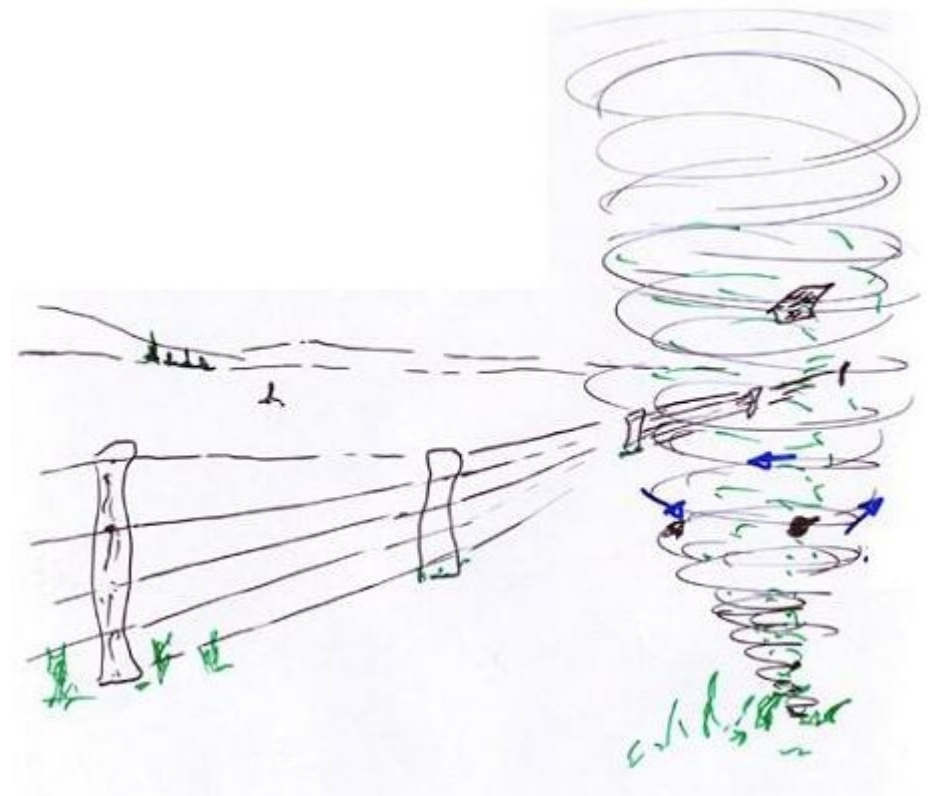


# Modelling disturbed micrometeorological flows



- will illustrate the prevalent way of thinking relative to flow disturbances
- show we have some skill in the mathematical representation of disturbed micromet flows. The basic limitation is the closure problem: RANS (Reynolds-Averaged Navier-Stokes) models far from perfect; as yet LES (Large Eddy Simulation) impractical for routine application to disturbed flows



## (Modelling) disturbed micrometeorological flows – example – “local advection”

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

- by inhomogeneity in the surface boundary conditions\*\* – inhomogeneity in surface properties and fluxes e.g.  $\Delta Q_{H0}$ ,  $\Delta Q_{E0}$ ,  $\Delta Z_0, \dots$  due to varying soil moisture, surface elevation/cover ,...
- by purely aerodynamic disturbances (windbreaks, hills, buildings,...)
- by a combination of these types of influences

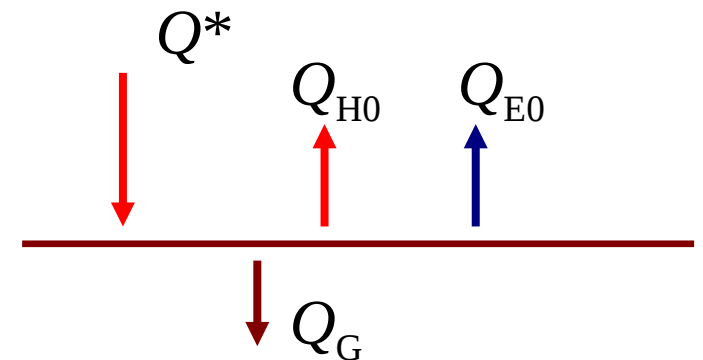
Note: the flow need not be disturbed at the boundary in order to be inhomogeneous

\* 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

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)

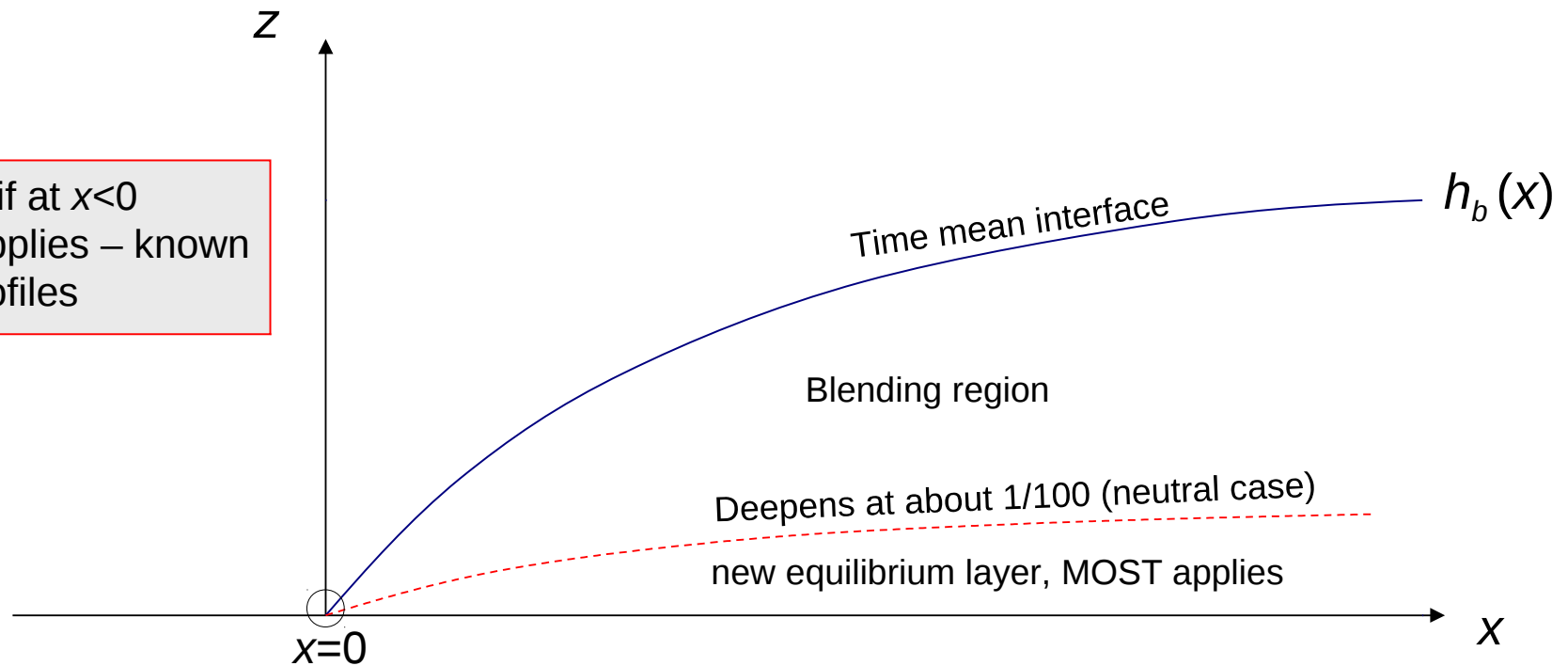
$$Q^* = Q_{H0} + Q_{E0} + Q_G$$



\*\*The surface energy budget

# The paradigm of the Internal Boundary Layer

Horiz. unif at  $x < 0$   
 MOST applies – known  
 inflow profiles



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 propagates 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  $\mathbf{P}$  responds to r.h.s. over *all* positions  $\mathbf{r}$ , weighted as  $|\mathbf{P} - \mathbf{r}|^{-2}$ . This implies a disturbance has upstream influence

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

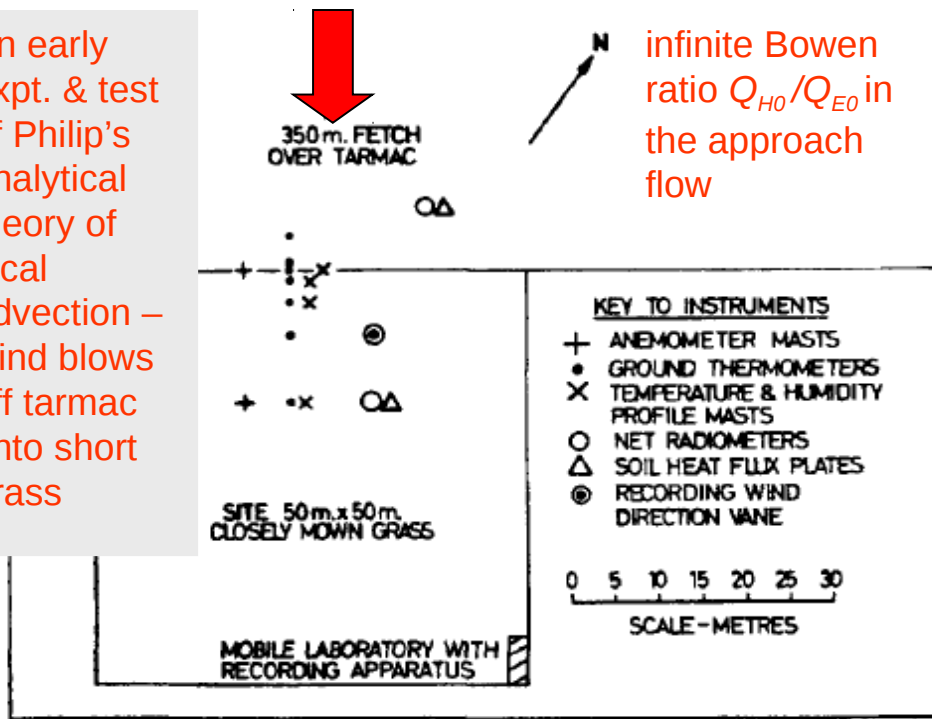


Figure 2. The experimental site, together with the disposition of instruments.

## The horizontal transport of heat and moisture – a micrometeorological study

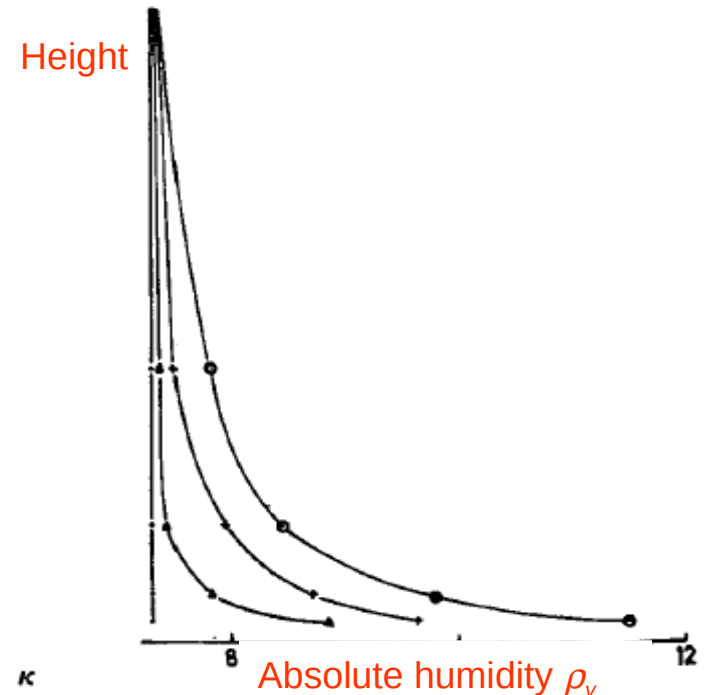
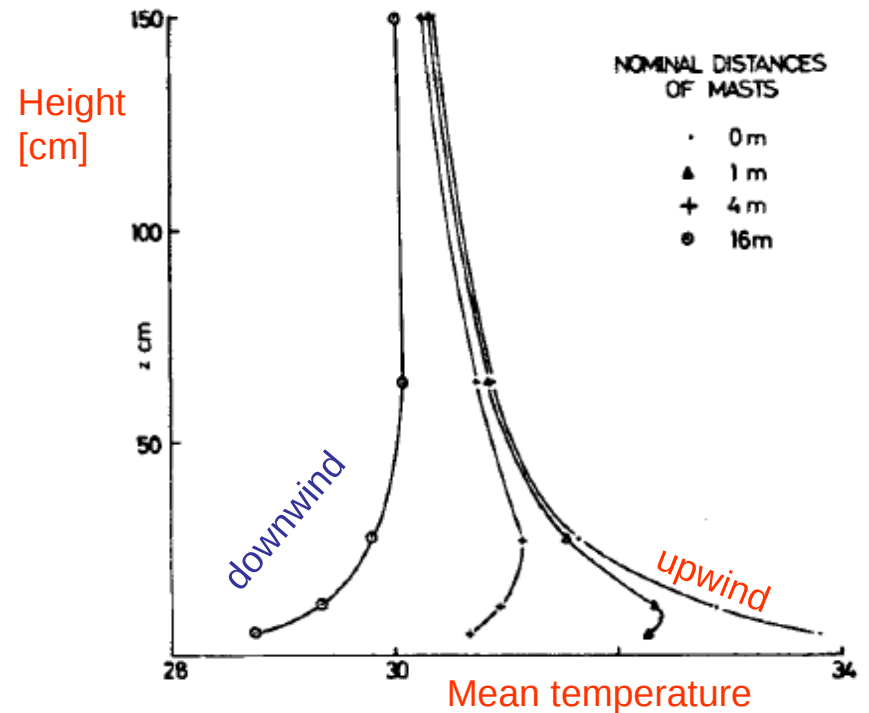
By N. E. RIDER\*, J. R. PHILIP and E. F. BRADLEY  
C.S.I.R.O., Division of Plant Industry, Canberra, Australia

QJRMS Vol. 89, 1963

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

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

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

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

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

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

invariably provides a further condition. Here  $R_s$  and  $R_a$  are the flux densities of atmospheric and short wave radiation at the surface;  $\tau$  is the reflection coefficient of the surface for short wave radiation;  $\epsilon$ , the surface emissivity;  $\sigma$ , 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;  $\rho_w$ , the density of liquid water; and  $E$ , the rate of evaporation. We notice that in Eq. (4)

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

where  $c$  and  $\rho$  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 :

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

Incoming terrestrial (longwave) radiation

Outgoing terrestrial (longwave) radiation

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).

