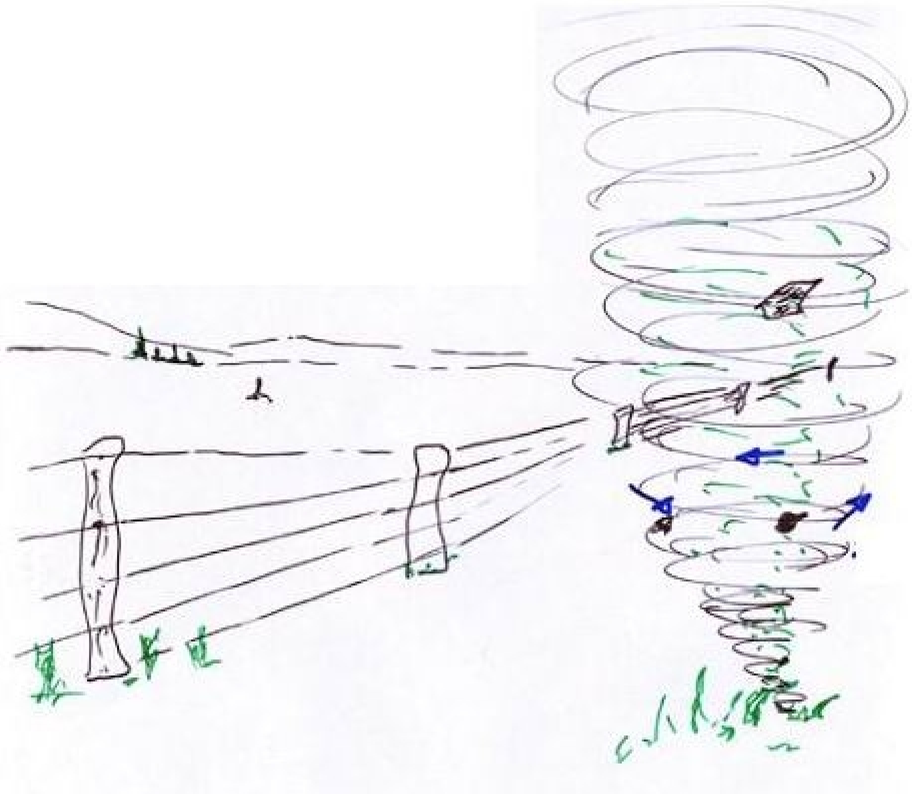
Disturbed micrometeorological flows, and their modelling

- 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 imperfect; LES (Large Eddy Simulation) remains impractical for routine application to disturbed flows



(Modelling) 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

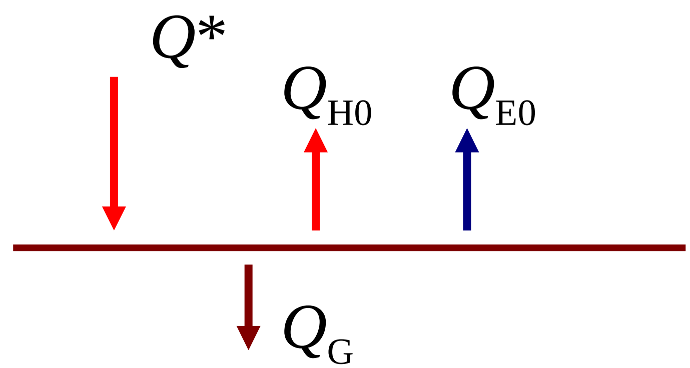
Note: the flow need not be disturbed at the boundary in order to be inhomogeneous

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)

