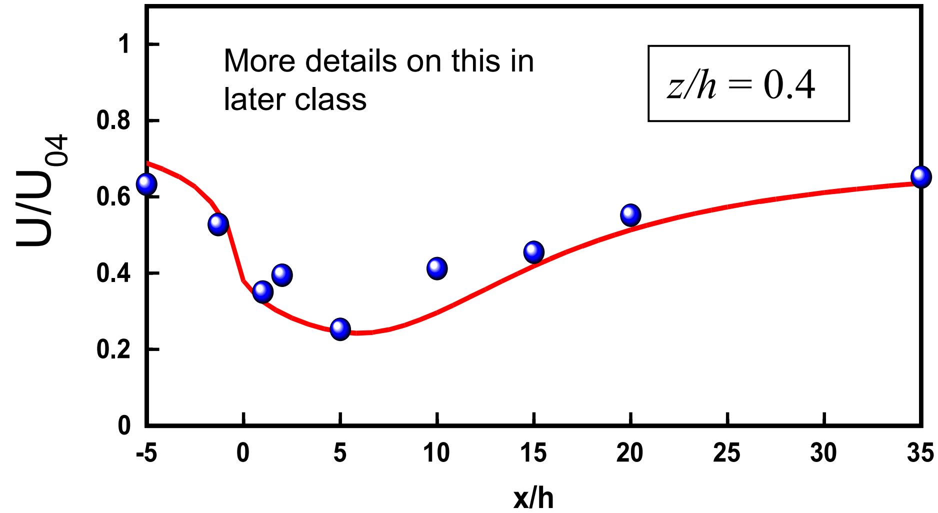
Inference of gas emissions (NH_3, CH_4) from agricultural lagoons… familiar techniques are predicated on a horizontally-uniform flow and existence of a constant flux layer over the source.

Generate a "synthetic" lagoon flow and test several micromet methods

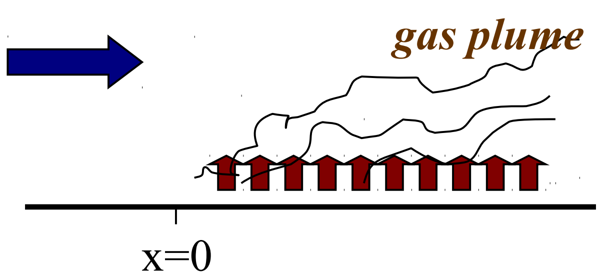
Preparation entailed comparing RWC with the La Crau local advection expt. – adding a passive scalar and comparing with Project Prairie Grass – and adding a windbreak momentum sink to test model's treatment of windbreak flow (covered in detail elsewhere). The RWC model performed very well in all tests

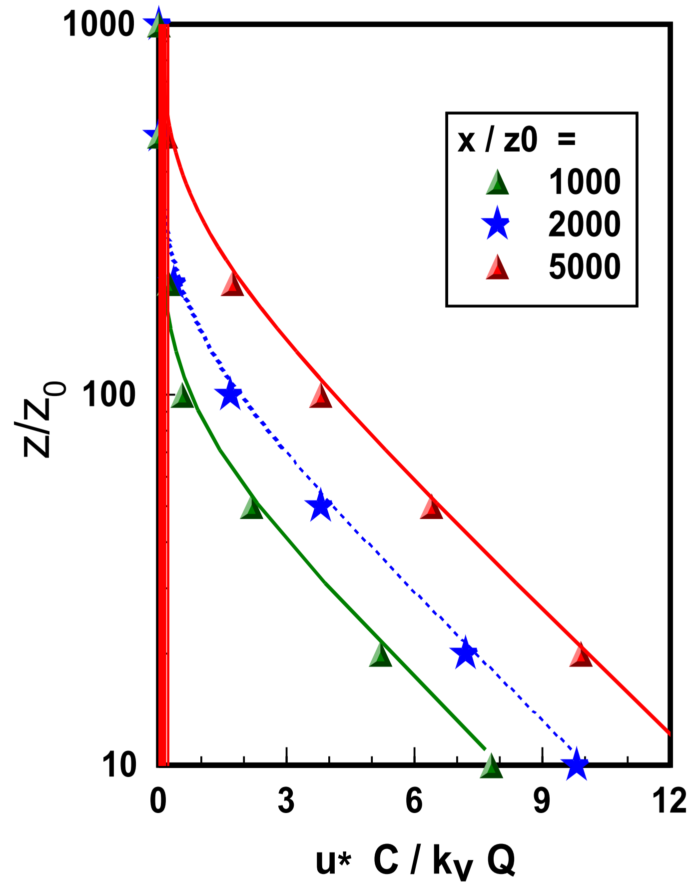


Mean wind reduction behind a long porous fence (h/z_0=600, k_r=2) mounted perpendicular to neutrally-stratified flow… Mulhearn & Bradley field observations versus solution of (augmented) RWC conservation equations:

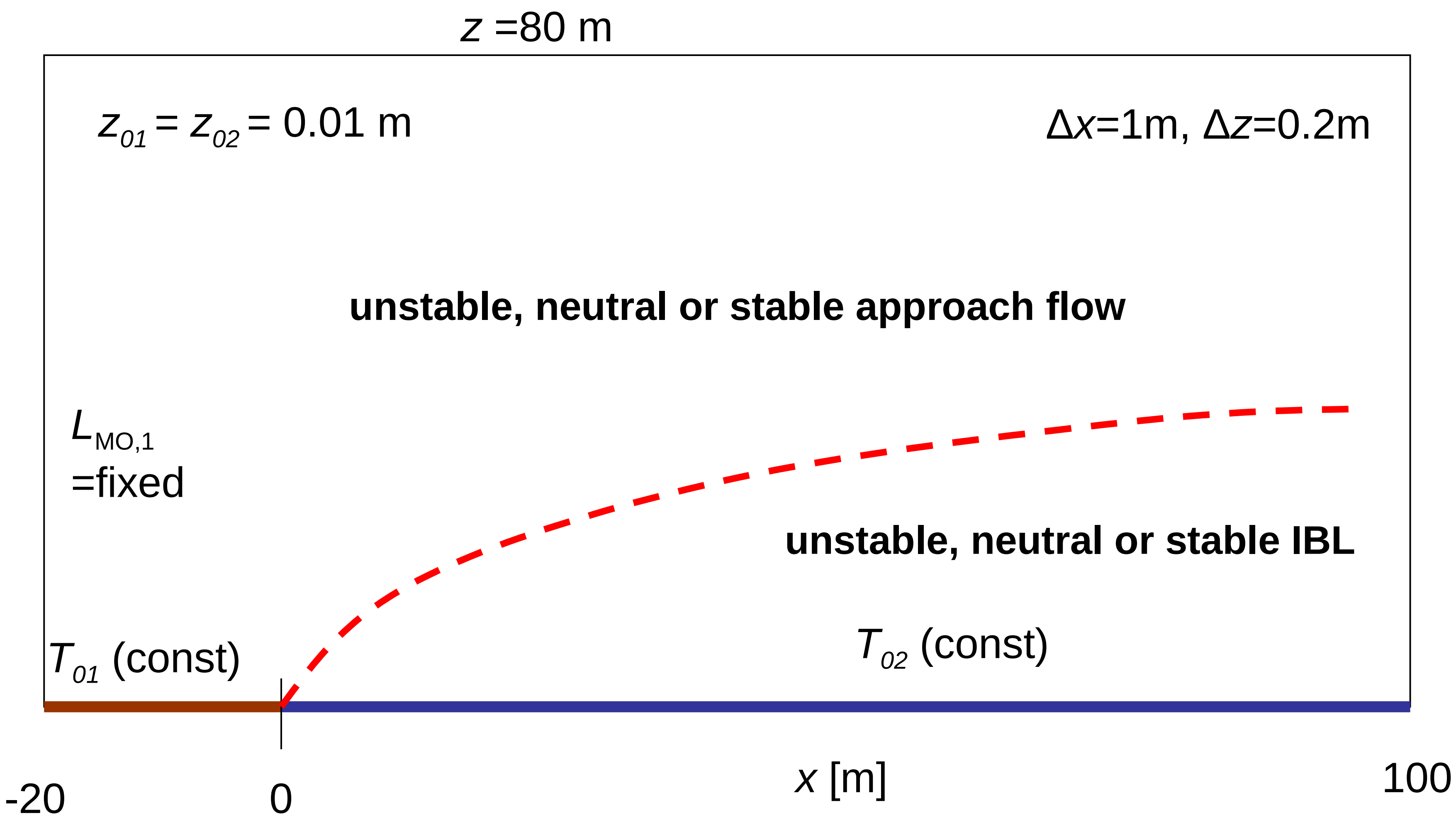


Tracer concentration field from a ground level area source at x>0, in horizontally-uniform and neutral flow… lines from RWC, symbols from the well-mixed Lagrangian stochastic model for this flow (known to agree with Prairie Grass):