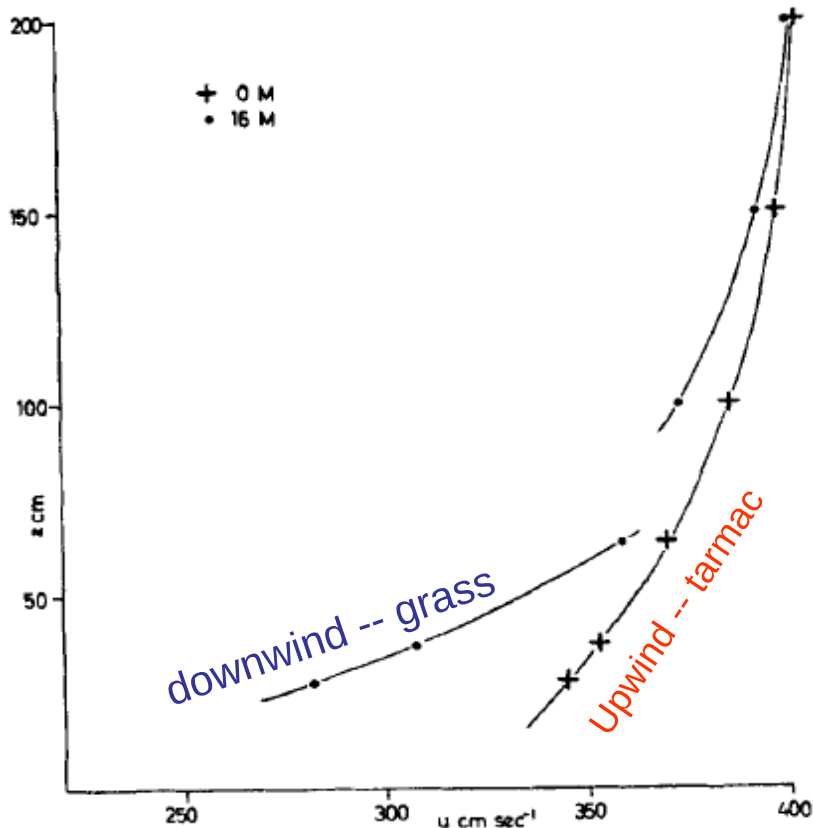
$$\bar{T}_{eq} = \bar{T} + \frac{\bar{e}}{\gamma} \quad \text{y the psychrometric constant - temperature if all latent heat converted to sensible}$$

$$\bar{D} = e_{\text{sat}}(\bar{T}) - \bar{e} \quad \text{e is vapour pressure}$$

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

LHS constrained at gnd by sfc energy balance

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



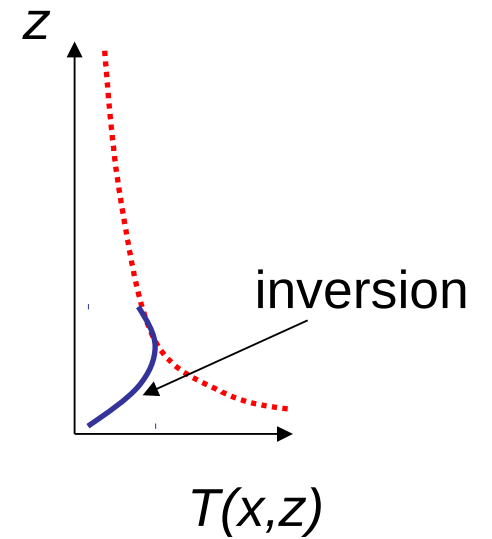
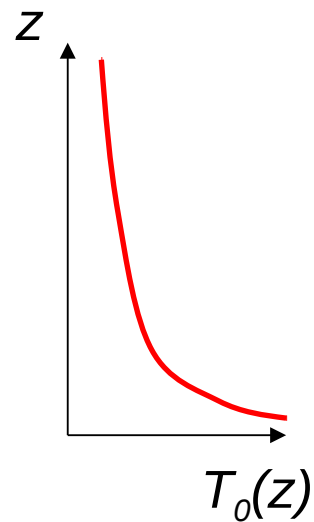
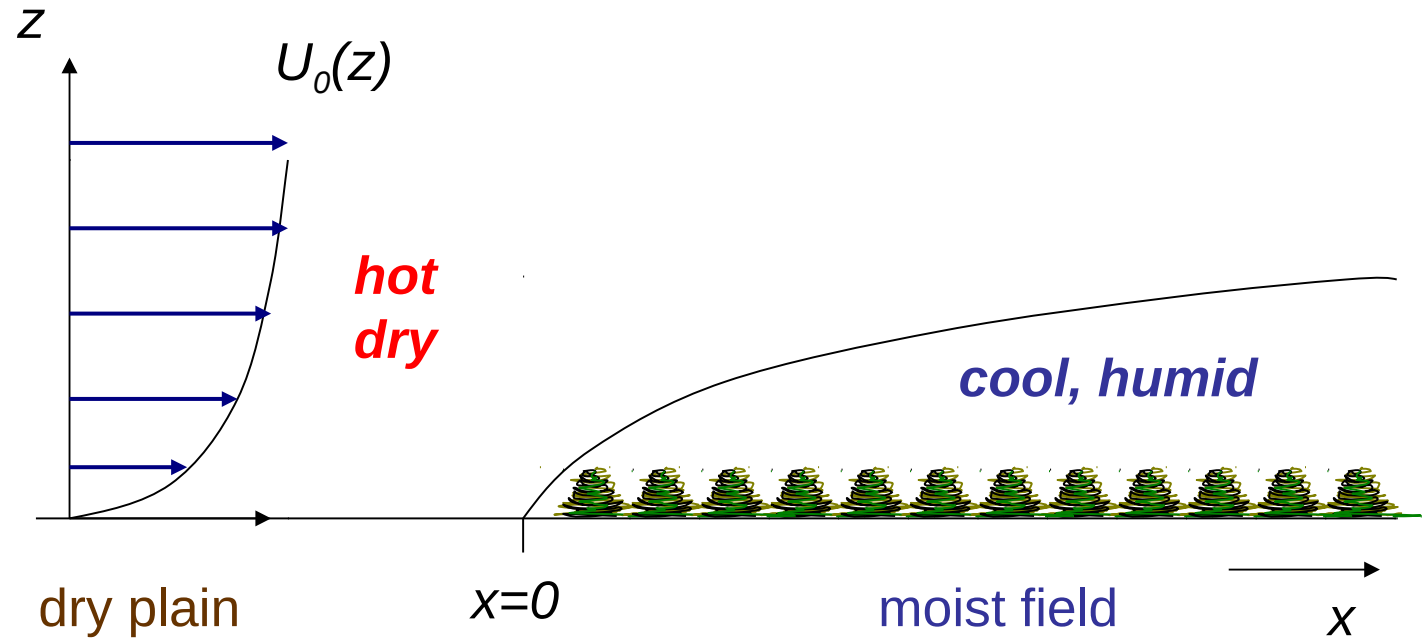
Observed variation of the profile of the (advecting) mean wind speed implies one should account for conservation of momentum as well as heat and latent heat

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

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

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

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



# Rao-Wyngaard-Coté 2<sup>nd</sup>-order closure model of local advection:

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

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

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

**U-mtm:**

$$\underbrace{\frac{\partial}{\partial x} [U U + \overline{u'^2}]}_{\text{hor.flx.}} + \underbrace{\frac{\partial}{\partial z} [W U + \overline{w'u'}]}_{\text{vrt.flx.}} = -\frac{1}{\rho_0} \left( \frac{\partial P}{\partial x} \right) \text{ pressure disturbance (not allowed for in original RWC treatment)}$$

**$\sigma_u^2$ :**

$$\underbrace{\frac{\partial}{\partial x} \left[ U \overline{u'^2} - a_t \tau \overline{u'^2} \frac{\partial \overline{u'^2}}{\partial x} \right]}_{\substack{\text{hor.flx.} \\ \text{adv.} \quad \text{diff.}}} + \underbrace{\frac{\partial}{\partial z} \left[ W \overline{u'^2} - a_t \tau \overline{w'^2} \frac{\partial \overline{u'^2}}{\partial z} \right]}_{\substack{\text{vrt.flx.} \\ \text{effective diffusivity}}} = \underbrace{-2\overline{u'^2} \frac{\partial U}{\partial x} - 2\overline{u'w'} \frac{\partial U}{\partial z}}_{\text{shear production}} - \dots$$

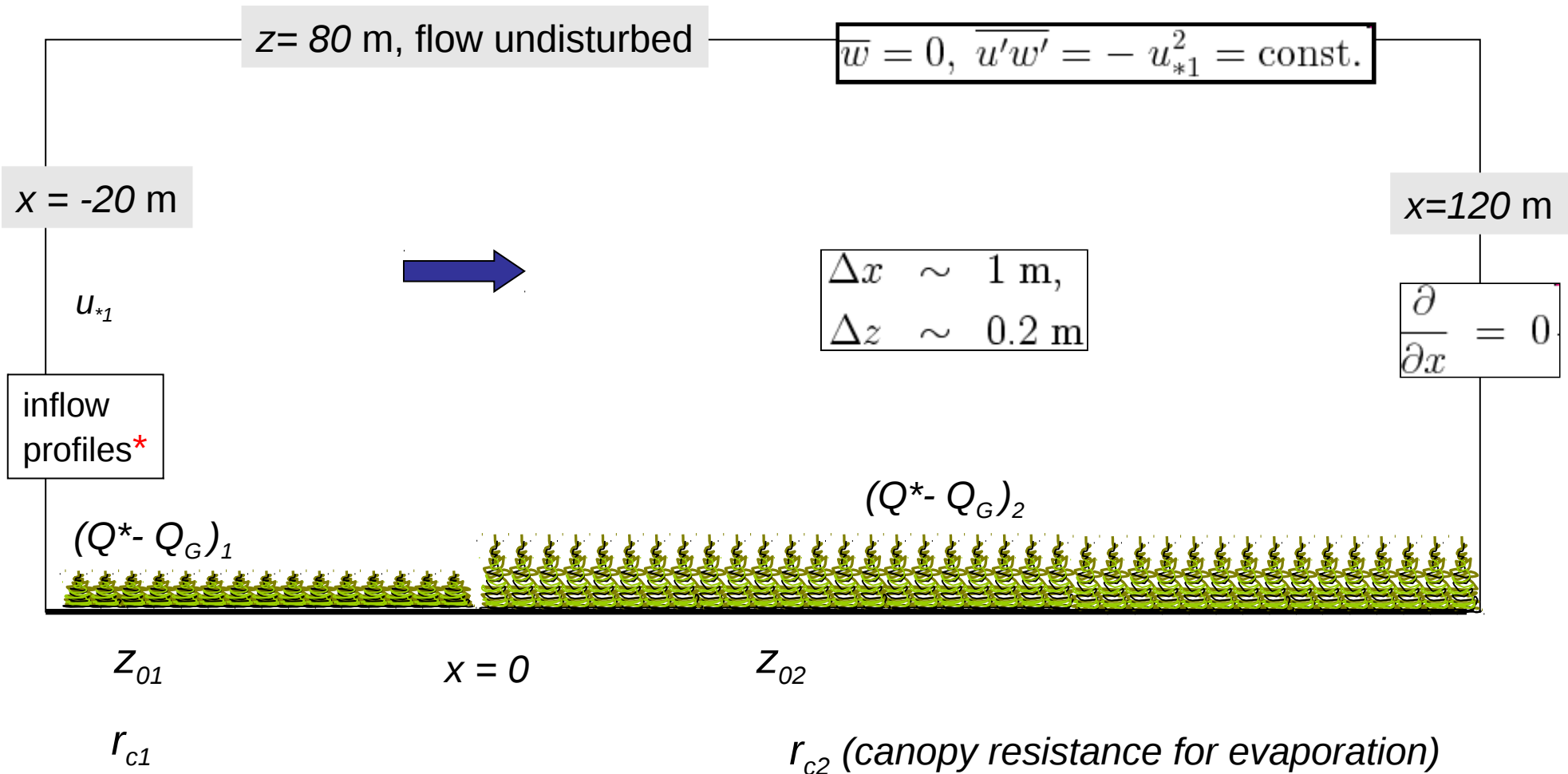
$$\dots - \frac{2}{3} \epsilon - \frac{c_{11}}{\tau} \left[ \overline{u'^2} - \frac{2}{3} k \right] + \frac{\partial}{\partial x} \left[ a_t \tau \overline{u'w'} \frac{\partial \overline{u'^2}}{\partial z} \right] + \frac{\partial}{\partial z} \left[ a_t \tau \overline{u'w'} \frac{\partial \overline{u'^2}}{\partial x} \right]$$


**dissip'n**                      **redistrib.**

$$\tau = \frac{2k}{\epsilon} \quad \text{a turbulence time scale}$$

The closure constants  $a_t$ ,  $c_{11}$ , etc. are **not free** – constrained by forcing the model to (analytically) reproduce the ideal hh\_NSL

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



  
 ELSEVIER  
 Agricultural and Forest Meteorology 107 (2001) 207–225  
 www.elsevier.com/locate/agrformet  
 AGRICULTURAL AND FOREST METEOROLOGY  
 \*\*  
 Micro-meteorological methods for estimating surface exchange with a disturbed windflow  
 John D. Wilson<sup>a,\*</sup>, Thomas K. Flesch<sup>a</sup>, Lowry A. Harper<sup>b</sup>  
<sup>a</sup> Department of Earth and Atmospheric Sciences, University of Alberta, Edmonton, Alta., Canada T6G 2E3  
<sup>b</sup> United States Department of Agriculture, Watkinsville, GA, USA

\*to obtain profiles for the inflow boundary, the model eqns are solved with all  $\partial/\partial x$  terms set to zero, and with conditions appropriate to the upwind surface, i.e.  $u_{*1}, z_{01}, (Q^* - Q_G)_1$

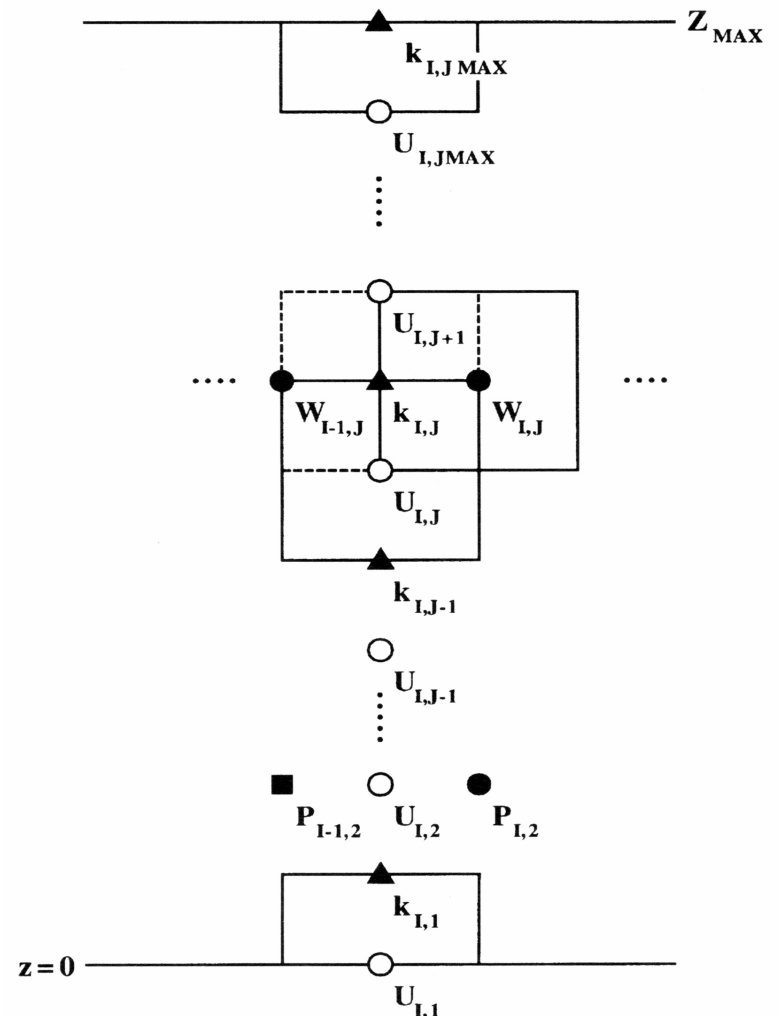
# Computational method – illustrated for a case with symmetry along y axis

- steady state
- coupled, non-linear equations
- staggered grid, index gridpoints I,J
- linearize (e.g.)  $U_{I,J} U_{I,J}$  as  ${}^m U_{I,J} {}^{m+1} U_{I,J}$  where  $m$  denotes the  $m$ th iterative guess and  $m+1$  the next guess
- use an ADI (alternating direction implicit) method...  
because  $U_{I,J}$  is coupled by "neighbour eqns" to its immediate neighbours on the grid
- incorporate "relaxation" to help ensure convergence, e.g.

$$U_{I,J}^{m+1} \leftarrow \alpha U_{I,J}^m + (1-\alpha) U_{I,J}^{m+1}$$

with relaxation parameter  $\alpha < 1$

- continue to iterate until momentum is conserved cell by cell and for the domain in total



# La Crau run 42 – specification of controlling boundary conditions



Vincent Van Gogh: "Harvest at La Crau"

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

$$T_1(3.05m) = 24.08 \text{ K}$$

$$Q_1(3.05m) = 6.6 \text{ g kg}^{-1}$$

$$u_{*1} = 0.63 \text{ m s}^{-1}$$

$$z_{01} = 0.01 \text{ m}$$

$$Q_{*1} - Q_{G1} = 434 \text{ W m}^{-2}$$

$$Q_{H1} = 362 \text{ W m}^{-2}$$

$$z_{02} = 0.07 \text{ m}$$

$$Q_{*2} - Q_{G2} = 500 \text{ W m}^{-2}$$

$$r_{c2} = 47 \text{ s m}^{-1}$$

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

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

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

- $\lambda$  the latent heat of vapourization;  $\epsilon_{sa}$  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

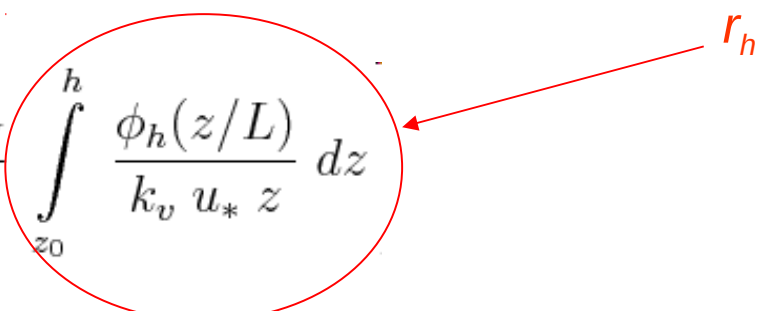
e.g. let  $r_h$  be the transfer resistance for heat between levels  $z=z_0$  to  $z=h$ , defined by