Meaning, they are not MO profiles

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

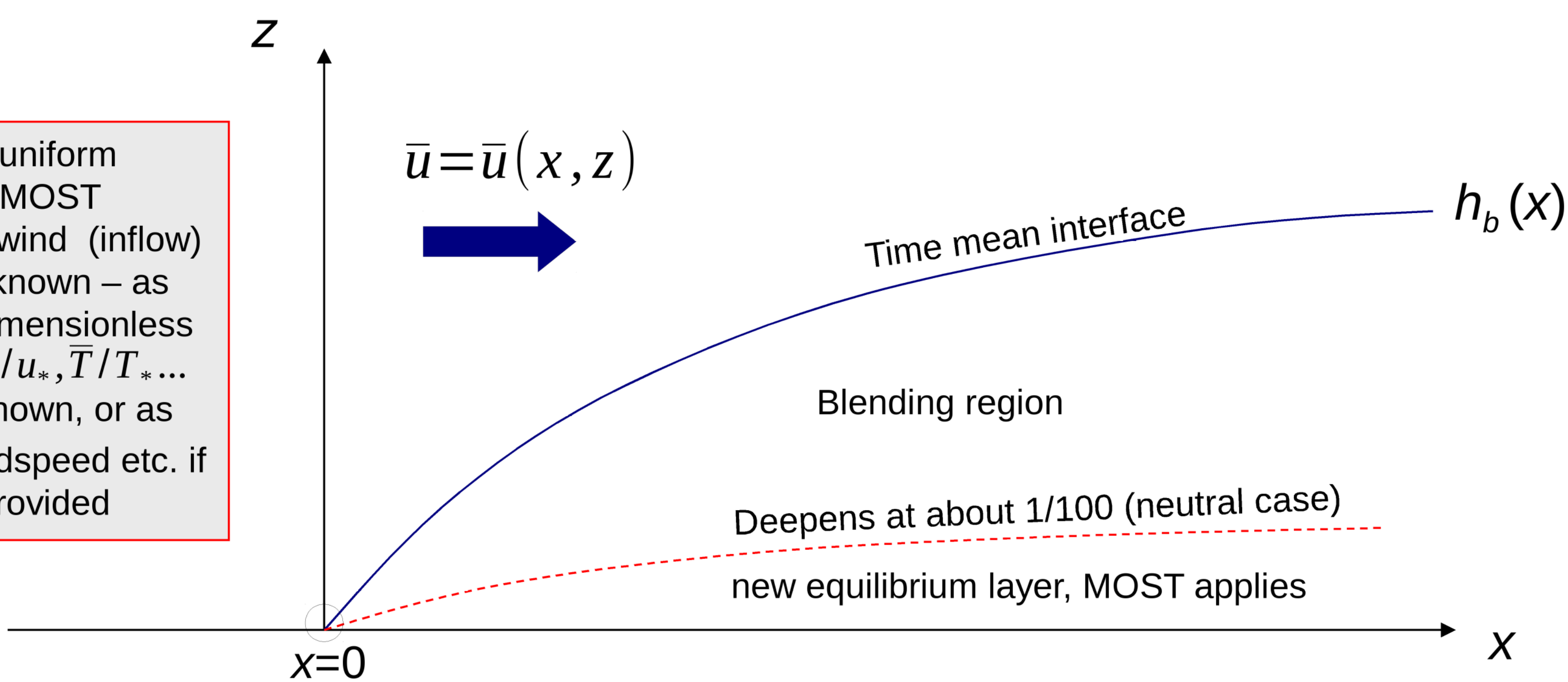


**The surface energy budget

* 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

The paradigm of the Internal Boundary Layer

Horizontally uniform at x<0 ... so MOST applies – upwind (inflow) profiles are known – as profiles of dimensionless properties $\overline{u}/u_*, \overline{T}/T_* \ldots$ if u_* & L unknown, or as physical windspeed etc. if u_* & L are provided



Paradigm:

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

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

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

This Poisson eqn derives from the Reynolds eqns. Solution for mean pressure at point with coordinate **R** responds to r.h.s. over *all* positions **r**, weighted as |**R** – **r** |$^{-2}$. This implies a disturbance has *upstream* influence (x is *not* "one-way")

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

infinite Bowen ratio Q_{H0}/Q_{E0} in the approach flow

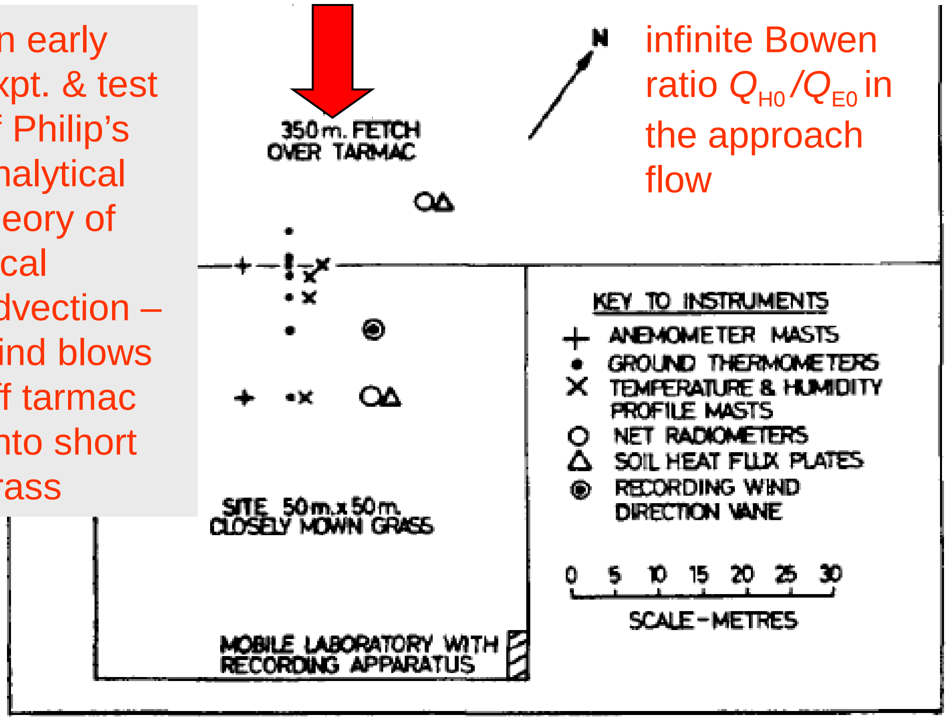
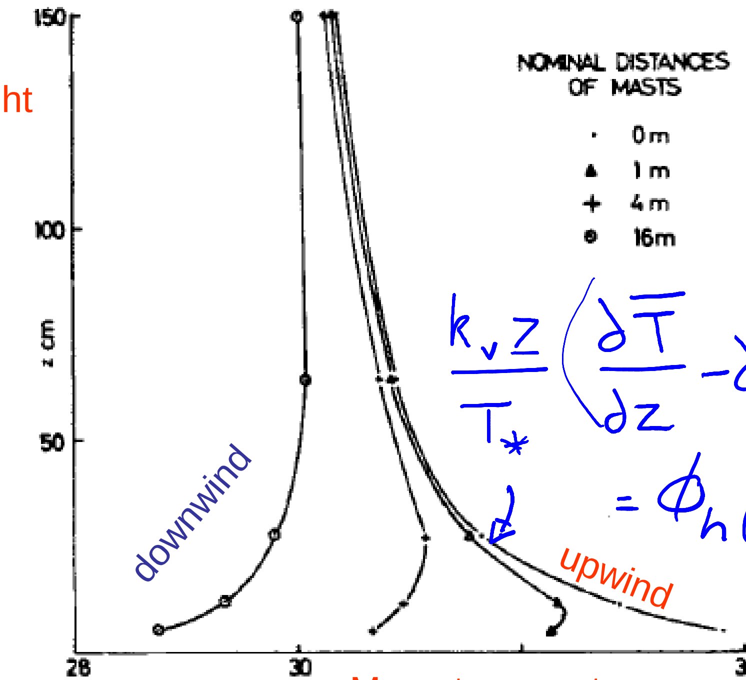