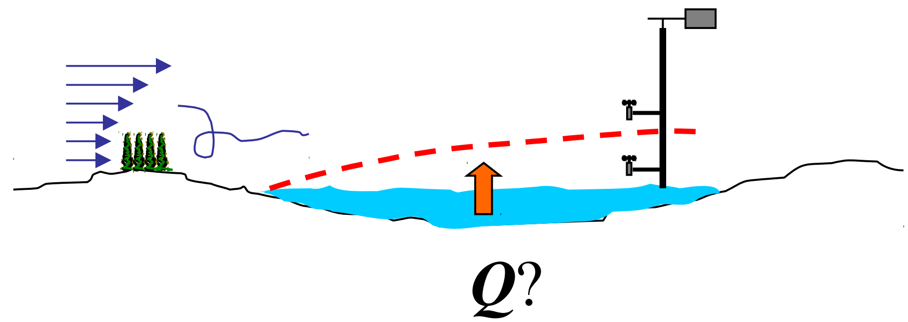


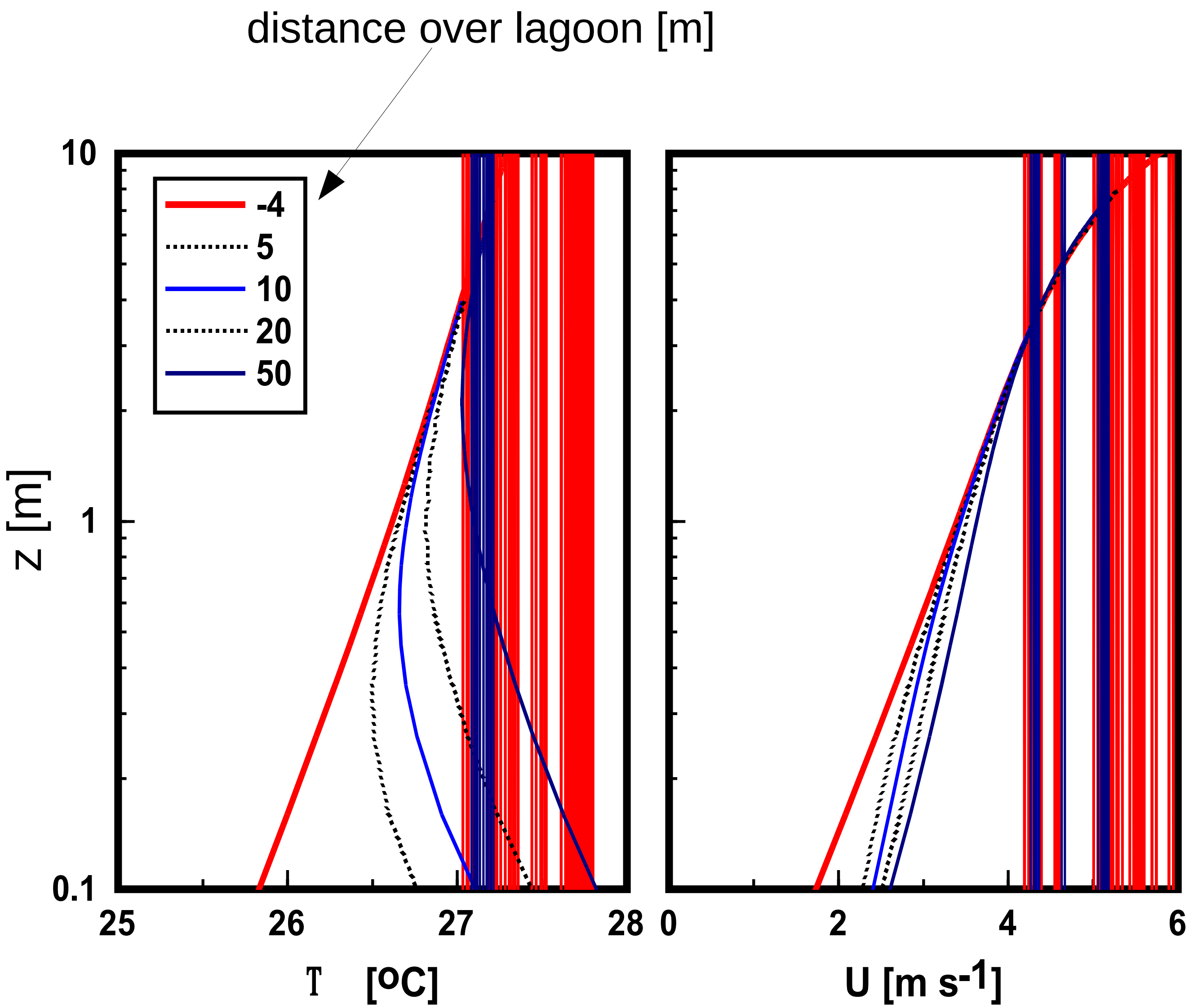


Proved model is competent to generate disturbed field of wind, temperature, humidity, tracer gas (and their fluxes)… now generate synthetic lagoon flow…