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

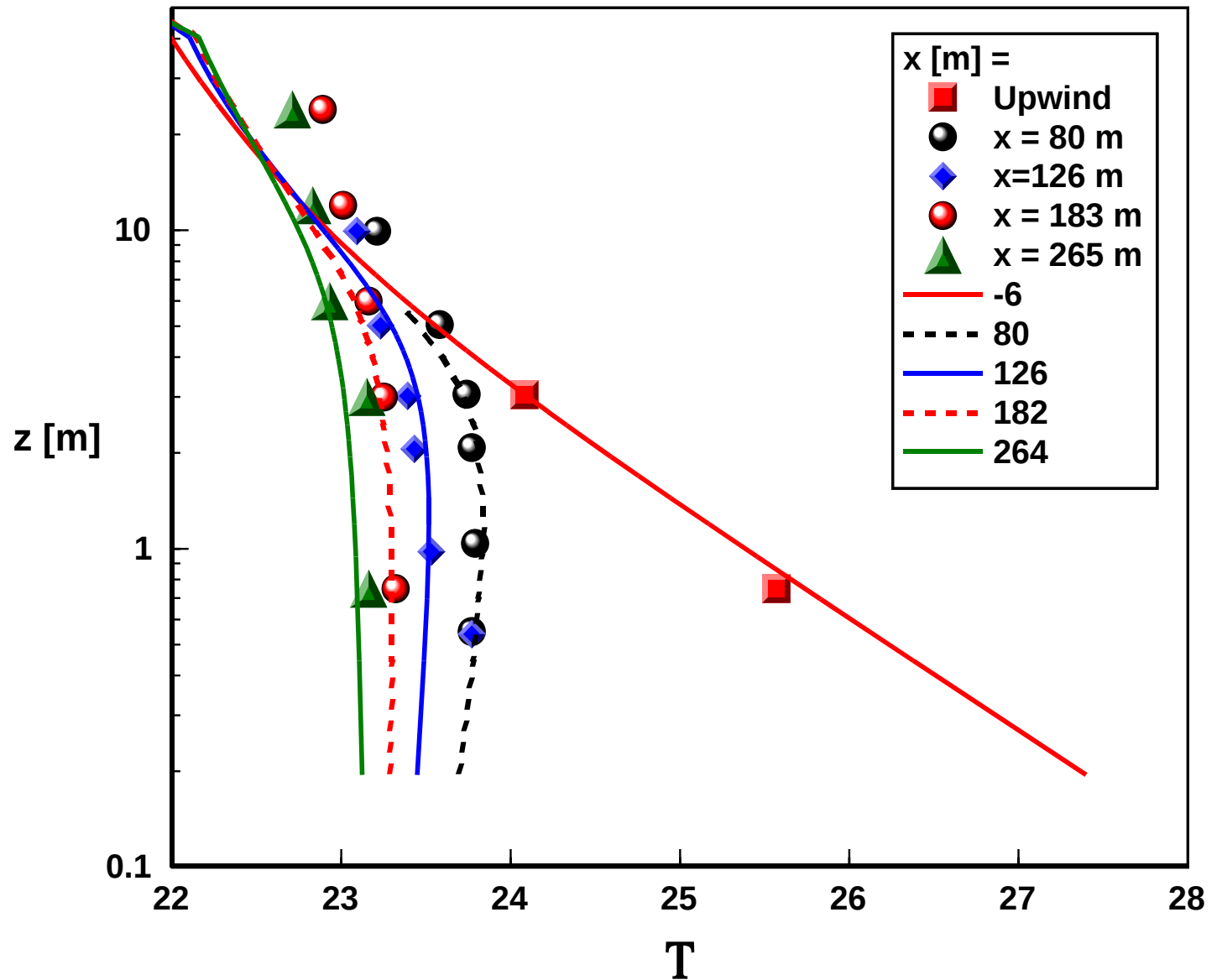
If the flux is height-independent it is easy to prove that  $r_h = \int_{z_0}^h \frac{dz}{K(z)}$

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

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

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


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

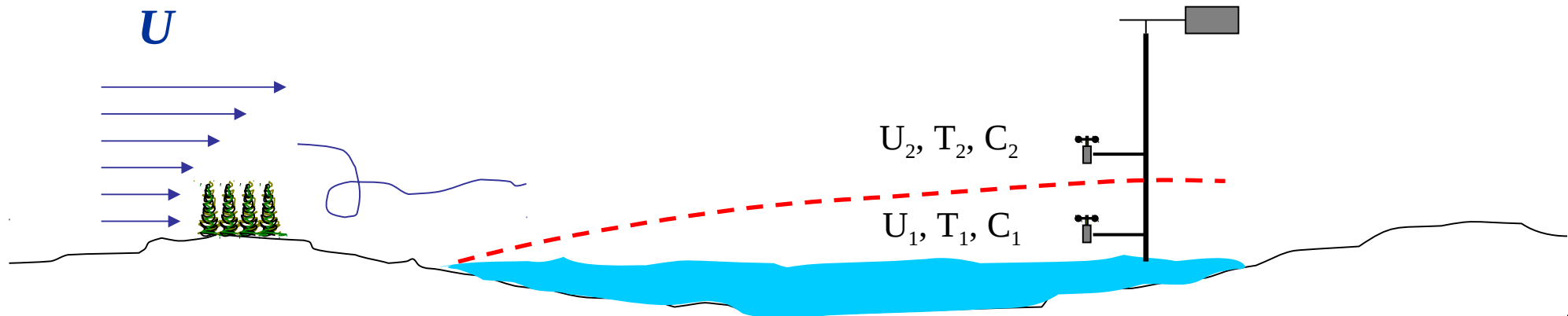


## An application of the RWC local advection model...

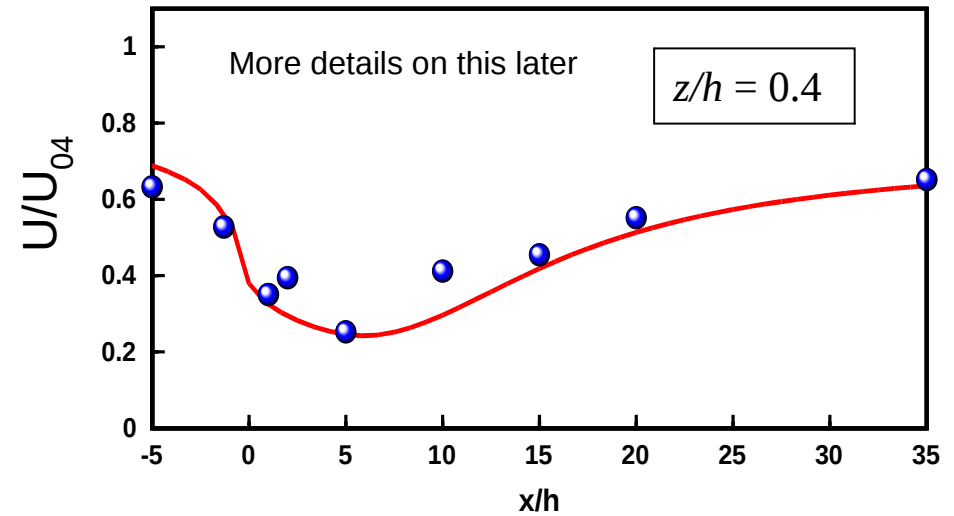
Inference of gas emissions ( $\text{NH}_3$ ,  $\text{CH}_4$ ) from agricultural lagoons... familiar techniques are predicated on a horizontally-uniform flow and existence of a constant flux layer over the source.

Generate a “synthetic” lagoon flow and test several micromet methods

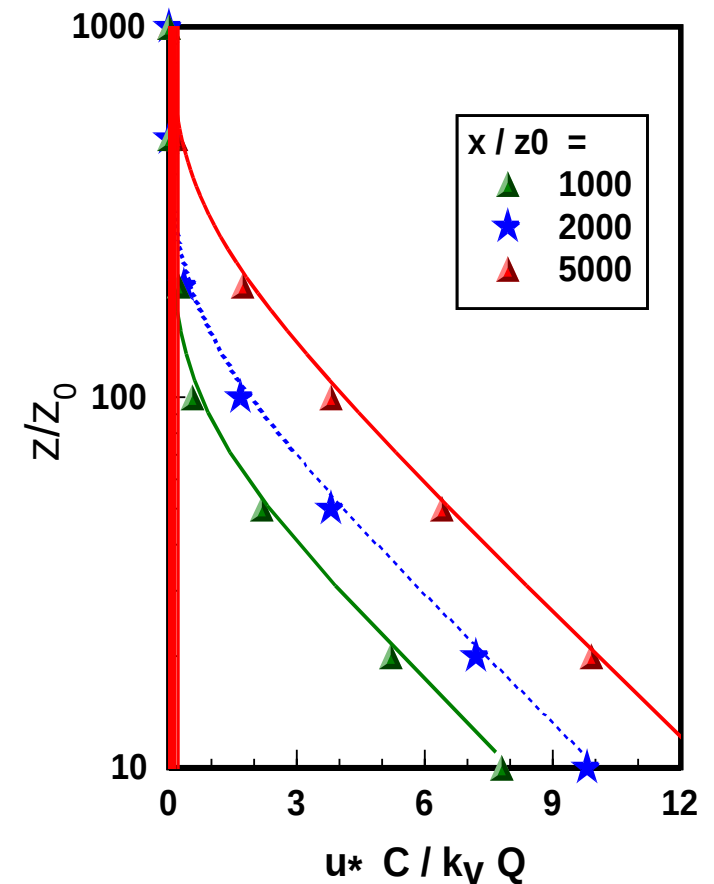
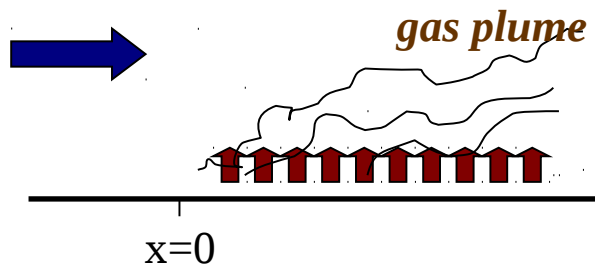
Preparation entailed comparing RWC with the La Crau local advection expt. – adding a passive scalar and comparing with Project Prairie Grass – and adding a windbreak momentum sink to test model’s treatment of windbreak flow (covered in detail elsewhere). The RWC model performed very well in all tests



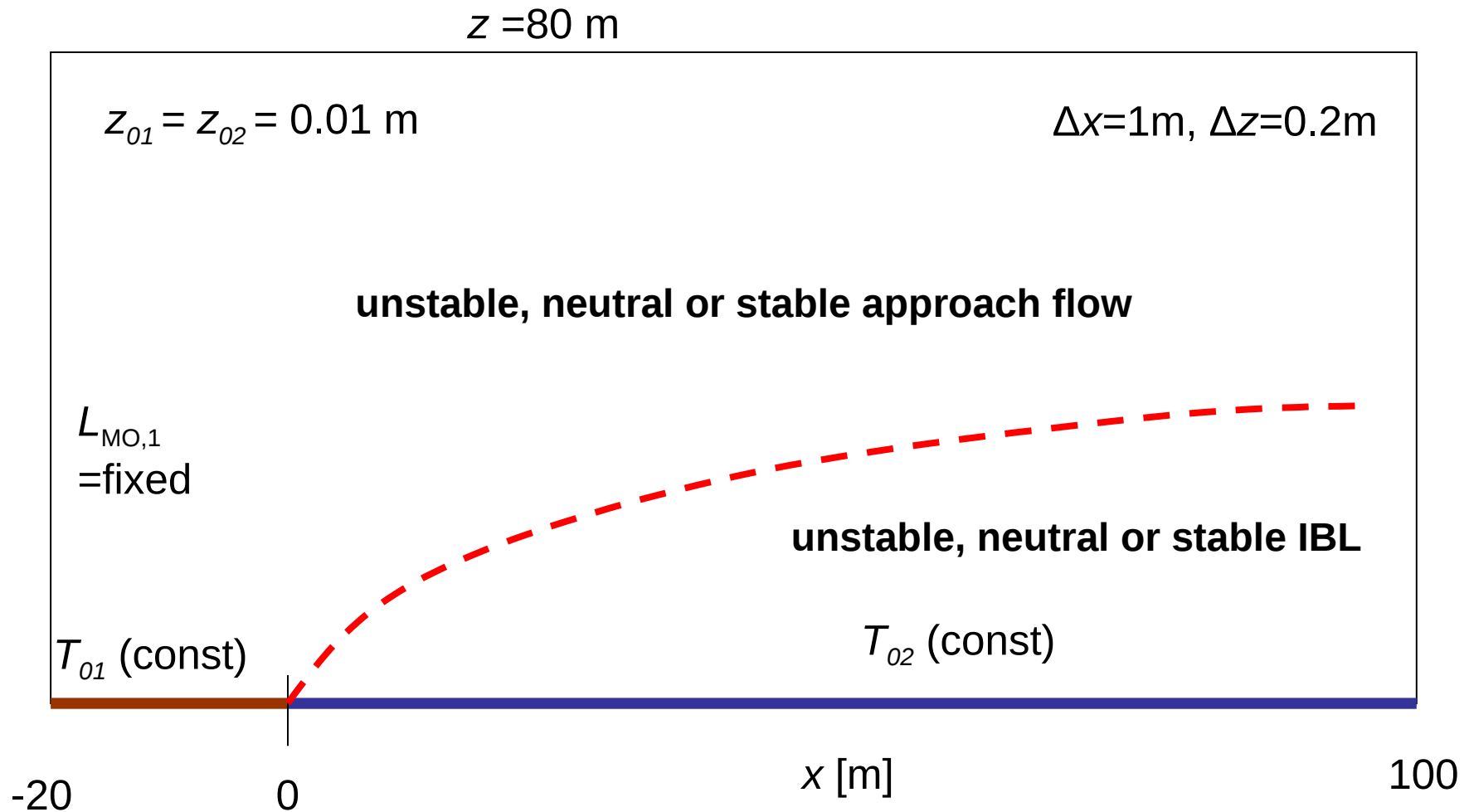
Mean wind reduction behind a long porous fence ( $h/z_0=600$ ,  $k_r=2$ ) mounted perpendicular to neutrally-stratified flow... Mulhearn & Bradley field observations versus solution of (augmented) RWC conservation equations:



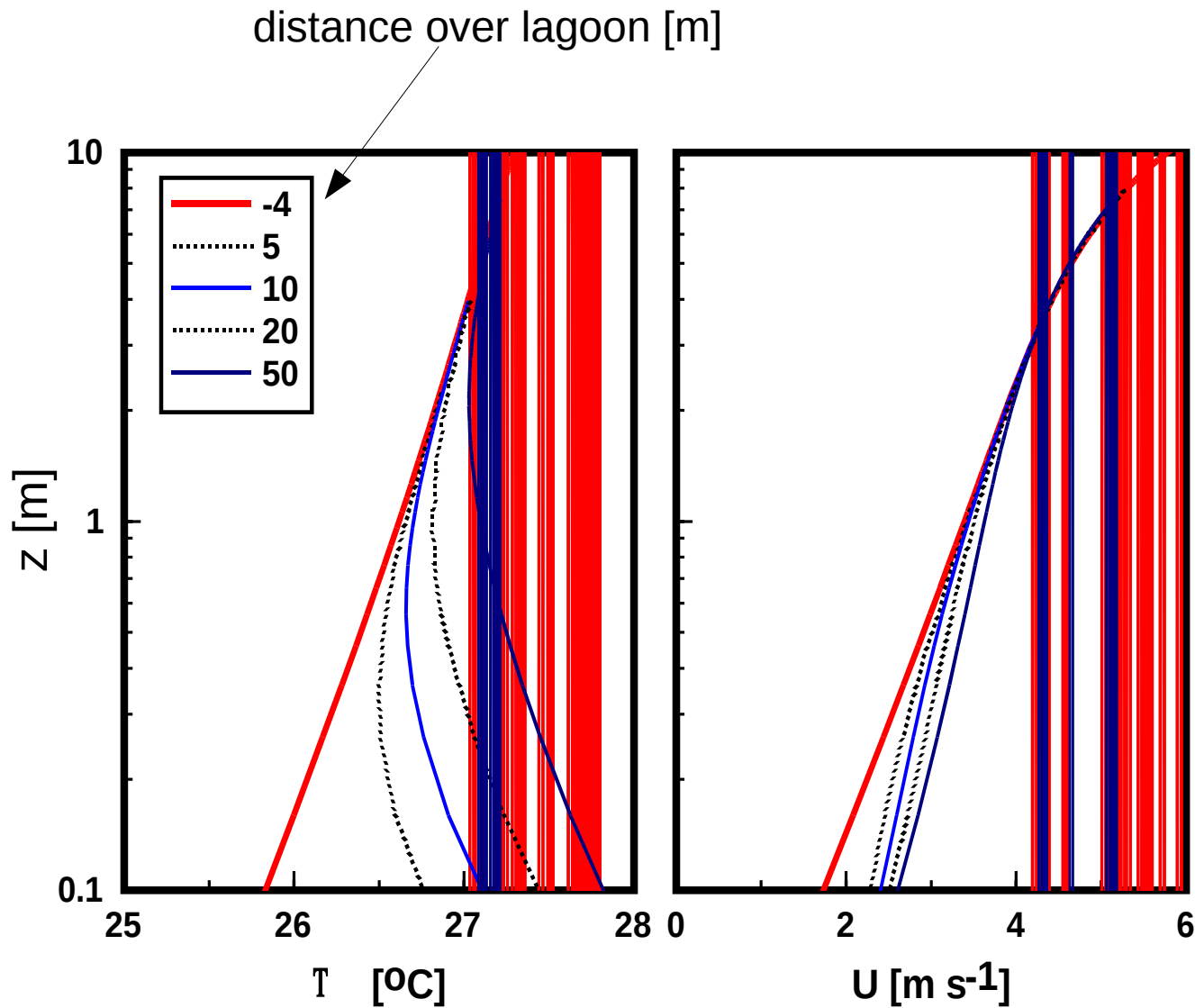
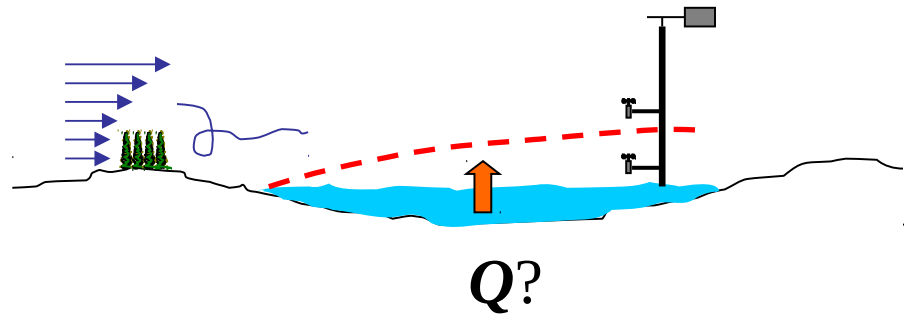
Tracer concentration field from a ground level area source at  $x>0$ , in horizontally-uniform and neutral flow... lines from RWC, symbols from the well-mixed Lagrangian stochastic model for this flow (known to agree with Prairie Grass):



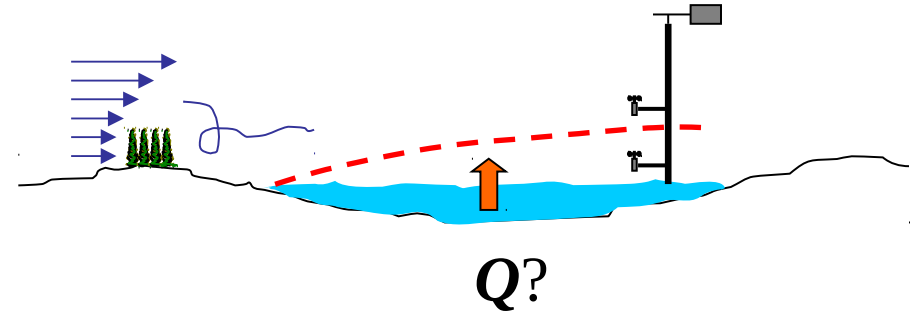
Proved model is competent to generate disturbed field of wind, temperature, humidity, tracer gas (and their fluxes)... now generate synthetic lagoon flow...



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



# Performance of flux-estimators...



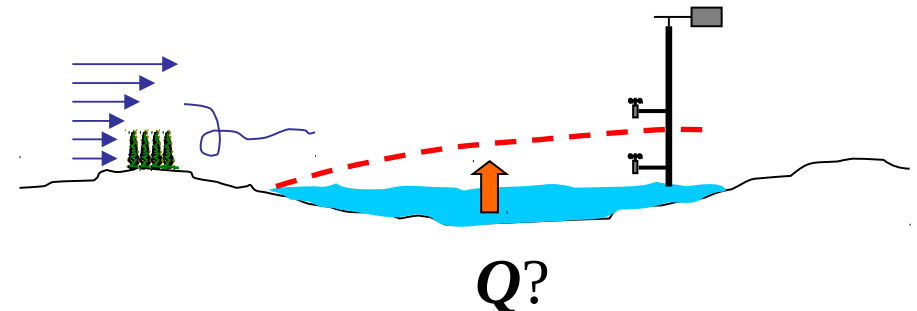
Lagoon	Upwind		Downwind (x = 50 m)		Flux-Gradient			IHF	BLS	
	T <sub>up</sub> (C)	L (m)	T <sub>lag</sub> (C)	L (m)	z <sub>1</sub> = 0.15 z <sub>2</sub> = 0.4	z <sub>1</sub> = 0.4 z <sub>2</sub> = 0.65	z <sub>1</sub> = 0.4 z <sub>2</sub> = 1.4		up	down
A	25	-23	25	-12	0.91	0.78	0.72	1.09	0.82	0.89
B	25	-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  $x = 50\text{m}$  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 disturbed flow except in the growing equilib. layer) very poor in some cases

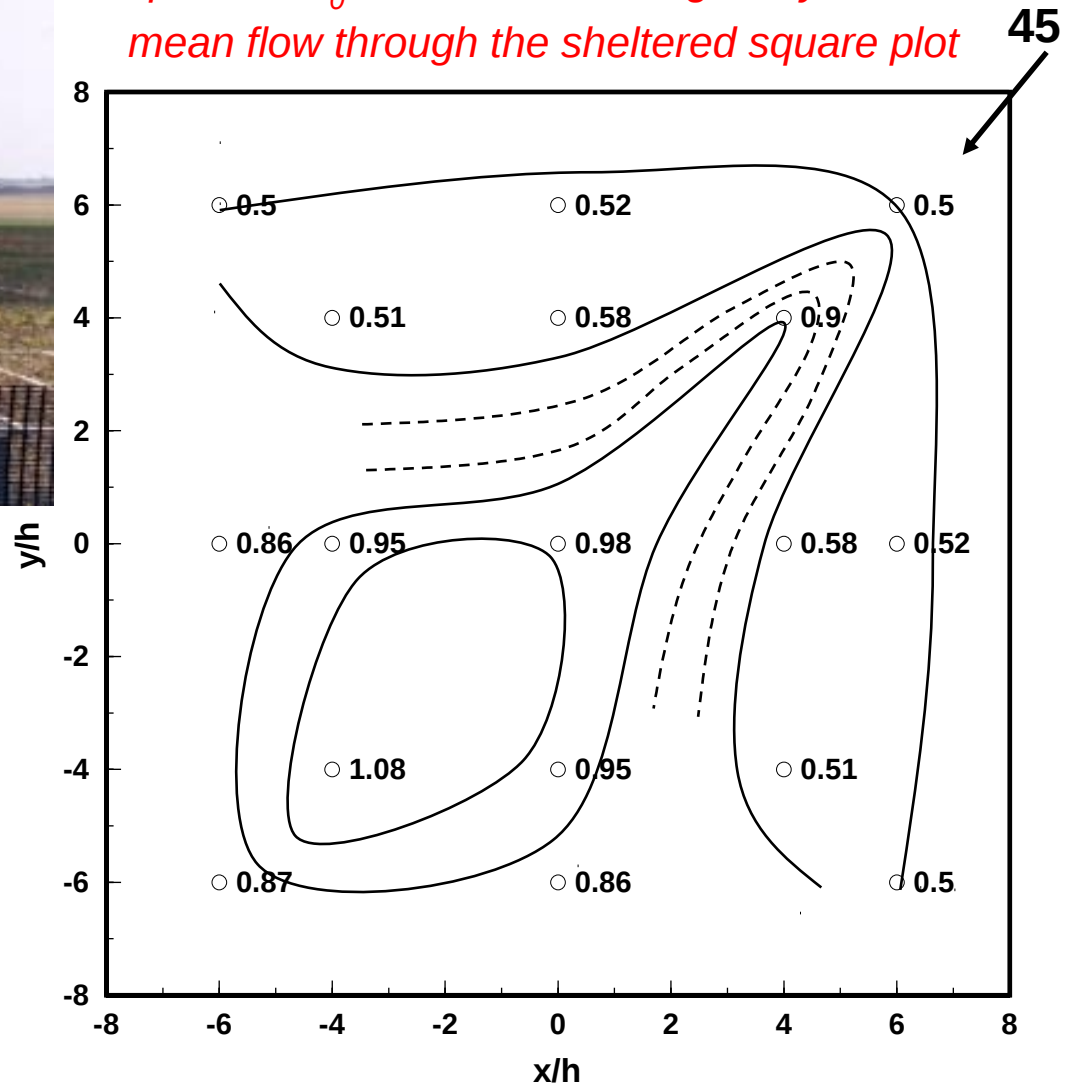


## Disturbed micro-meteorological flows (ctd): flow around windbreaks

- basic effects observed
- elements of the theoretical description



*Contours of measured relative mean wind speed  $U/U_0$  at  $z/h=0.5$ , for diagonally-incident mean flow through the sheltered square plot*

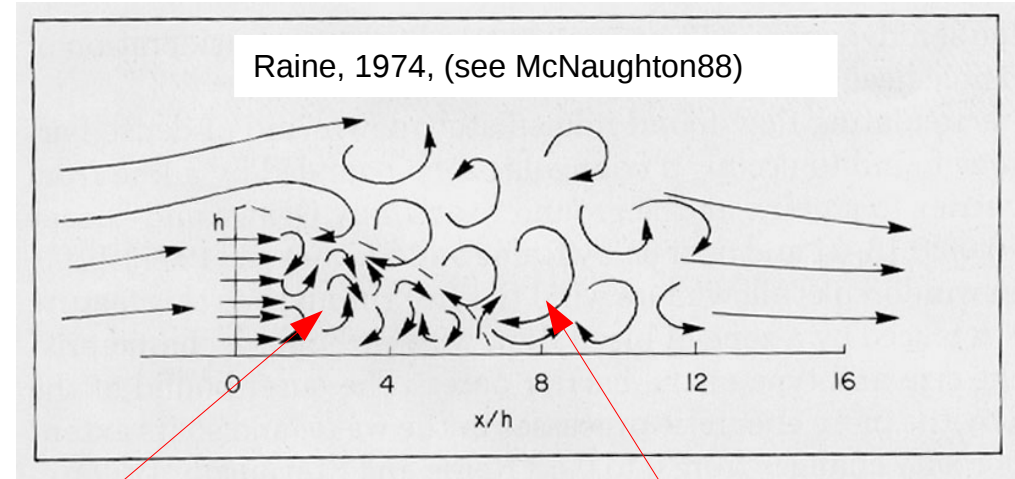
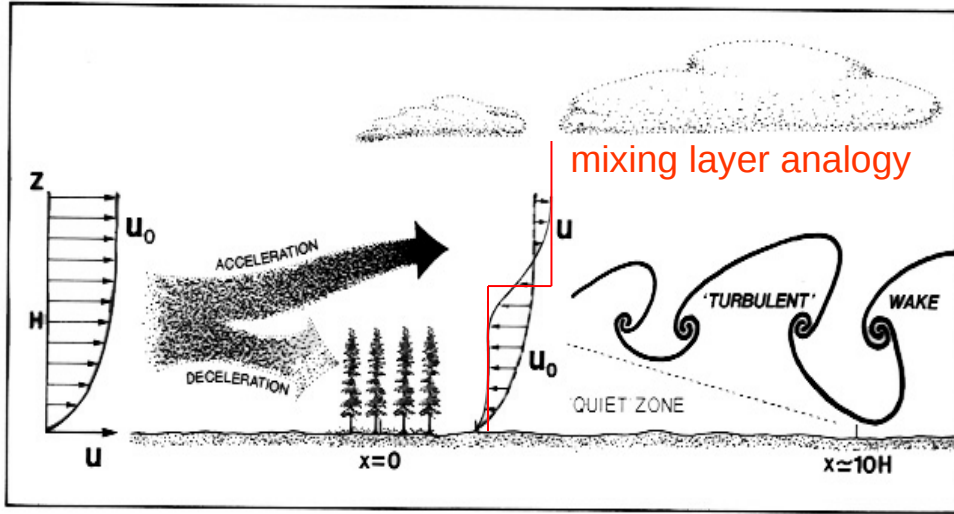


Useful review: pages 17-24 in McNaughton (1988, Agric., Ecosys. & Environ. Vol. 22/23, 17-39)