Figure 2. The experimental site, together with the disposition of instruments.

Height [cm]



Mean temperature

$$\frac{k_v z}{T_*}\left(\frac{\partial \overline{T}}{\partial z} - \gamma_d\right) = \phi_h\left(\frac{z}{L}\right)$$

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

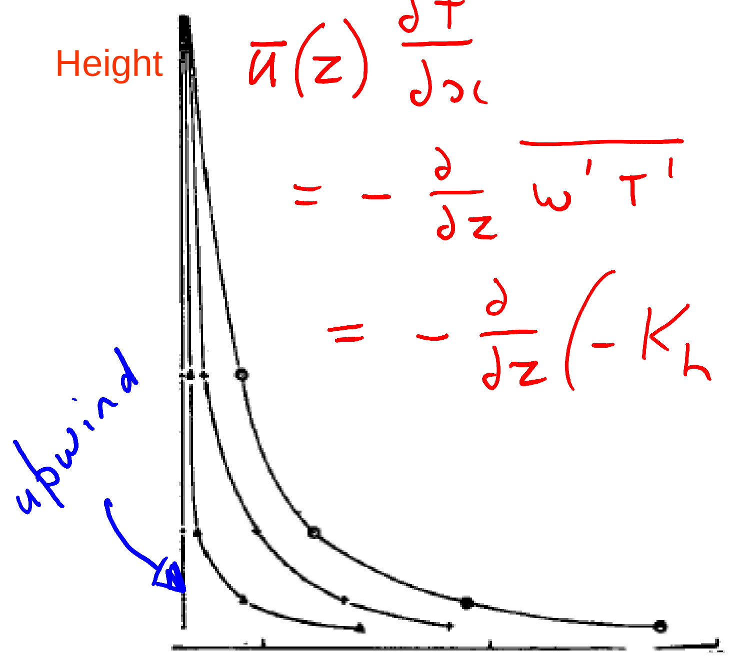
$$u\frac{\partial T}{\partial x} = \frac{\partial}{\partial z}\left[K_T\frac{\partial T}{\partial z}\right] \quad (1)$$

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

Interpretive paradigm = eddy diffusion; steady state; divergences of sensible and latent heat flux vectors vanish; imposed power law profiles of mean wind speed "u" and diffusivity "K_T, K_e"; these properties held independent of x , i.e. $u=u(z)$ etc.

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

Height



Absolute humidity ρ_v

$$\overline{u}(z)\,\frac{\partial \overline{T}}{\partial x} = -\frac{\partial}{\partial z}\overline{w'T'} = -\frac{\partial}{\partial z}\left(-K_h\frac{\partial \overline{T}}{\partial z}\right)$$

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 :

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

Incoming terrestrial (longwave) radiation

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

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

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)

Outgoing terrestrial (longwave) radiation

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

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 :

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

assumes the downwind surface is saturated

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" $\gamma \equiv \dfrac{p\ c_p}{0.622\ L_v}$, and T_{eq} the temp. if all latent heat converted to sensible heat