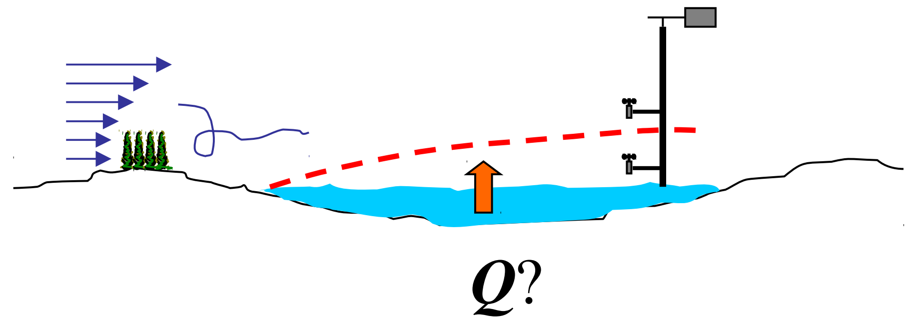


Stable approach flow encounters a warm lagoon, $T_{lag}=T_{up}+5$ (case F)





Performance of flux-estimators…



Lagoon	Upwind			Downwind (x = 50 m)			Flux-Gradient			IHF	BLS	
		T_{up} (C)	L (m)		T_{lag} (C)	L (m)	z_1= 0.15 z_2 = 0.4	z_1= 0.4 z_2 = 0.65	z_1= 0.4 z_2 = 1.4		up	down
A		25	-23		25	-12	0.91	0.78	0.72	1.09	0.82	0.89
B		25	-23		30	-7	0.94	0.83	0.79	1.08	0.82	0.87
C		25	-23		20	-27	0.85	0.71	0.63	1.10	0.82	0.89
D		25	-23		15	103	0.82	0.62	0.52	1.09	0.82	0.88
E		25	48		25	23	0.61	0.43	0.36	1.03	0.80	0.86
F		25	48		30	-22	0.69	0.56	0.53	1.03	0.80	0.89
G		25	48		20	6	0.51	0.29	0.19	1.03	0.80	0.87
H		25	48		15	2	0.43	0.18	0.09	1.03	0.80	1.15
I		20	-2300		30	-6	0.92	0.86	0.85	1.05	0.88	0.94
J		20	-2300		10	5	0.61	0.37	0.23	1.06	0.88	0.87

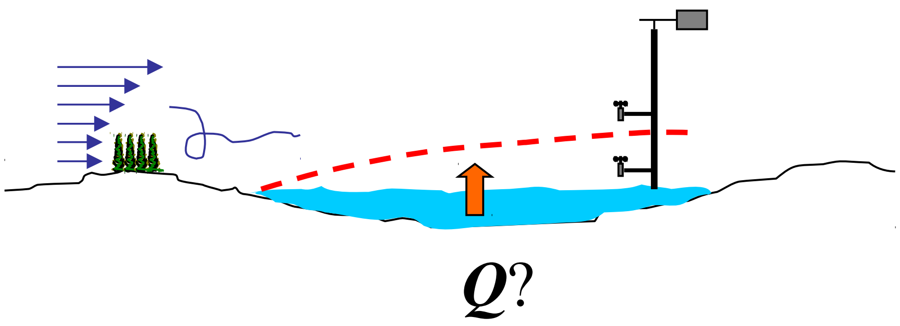
Conclusions…

RWC local advection model does plausible job of calculating disturbed microclimate, as judged by its comparison with

- observed development of (T,Q) in flow from dry to moist land
- tracer dispersion (indirectly verified against Prairie Grass)
- reduction in mean wind speed behind a fence

When flux estimators are applied to synthetic "data" at x = 50m over the lagoon, Integrated Horizontal Flux method (i.e. mass balance) excellent, 10% or better (model-independent, but practicality depends on geometric simplicity); backwards LS (model-based, source-receptor method) also very good (20%) despite neglect of flow disturbance; flux-gradient method (which assumes existence of a constant flux layer that does not prevail in distrubed flow except in the growing equilib. layer) very poor in some cases

Broader conclusion relative to eas572 – this and other examples will illustrate the prevalent way of thinking relative to flow disturbances; and show we have some skill in the mathematical representation of disturbed micromet flows. The basic limitation is the closure problem (RANS models far from perfect); as yet LES impractical for routine application to disturbed flows



Disturbed micrometeorological flows