# Disturbed micro-meteorological flows (ctd): flow around windbreaks



# Overview of effects of windbreak – on mean wind, turbulence, temperature...



- mean wind reduction, turbulent wake

$$\int_{z_0}^{\infty} \bar{u}(z) dz = \int_{z_0}^{\infty} \bar{u}_0(z) dz$$

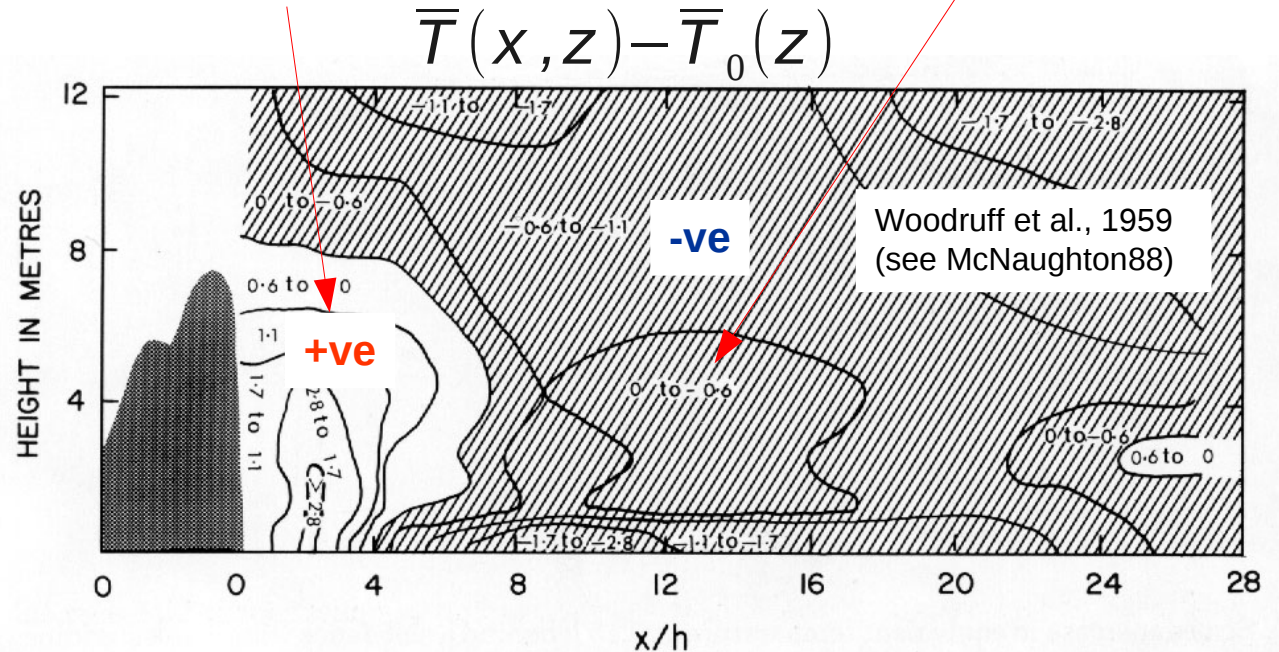
- altered scalar fields

midday summertime anomaly  
in mean temperature [°C],

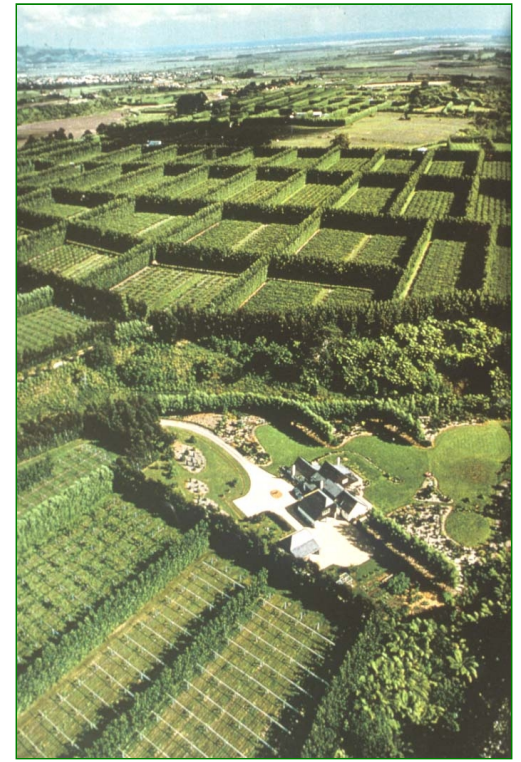
“quiet zone” of reduced TKE

Reduced  $K_h$ , so need  
stronger mean temp  
gradient to carry  $Q_H$

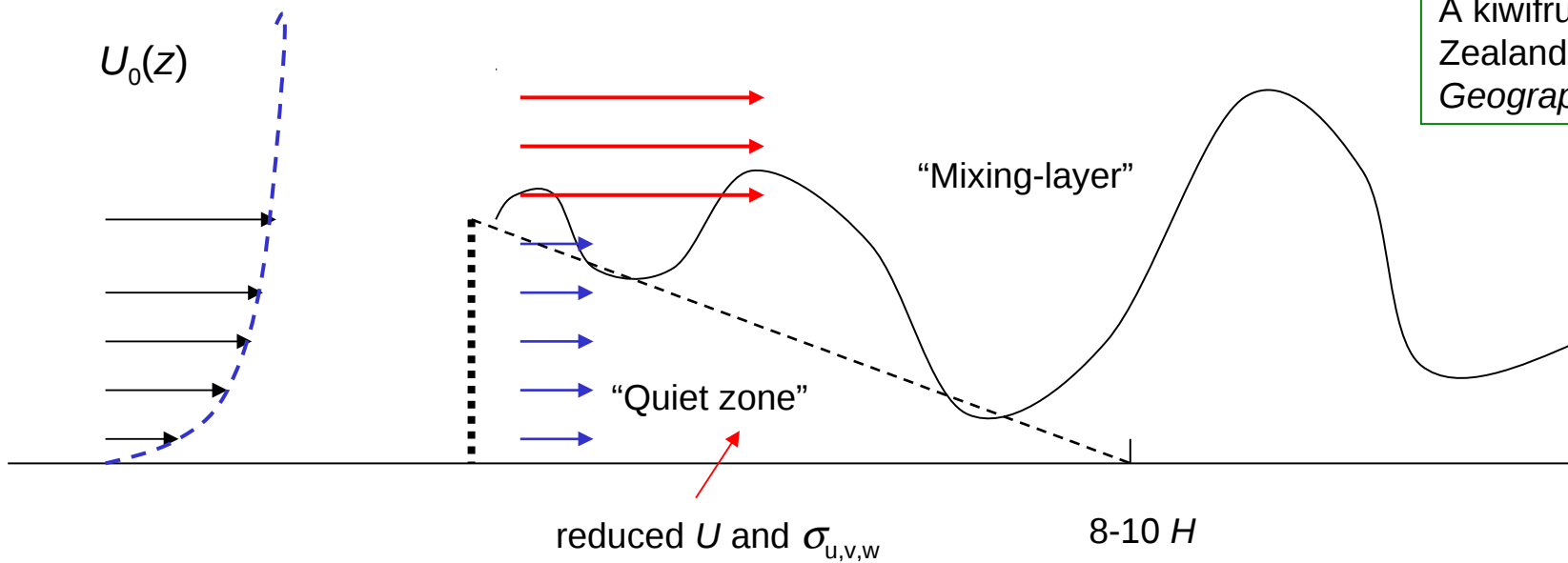
Enhanced  $K_h$



# Overview of effects of windbreak – on mean wind, turbulence, temperature...



A kiwifruit orchard in New Zealand (from *National Geographic*)



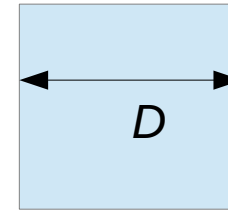
# Overview of effects of windbreak

*Agricultural and Forest Meteorology*, 48 (1989) 185-199  
Elsevier Science Publishers B.V., Amsterdam — Printed in The Netherlands

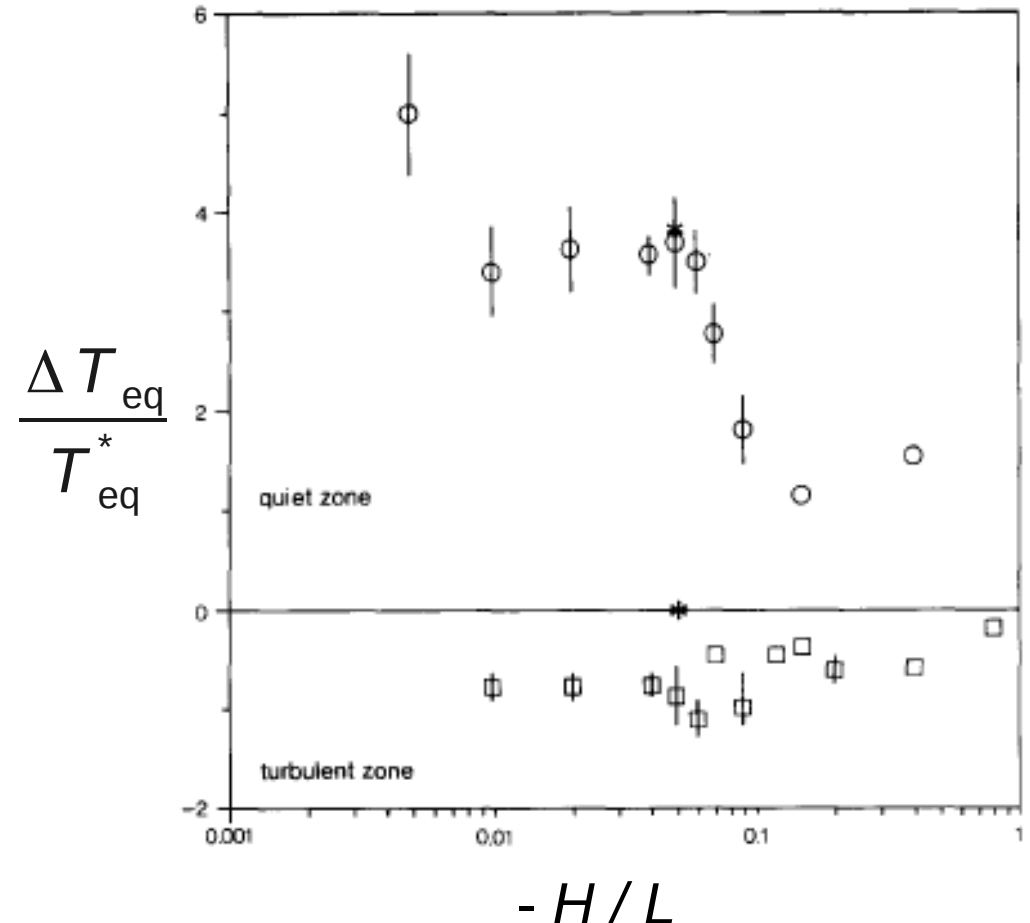
## THE MICROCLIMATE IN THE CENTRE OF SMALL SQUARE SHELTERED PLOTS

J.C. ARGETE\* and J.D. WILSON

- same surface flux of thermodynamic energy  $Q_{H0} + Q_{E0}$  ( $\equiv -\rho c_p u_* T_{eq}$ ) along with reduced eddy diffusivity in quiet zone results in higher  $T_{eq}$
- larger plot size  $D/H=16$  places centre of plot beyond the quiet zone... eddy diffusivity increases in the wake zone



*Normalized difference in mean equivalent temperature between plot centre and same height in the open, for plot widths  $D/H=8$  (circles) and  $D/H=16$  (squares)*

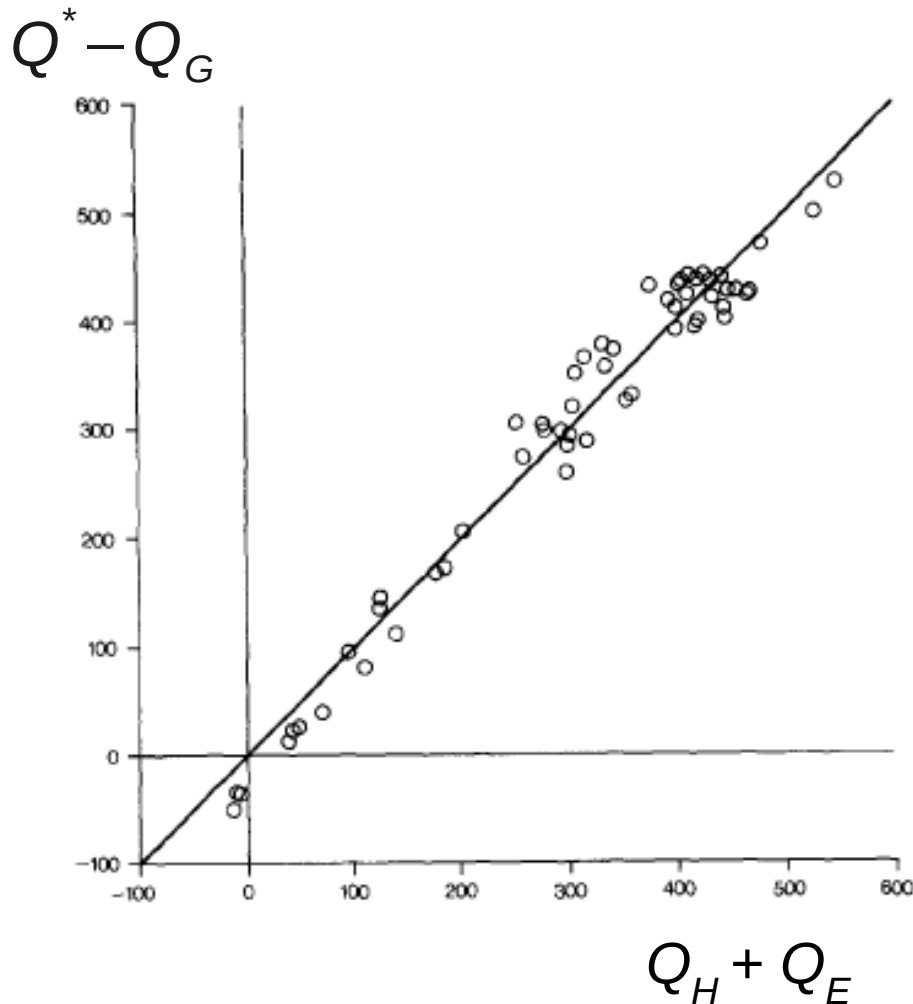


# Digression – quality of fluxes inferred from profiles

*Agricultural and Forest Meteorology*, 48 (1989) 185–199  
Elsevier Science Publishers B.V., Amsterdam — Printed in The Netherlands

## THE MICROCLIMATE IN THE CENTRE OF SMALL SQUARE SHELTERED PLOTS

J.C. ARGETE\* and J.D. WILSON



- fitted MO profiles to measured profiles of  $U$ ,  $T$ ,  $Q$  and inferred  $u^*$ ,  $L$ , and fluxes  $Q_H$ ,  $Q_E$
- measured net radiation  $Q^*$  with net radiometer and  $Q_G$  with soil heat flux plate

# Windbreak flow: theory & observations regarding an idealized case

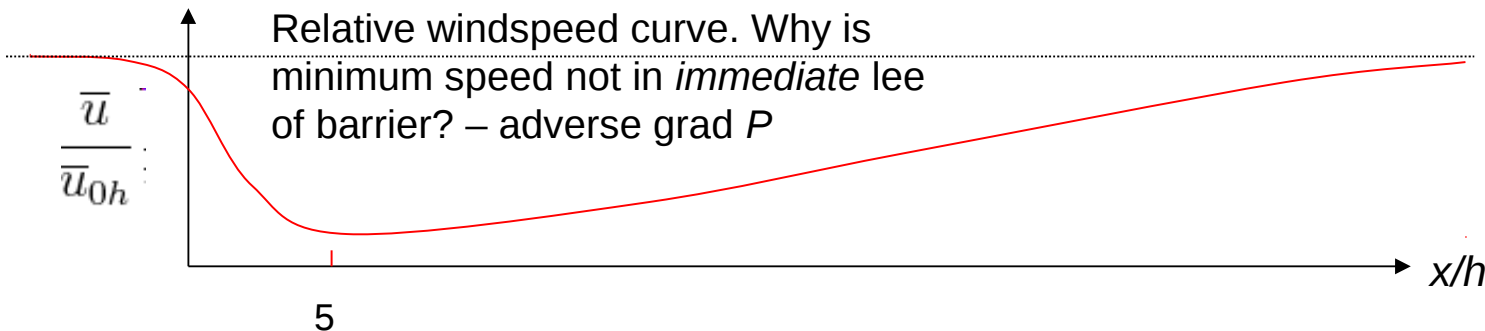
- infinitely long but thin porous barrier (height  $h$  or  $H$ , porosity  $\phi$ ), aligned along  $y$ -axis
- approach flow is neutrally stratified and mean wind direction is normal to the barrier
- by symmetry,  $\bar{v} = 0$  and  $\frac{\partial}{\partial y} = 0$  for any statistic
- things we'd like to be able to anticipate: spatial patterns in

$$\frac{\Delta \bar{u}}{\bar{u}_{0h}} \quad \text{or} \quad \frac{\Delta \bar{u}(x, z)}{\bar{u}_0(z)} \quad (\text{is this strongly height-dependent?})$$

$$\frac{k}{\bar{u}_{0h}^2} \quad \text{or} \quad \frac{k}{u_{*0}^2}$$

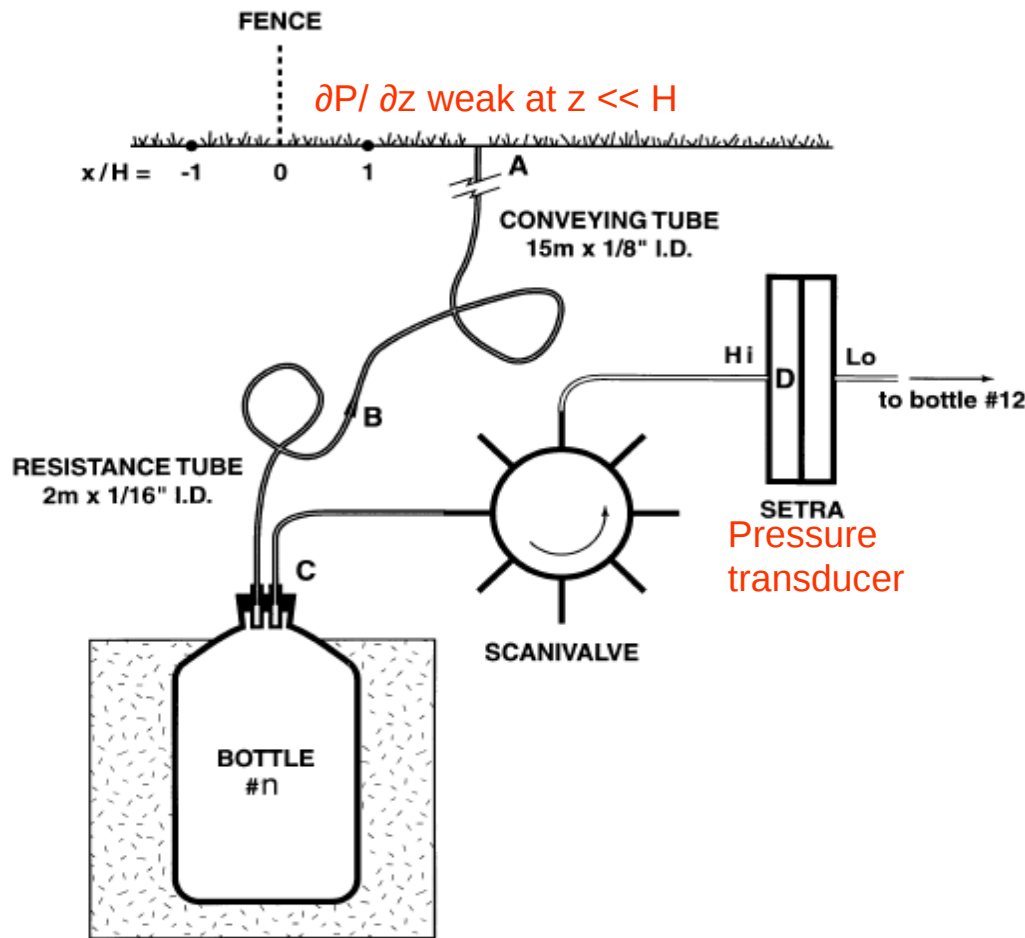
“resistance coefficient” (defined over)

as function of:  $\frac{x}{h}, \frac{z}{h}, \frac{h}{L}, \frac{L}{\delta}, \frac{h}{z_0}, \phi, k_r, \frac{\bar{u}_{0h} h}{\nu}, \dots?$