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

e is vapour pressure

$$Q_H + Q_E = -\ \rho_0 c_p K \frac{\partial \overline{T}_{eq}}{\partial z}$$

LHS constrained at gnd by sfc energy balance

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

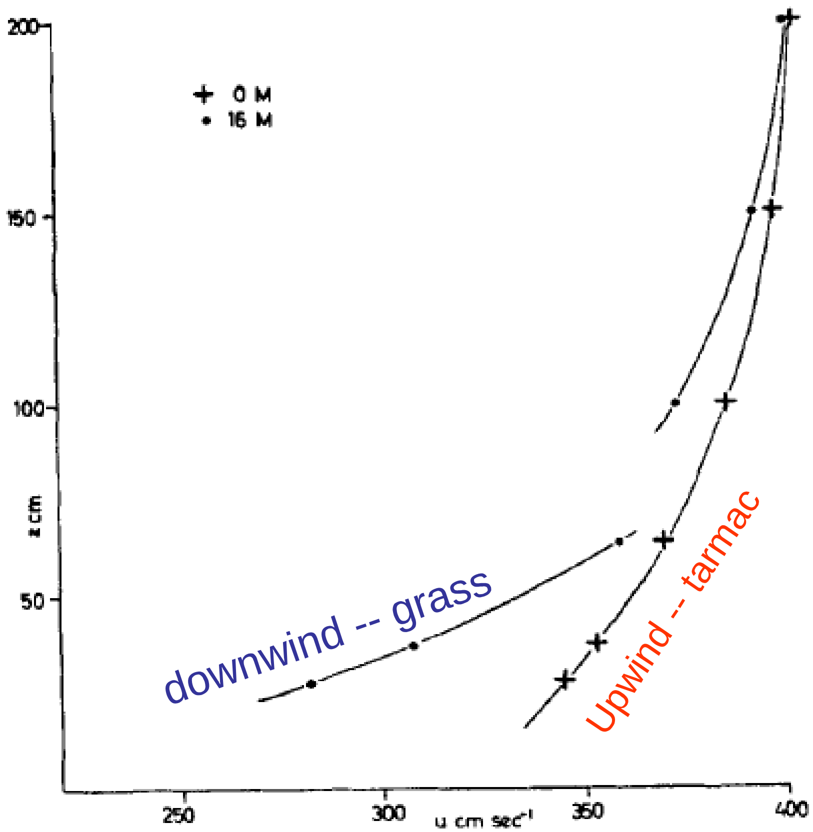


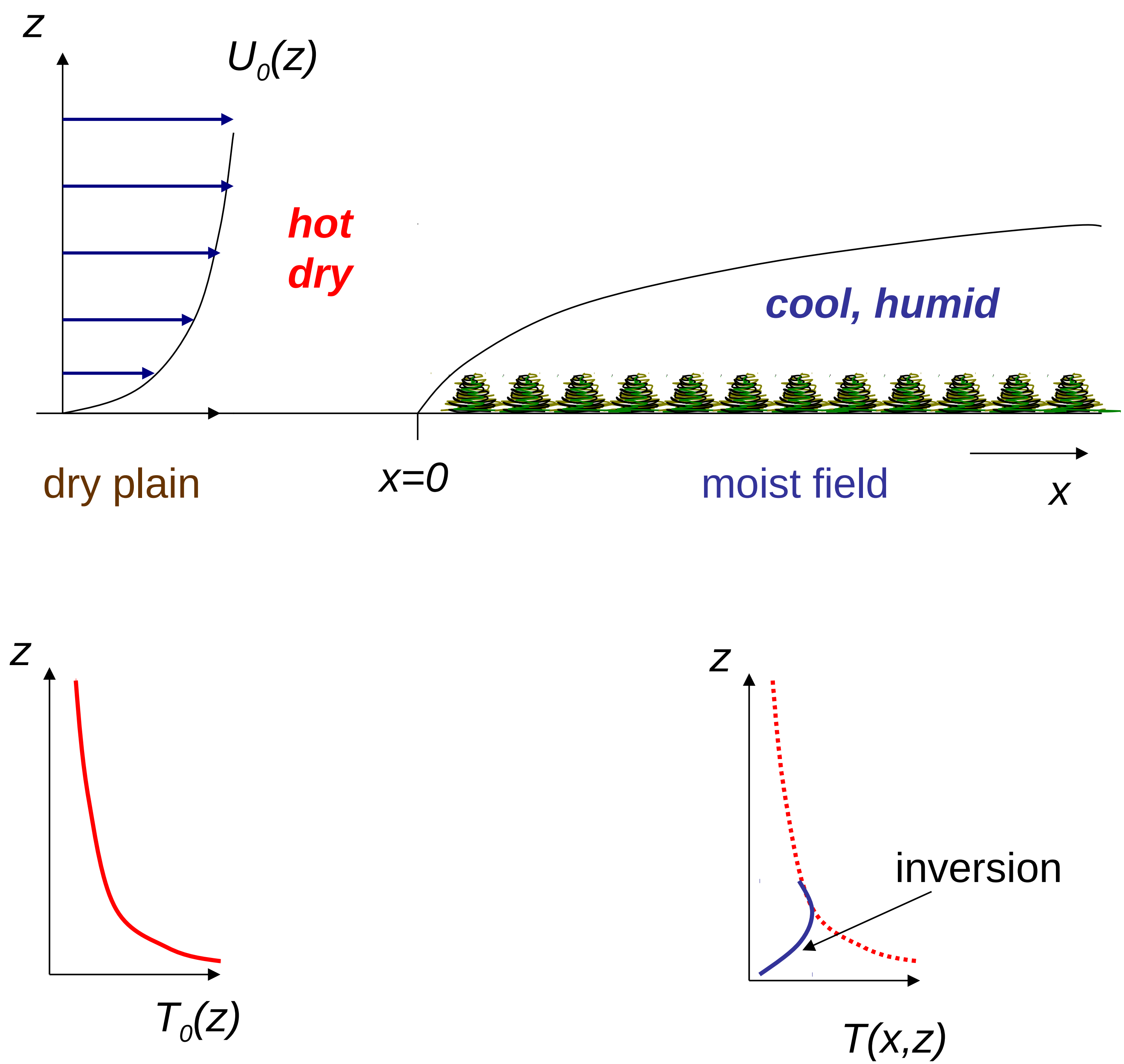
Figure 5. Typical example of the vertical profiles of wind speed at $x = 0$ and $x = 1600$ (observation No. 11).

Observed horizontal variation of the profile of the (advecting) mean wind speed implies one should extend the model by accounting for the disturbed momentum budget

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

$$Q_{H0} + Q_{E0} = \text{const.} = -\rho_0 c_p \left[K \frac{\partial \overline{T}_{eq}}{\partial z} \right]_0$$

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):

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

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

U-mtm:

$$\underbrace{\frac{\partial}{\partial x}\left[U\,U + \overline{u'^2}\right]}_{\textbf{hor.flx.}} + \underbrace{\frac{\partial}{\partial z}\left[W\,U + \overline{w'u'}\right]}_{\textbf{vrt.flx.}} = -\frac{1}{\rho_0}\frac{\partial P}{\partial x}$$

pressure disturbance (not allowed for in original RWC treatment)

σ_u^2: $\equiv \overline{u'^2}$

hor.flx. **vert.flx.**

effective diffusivity

shear production

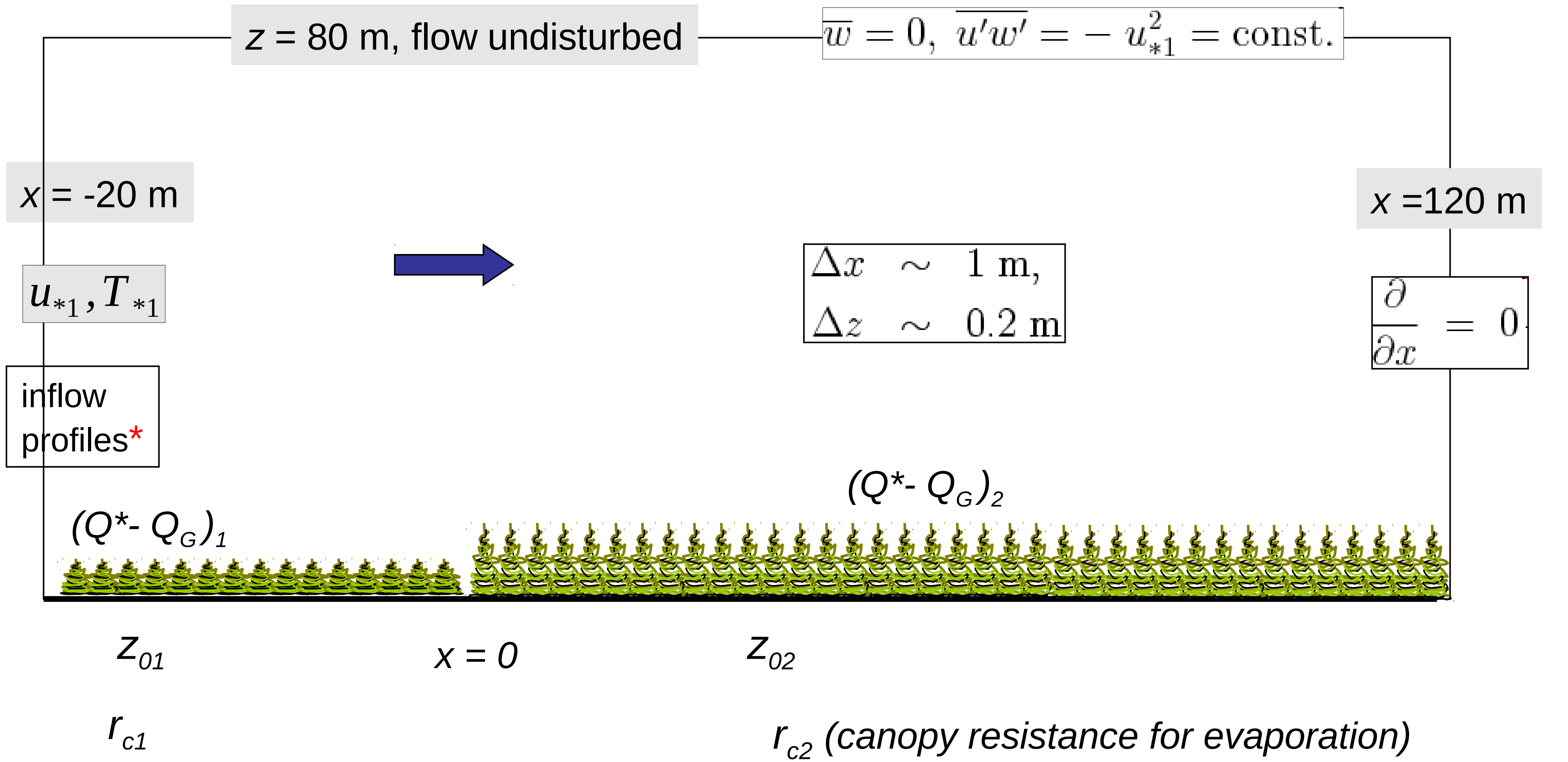
$$\frac{\partial}{\partial x}\Big[\underbrace{U\,\overline{u'^2}}_{\textbf{adv.}} - \underbrace{a_t\,\tau\,\overline{u'^2}\,\frac{\partial \overline{u'^2}}{\partial x}}_{\textbf{diff.}\ \ \overline{u'^3}}\Big] + \frac{\partial}{\partial z}\left[W\,\overline{u'^2} - a_t\,\tau\,\overline{w'^2}\,\frac{\partial \overline{u'^2}}{\partial z}\right] = \underbrace{-2\overline{u'^2}\,\frac{\partial U}{\partial x} - 2\overline{u'w'}\,\frac{\partial U}{\partial z}}_{\textbf{shear production}} - \ldots$$

$$\ldots \underbrace{-\frac{2}{3}\epsilon}_{\textbf{dissip'n}} \underbrace{-\frac{c_{11}}{\tau}\left[\overline{u'^2} - \frac{2}{3}k\right]}_{\textbf{redistrib.}} + \underbrace{\frac{\partial}{\partial x}\left[a_t\,\tau\,\overline{u'w'}\,\frac{\partial \overline{u'^2}}{\partial z}\right] + \frac{\partial}{\partial z}\left[a_t\,\tau\,\overline{u'w'}\,\frac{\partial \overline{u'^2}}{\partial x}\right]}_{\textbf{transport}}$$

$$\tau = \frac{2k}{\epsilon}$$ 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 2nd-order closure model to La Crau experiment:



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Agricultural and Forest Meteorology 107 (2001) 207–225
AGRICULTURAL AND FOREST METEOROLOGY
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Micro-meteorological methods for estimating surface exchange with a disturbed windflow

John D. Wilson[a,*], Thomas K. Flesch[a], Lowry A. Harper[b]
[a] Department of Earth and Atmospheric Sciences, University of Alberta, Edmonton, Alta., Canada T6G 2E3
[b] United States Department of Agriculture, Watkinsville, GA, USA

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

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,

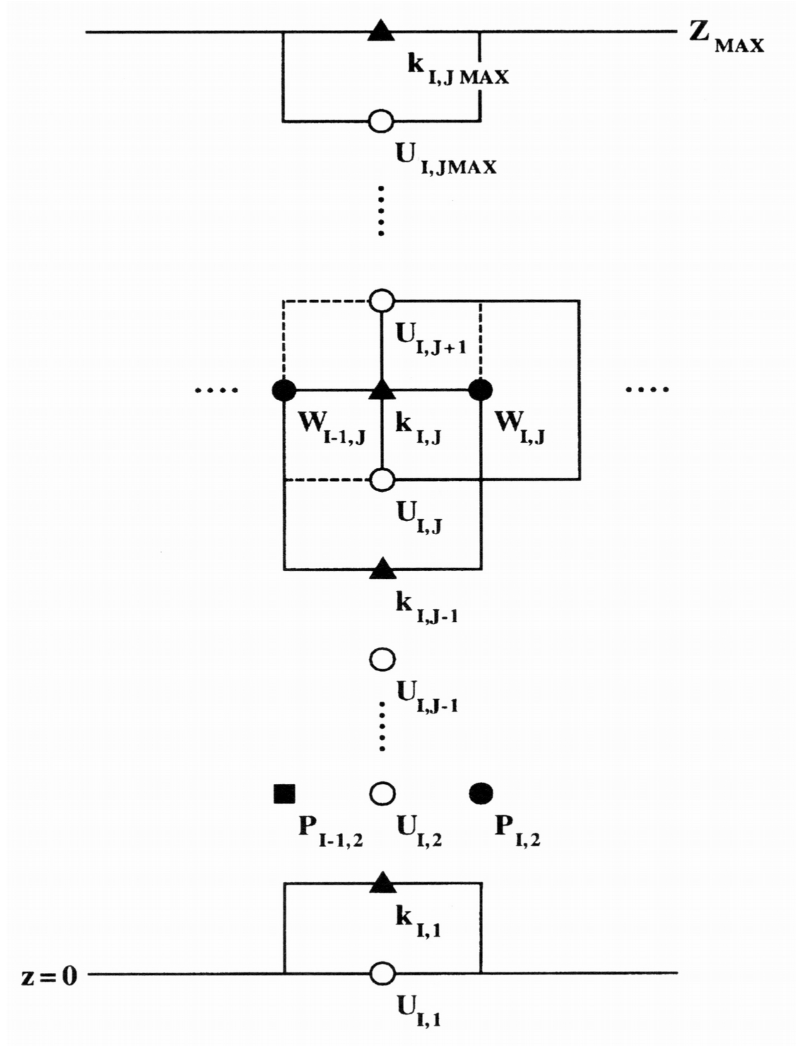
e.g. after obtaining new guess *m*+1, modify it to

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

with relaxation parameter α<1

- continue to iterate until momentum is conserved, both

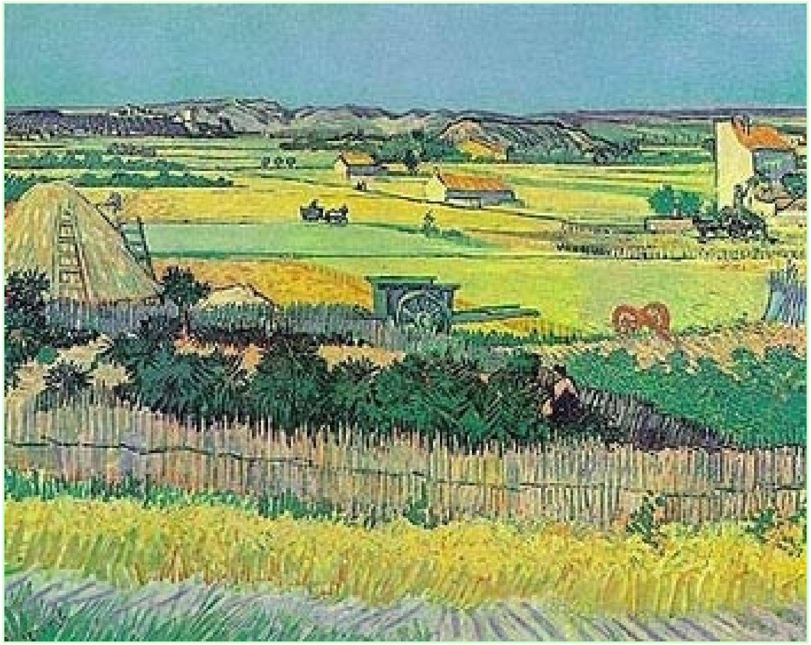
cell by cell and globally (i.e. for the domain in total)