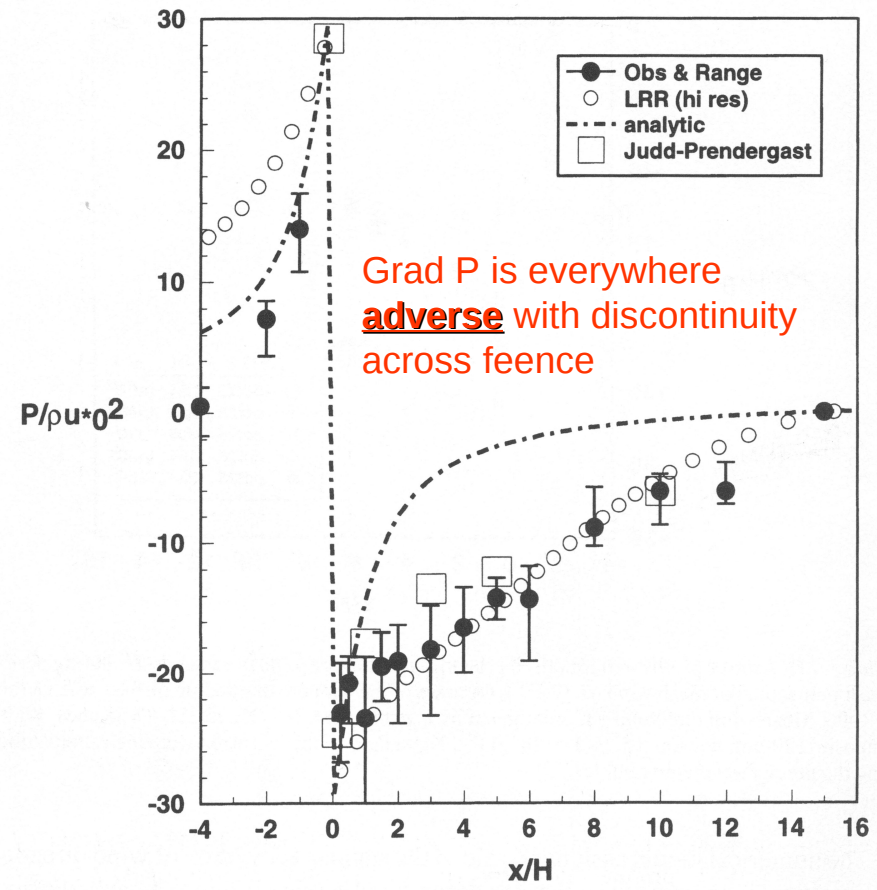
Why the “recovery”? – downward turbulent transfer of  $u$ -momentum from the jet aloft, i.e. due to

$$\frac{\partial \overline{u'w'}}{\partial z}$$



Mean pressure jump across windbreak:  

$$\Delta P \sim 50 \rho u_{*0}^2$$



#### 4. Numerical Simulations

In Section 4 the field observations of pressure and windspeed will be compared with numerical simulations, i.e., solutions of the mean momentum equations (plus the continuity equation, and a turbulence closure). For example the  $\bar{u}$ -momentum equation is:

$$\frac{\partial}{\partial x} \left( \bar{u}^2 + \overline{u'^2} + \bar{p} \right) + \frac{\partial}{\partial z} \left( \bar{u} \bar{w} + \overline{u'w'} \right) = -k_r \bar{u}^2 \delta(x - 0) s(z, H).$$

Localized momentum sink at  $x = 0, z \leq H$ . Proportional to square of speed at barrier, and resistance coefficient  $k_r$ .

# Governing equations – barrier parameterized as momentum sink

- presence of the barrier implies multiply-connected space; formally, need to define flow variables as a suitable area- or volume-average
- interaction of the flow with barrier is not resolved; momentum loss has to be parameterized

$$\frac{\partial}{\partial x} \left( \overline{u^2} + \overline{u'^2} + \overline{p} \right) + \frac{\partial}{\partial z} \left( \overline{u w} + \overline{u' w'} \right) = S_u$$

For a **natural windbreak**, let  $a(x,z)$  be the “drag area density” ( $\text{m}^{-1}$ ) and  $c_d$  the drag coefficient

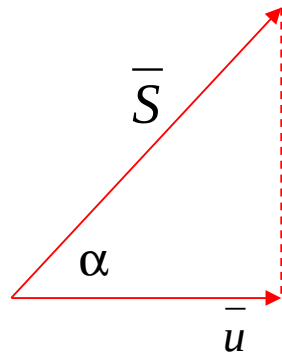
$$S_u = -c_d a(x,z) \overline{u} \sqrt{\overline{u^2} + \overline{v^2} + \overline{w^2}}$$

zero by symmetry restriction
of minor importance

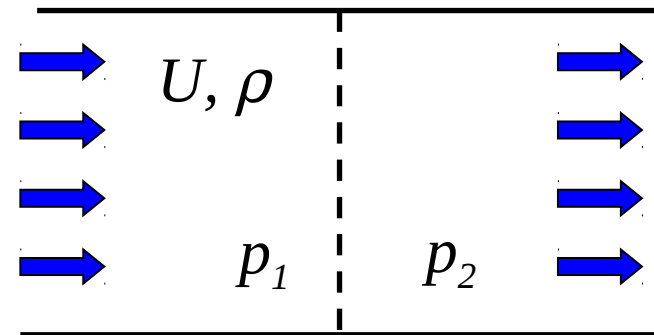
Drag is proportional to projection of  $(\overline{S})^2$  onto x-axis, where

$$\overline{S} \equiv \sqrt{\overline{u^2} + \overline{v^2} + \overline{w^2}}$$

$$(\overline{S})^2 \cos \alpha = (\overline{S})^2 \left( \overline{u} / \overline{S} \right) = \overline{u} \overline{S}$$



Definition of “resistance coefficient” with respect to a uniform stream forced through blocking **porous screen**



$$k_r = \frac{p_1 - p_2}{\rho U^2} \quad (\text{indep of } U \text{ for large } U)$$

$$S_u = -k_r \overline{u} \left| \overline{u} \right| \delta(x-0) s(z-h)$$

Step functn

- treat windbreak as a source of mean velocity deficit  $\Delta \bar{u}$
- treat the velocity deficit as a passive scalar that is advected by the undisturbed wind ( $\bar{u}_0$ ) and diffused by the turbulence (eddy diffusivity  $K_0$ )

$$\frac{\partial}{\partial x} \left( \bar{u}^2 + \overline{u'^2} + \bar{p} \right) + \frac{\partial}{\partial z} \left( \bar{u} \bar{w} + \overline{u'w'} \right) = S_u$$

(kinematic pressure)

neglect

Substitute  $\bar{u} = \bar{u}_0 + k_r \Delta \bar{u}$

$\bar{w} = k_r \Delta \bar{w}$

$\bar{p} = k_r \Delta \bar{p}$

“Perturbation expansion” in small parameter  $k_r$

Solve eqn only in downwind region. Solution is “driven” not by this inhomogeneity (ie. source term), but by an inflow boundary condition

Neglect terms in  $k_r^2$  (i.e. linearize) and write  $\overline{u'w'} = -K \frac{\partial \Delta \bar{u}}{\partial z}$

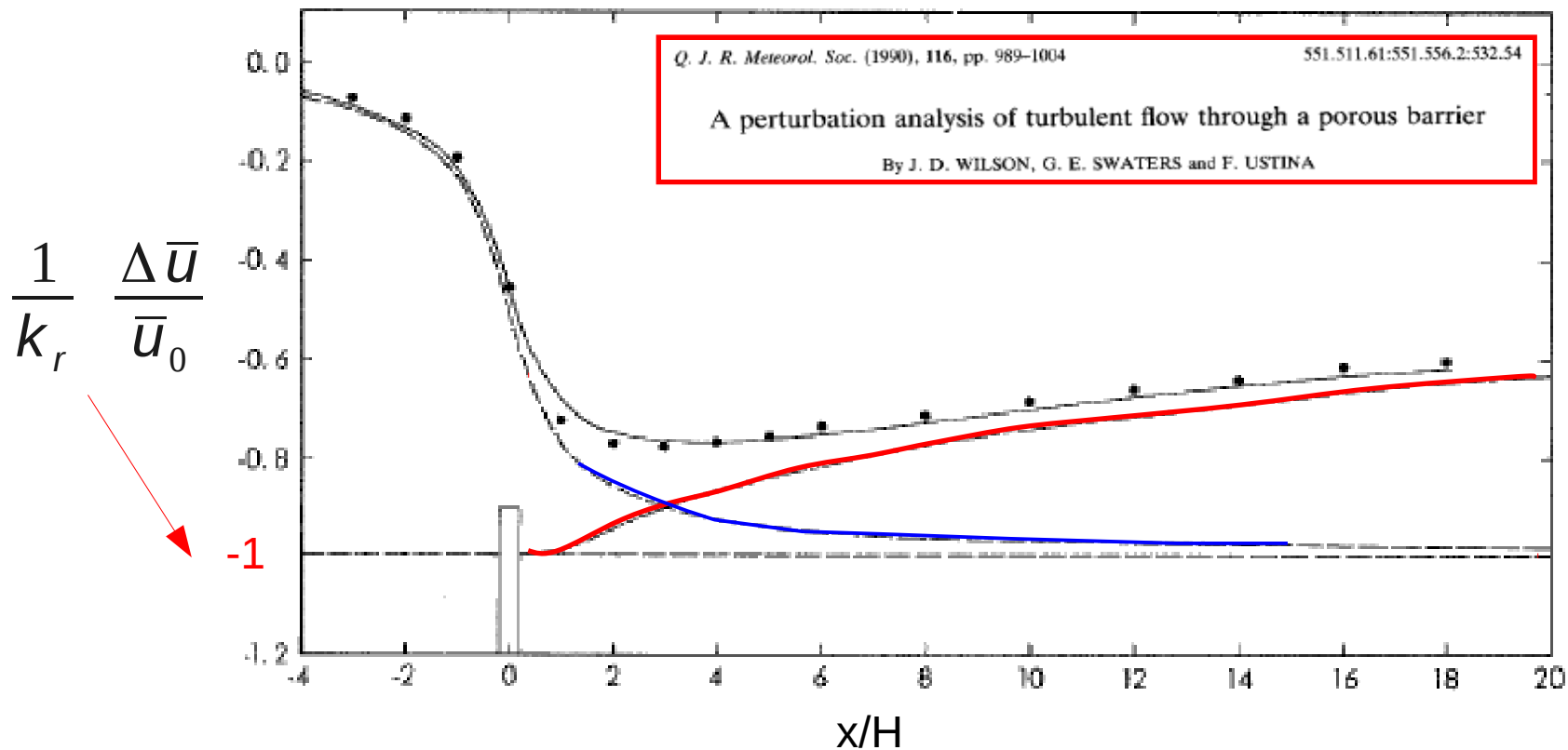
$$\bar{u}_0 \frac{\partial \Delta \bar{u}}{\partial x} + \Delta \bar{w} \frac{\partial \bar{u}_0}{\partial z} = - \frac{\partial \Delta \bar{p}}{\partial x} + \frac{\partial}{\partial z} K \frac{\partial \Delta \bar{u}}{\partial z}$$

Further simplifications:  $\Delta \bar{w} = 0, \partial \bar{p} / \partial x = 0, K = K_0 = \text{const.}, \bar{u}_0 = \text{const.}$

# Kaiser's analytical solution for mean wind speed *downwind* (only) of barrier

– windbreak of height  $h$  represented as collection of strip sources of momentum deficit, each strip of width  $dz$  having strength  $dQ = k_r u_0^2 dz$

$$\frac{1}{k_r} \frac{\Delta \bar{u}}{\bar{u}_0} = -\frac{1}{2} \left[ \operatorname{erf} \left( \frac{h+z}{2\sqrt{x} K_0 / \bar{u}_0} \right) + \operatorname{erf} \left( \frac{h-z}{2\sqrt{x} K_0 / \bar{u}_0} \right) \right]$$



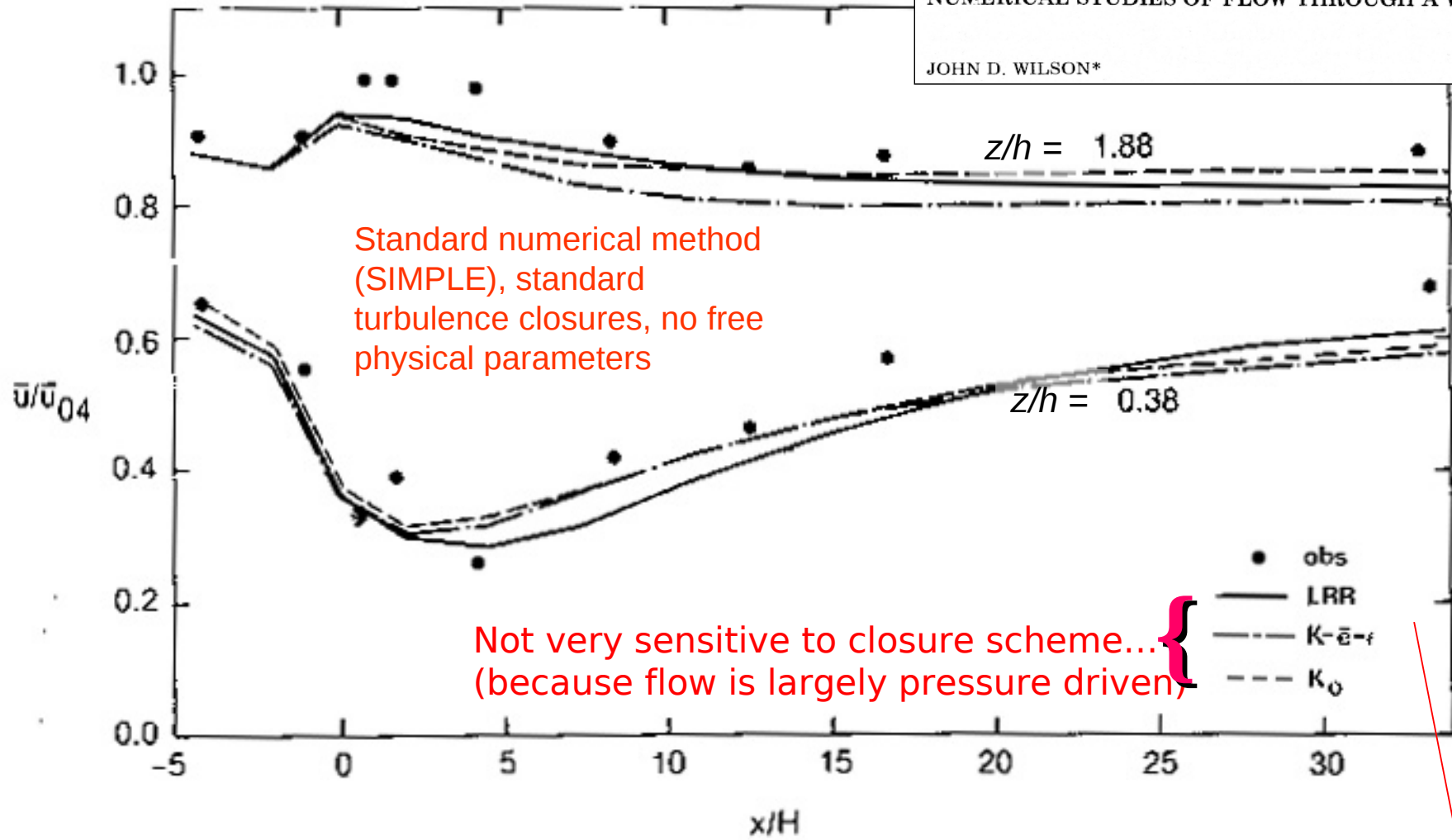
Kaiser's solution necessarily places minimum velocity at the barrier (source of momentum deficit) – unrealistic. Contrast with later analytic solutions that retain  $\text{grad } P$ . The dashed line – no recovery – neglects  $\overline{\partial u'w' / \partial z}$

# Numerical solution – various closures

*Journal of Wind Engineering and Industrial Aerodynamics*, 21 (1985) 119–154  
Elsevier Science Publishers B.V., Amsterdam – Printed in The Netherlands

NUMERICAL STUDIES OF FLOW THROUGH A WINDBREAK

JOHN D. WILSON\*



Bradley-Mulhearn (1983, *J. Wind Eng. Indust. Aerodyn.*, Vol. 15, 145-156)

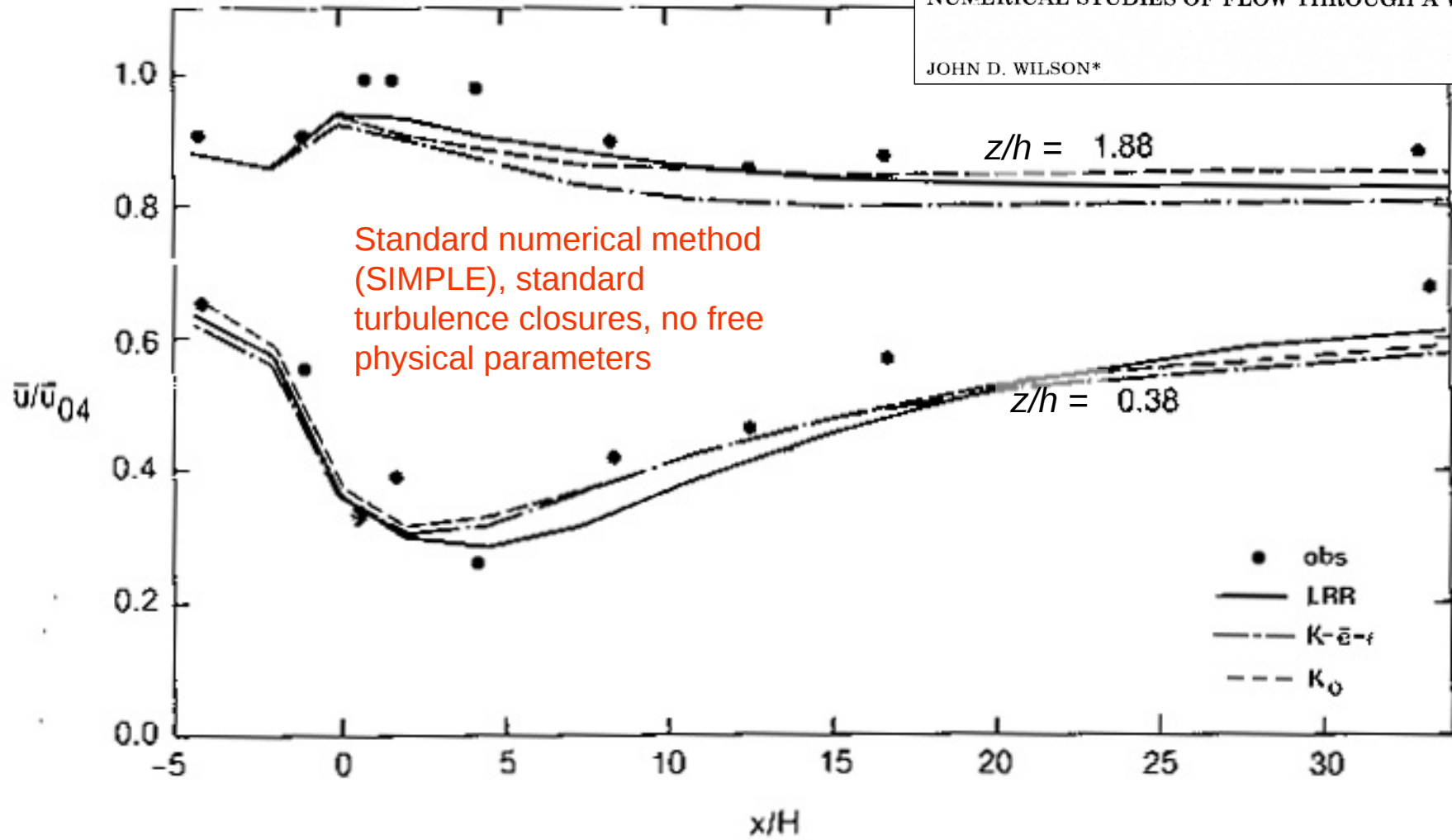
$$k_r = 2, \quad h/z_0 = 600, \quad |L| = \infty$$

$$(h = 1.2 \text{ m}, \quad z_0 = 0.002 \text{ m})$$

Closures :

- (1)  $K_0 = k_v u_{*0} z,$
- (2)  $k - \epsilon$
- (3) LRR second order closure

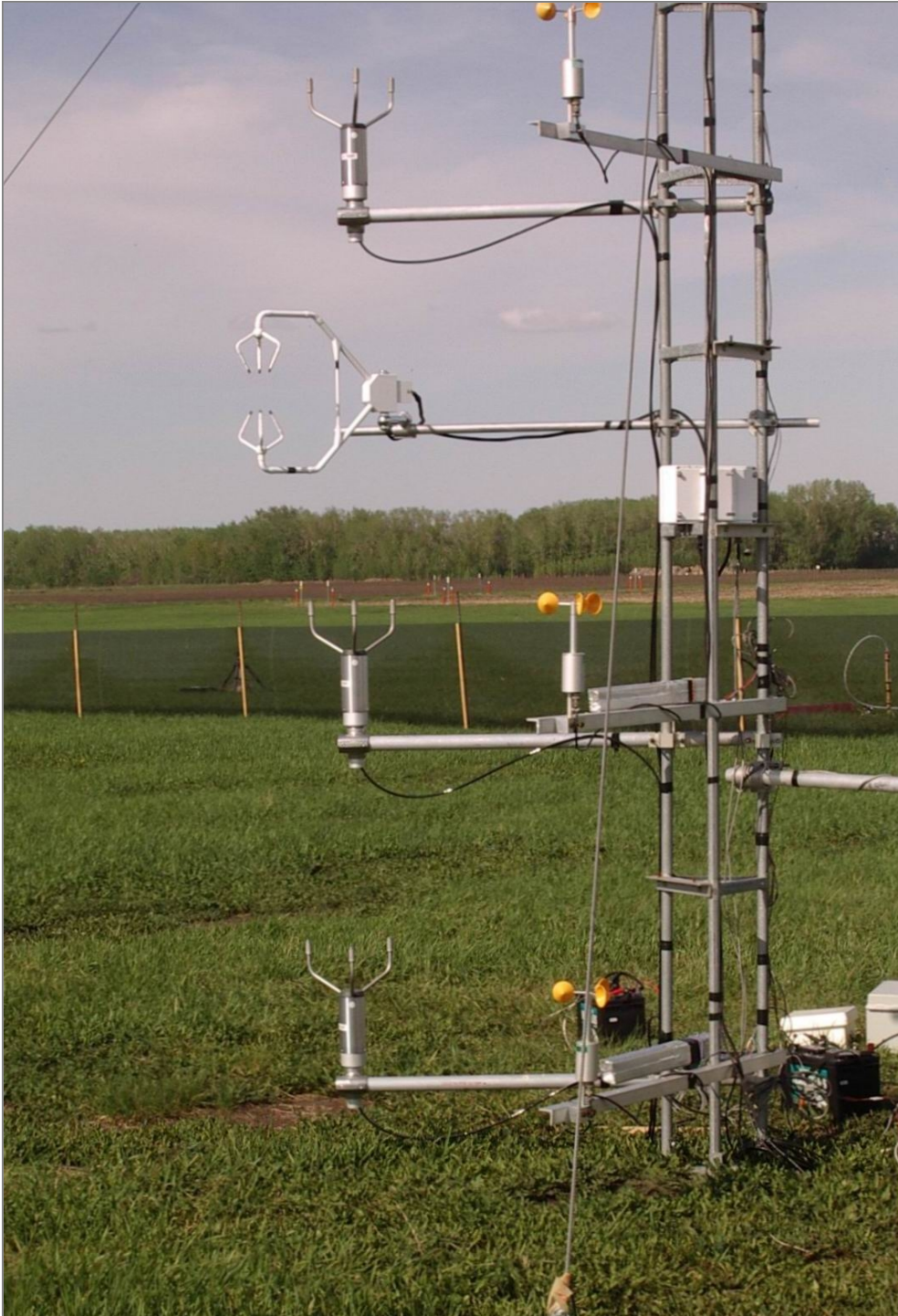
# Numerical solution of mtm eqns



Minimum mean wind speed occurs at about  $5H$  downwind of the barrier, and the fractional reduction in wind speed at that point is:

$$\frac{\Delta \bar{u}}{\bar{u}_0} \approx \frac{k_r}{(1+2k_r)^{0.8}}$$

## Windbreak experiment at Ellerslie



11 cup anemometers  
8 two-D sonic anemometers  
2 three-D sonic anem/thermometers (16 Hz)  
wind vane  
2 thermocouple  $\Delta T$ s

34 wind signals, 4  $T$  signals, 3 dataloggers



# Mean speed... effect of stratification ( $L$ ) in perpendicular flow

## Oblique, Stratified Winds about a Shelter Fence. Part II: Comparison of Measurements with Numerical Models

JOHN D. WILSON

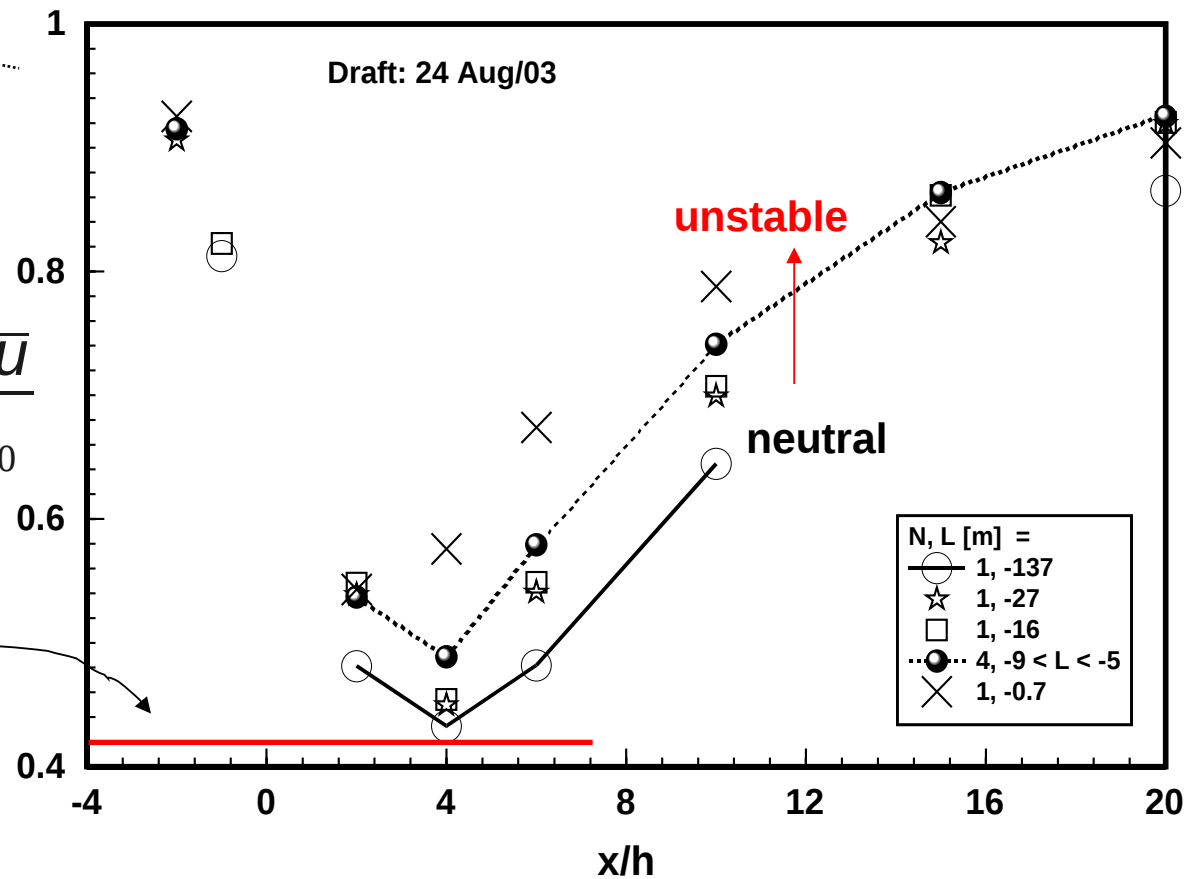


- neutral,  $L < -50$  m
- mod. instability,  $-50 < L < -20$  m
- extrm. instability,  $-5 < L < 0$  m

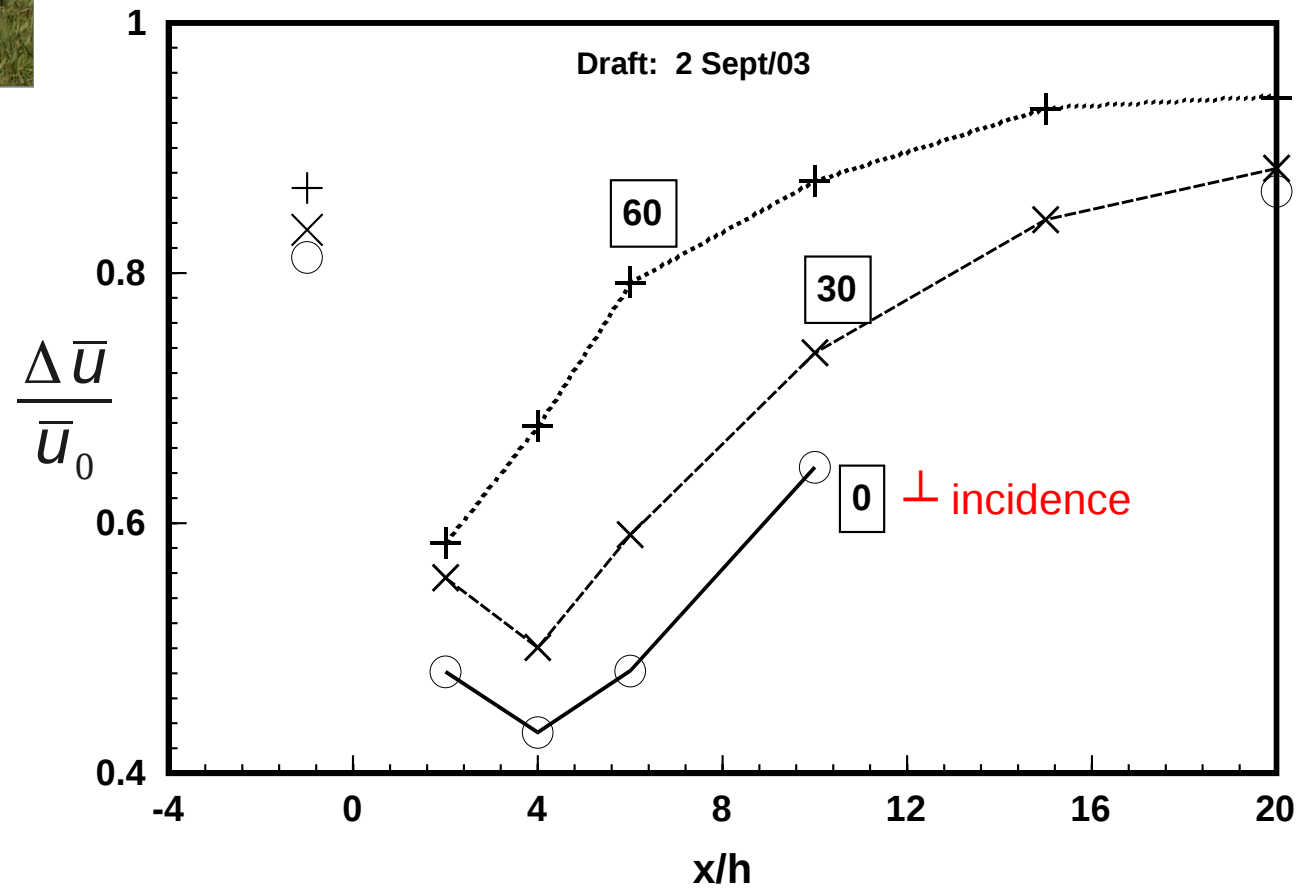
“relative windspeed”

$$\frac{\Delta \bar{u}}{\bar{u}_0}$$

$$\frac{\Delta \bar{u}}{\bar{u}_0} = \frac{k_r}{(1 + 2k_r)^{0.8}}$$



# Mean speed... effect of obliquity in neutrally-stratified winds

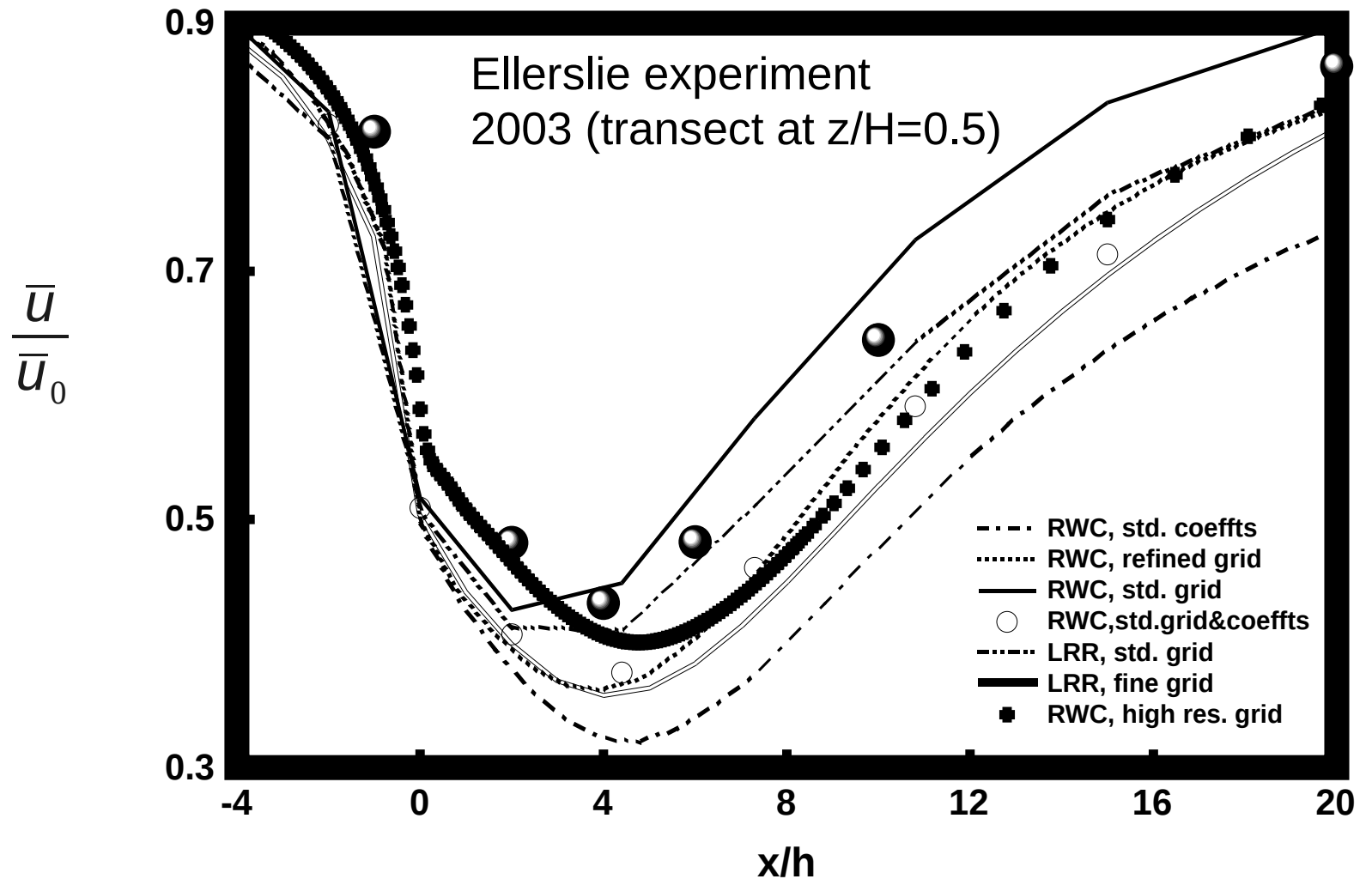


# Computed vs. observed transect - neutral, perpendicular flow...

Domain covers:  $-20 \leq x/h \leq 120$  ,  $z/h \leq 50$

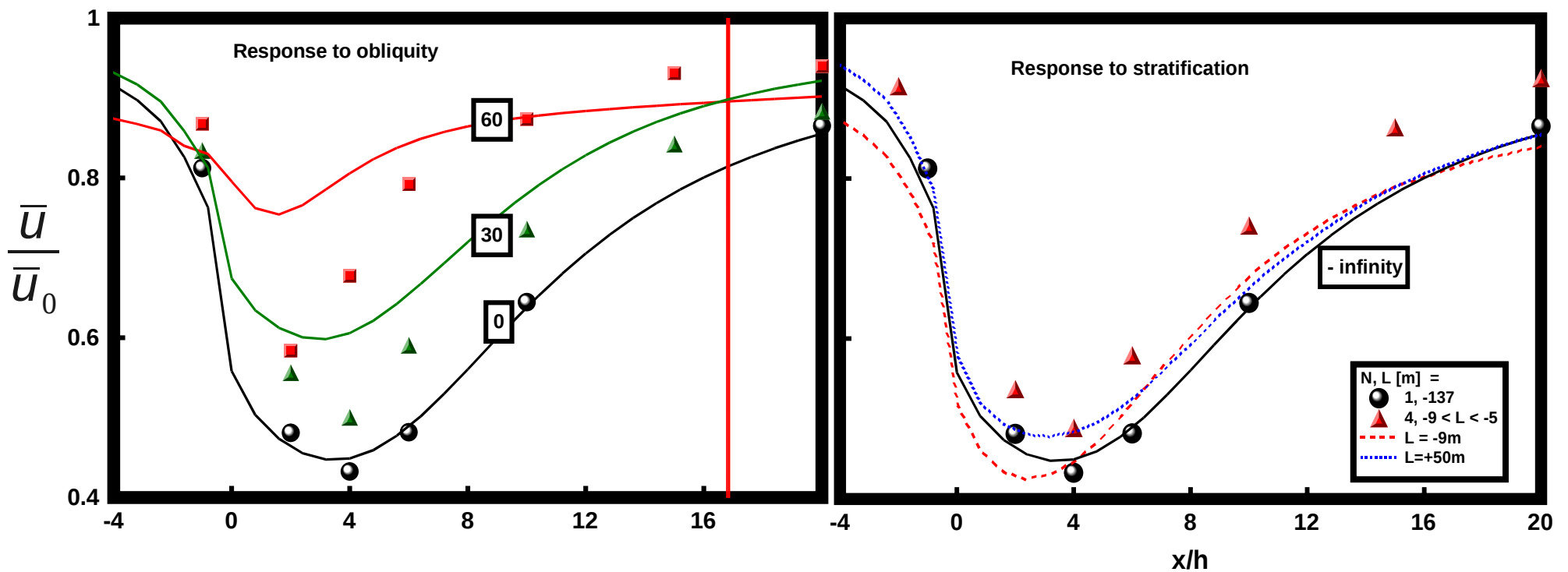
Resolution:  $\Delta x/h \leq 2$  ,  $\Delta z/h \leq 0.25$

Closure: Launder-Reece-Rodi or Rao-Wyngaard-Coté



# Computed vs. observed transects - responses to influence of obliquity and stratification...

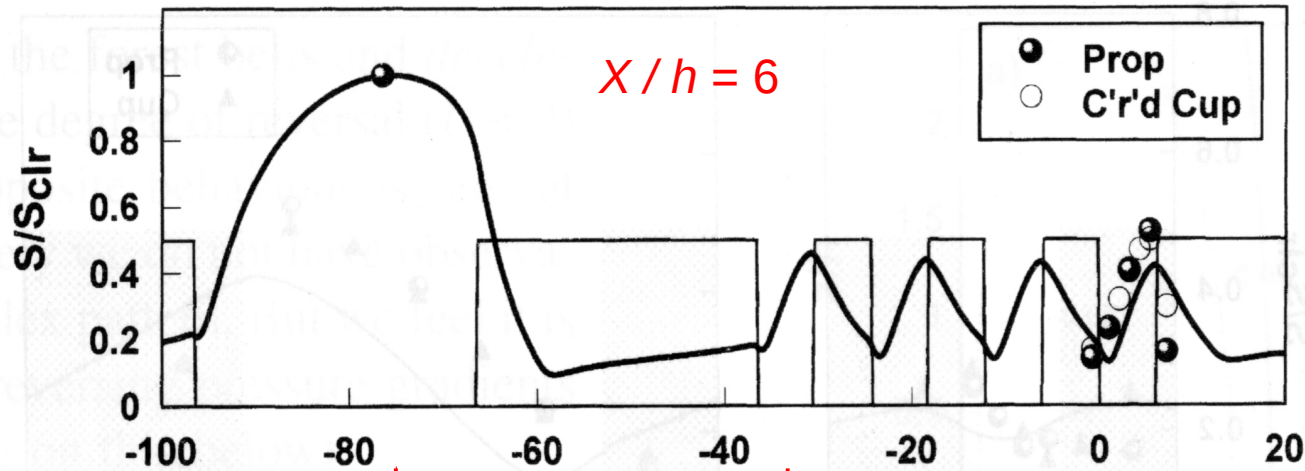
Rao-Wyngaard-Coté closure on a refined grid, with  $k_r = 1.8$  tuned away from experimental value (2.4) so that model's "potential shelter" curve (black) matches observation...



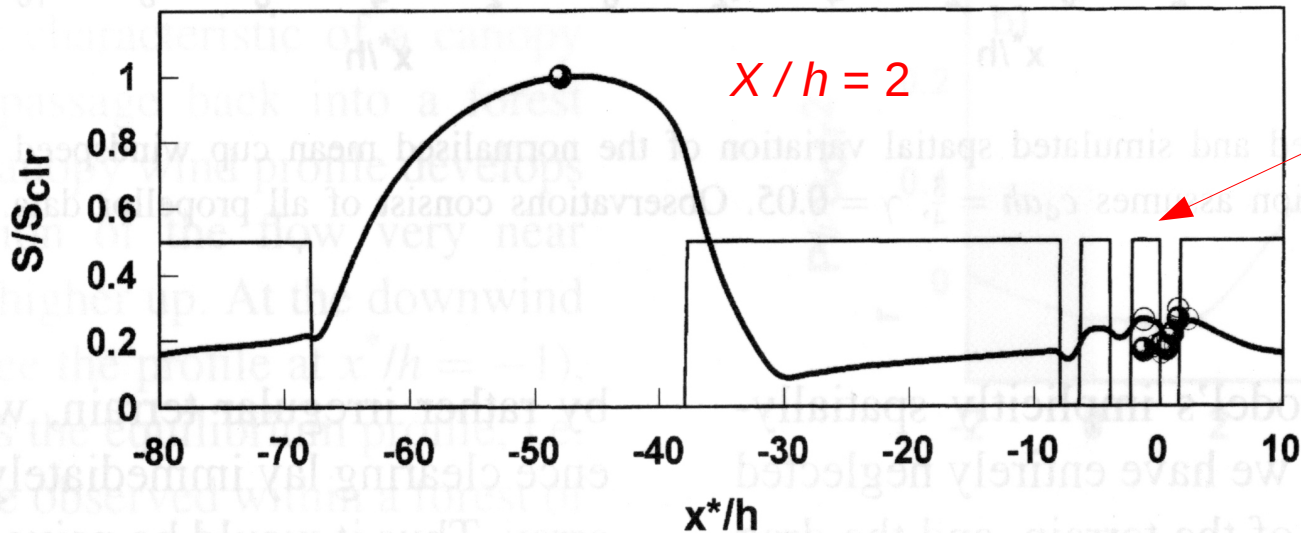
# Measured and modelled mean winds $S/S_{clr}$ in forest cutblocks (width $X$ )

Simpler closure  $K = \lambda(x, z) \sqrt{k(x, z)}$  with prescribed  $\lambda$  and one free parameter ( $c_d a h$ )

Reference speed  $S_{clr}$  is in a far-off (5 km distant) large clearing

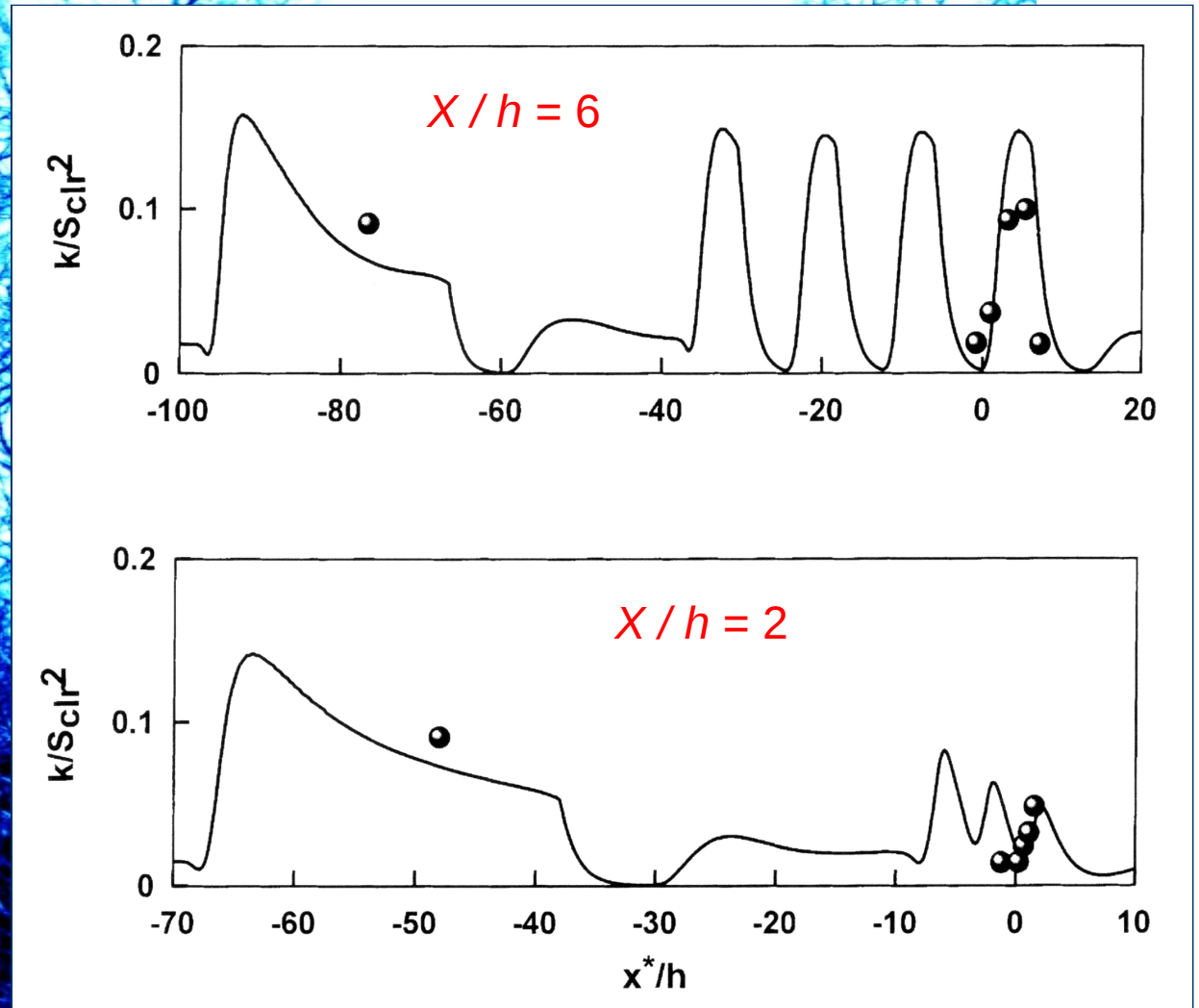


In real world, separated by 5 km of irregular terrain

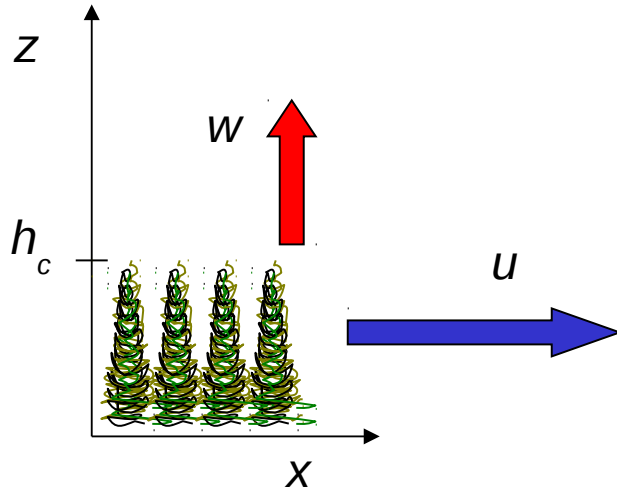


Manning, Alberta

# Measured and modelled TKE in forest cutblocks



## Conclusions:



- there are an infinity of different "disturbed flows"
- representation of turbulence statistics is the limiting factor in their modelling
- higher-order RANS closures more flexible than eddy diffusion scheme (in part as they possess more closure coeffs.); skepticism as to their objective superiority

- Large Eddy Simulation somewhat analogous to NWP – larger scales of motion are explicitly computed

- LES must parameterize effect of subgrid motion; typical closure

$$K = \sqrt{\alpha k} (\Delta x \Delta y \Delta z)^{1/3}$$

- LES burdensome computationally

- seldom compared directly with real horizontally-inhomogeneous flows