La Crau run 42 – specification of controlling boundary conditions

$$T_{*1} = \frac{-Q_{H1}}{\rho\ c_p\ u_*} \approx \frac{-362}{1 \times 1000 \times 0.6} \approx -0.6^{\circ}$$

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



Vincent Van Gogh: "Harvest at La Crau"

Surface treated as a "big leaf" and coupled to model atmosphere's lowest plane of gridpoints (at $z = z_0 + \Delta z \sim 0.2\ \mathrm{m}$) 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

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

(alt. to $Q_H = -\rho c_p \left(K \frac{\partial \overline{T}}{\partial z} - \gamma_d\right)$)

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

$$K^{MO} = \frac{k_v u_* z}{\phi(z/L)}$$

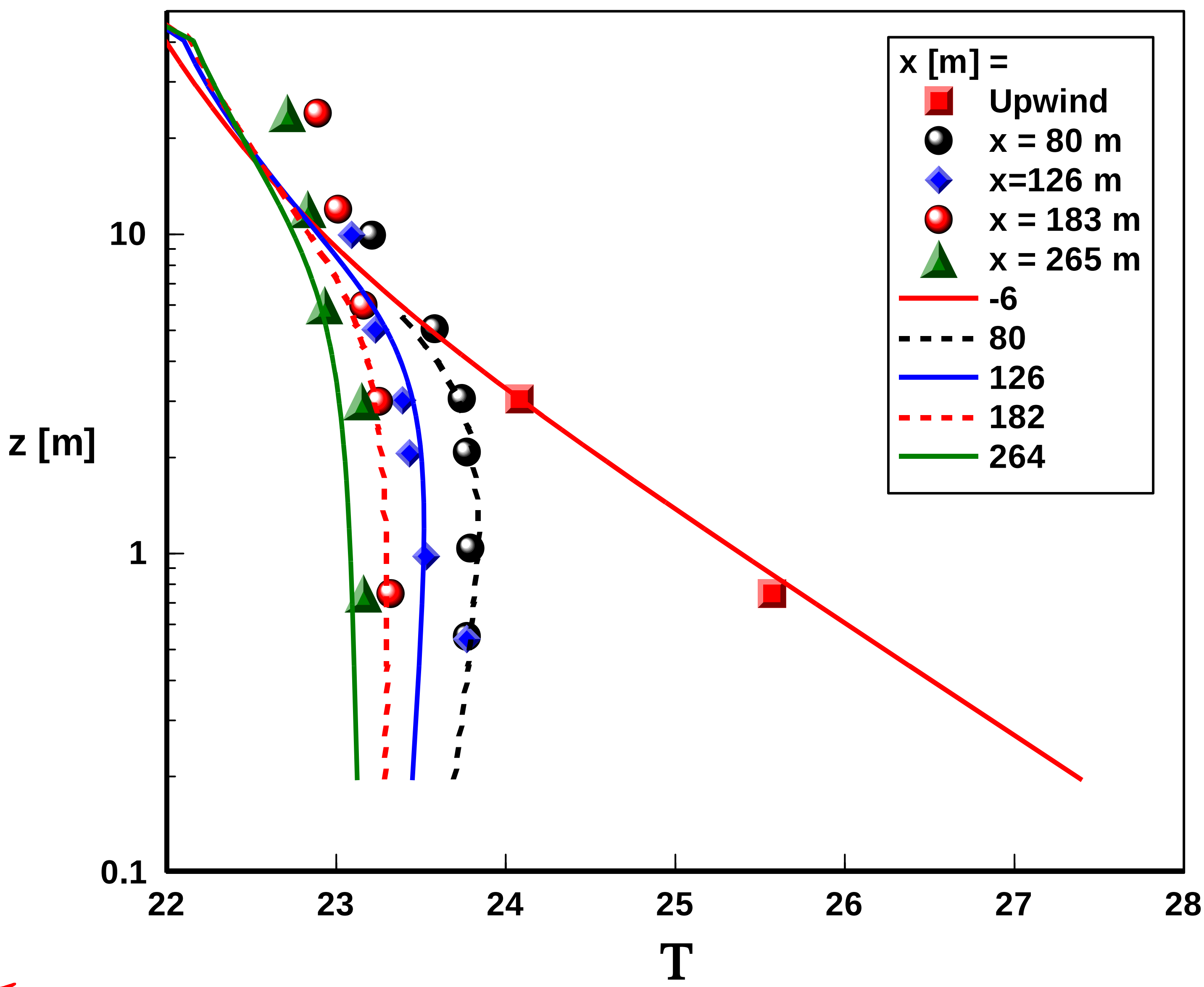
We can use MOST (entailing the assumption of height-independent flux) to calibrate the resistance:

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

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

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



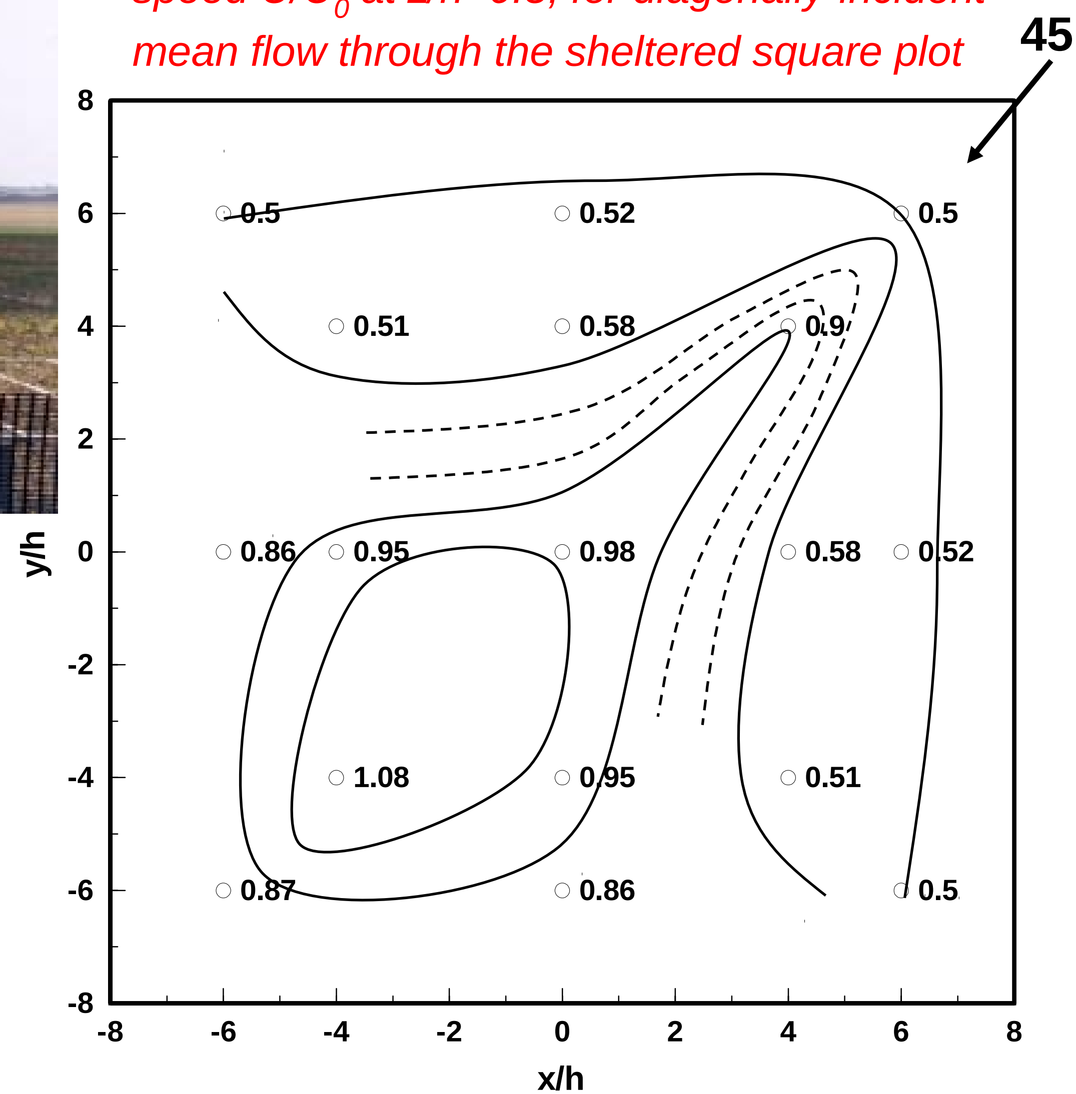
STOP HERE 22 MAR. 2016

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

- basic effects observed

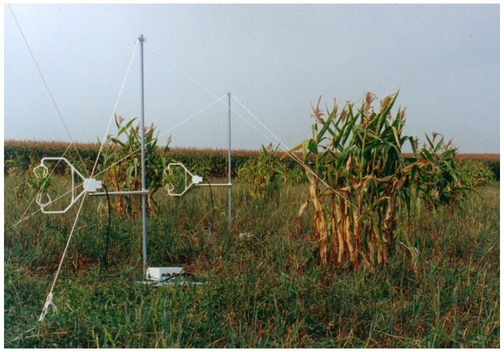
- elements of the theoretical description

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

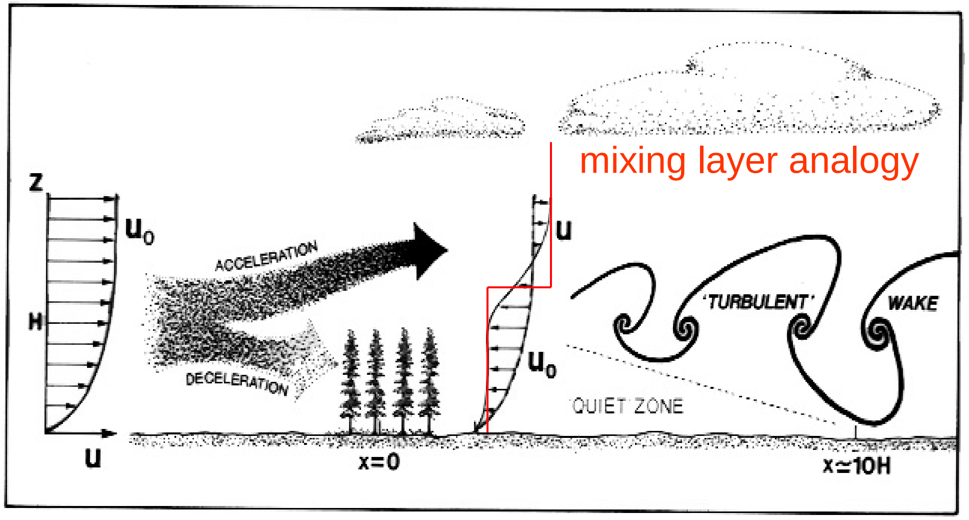


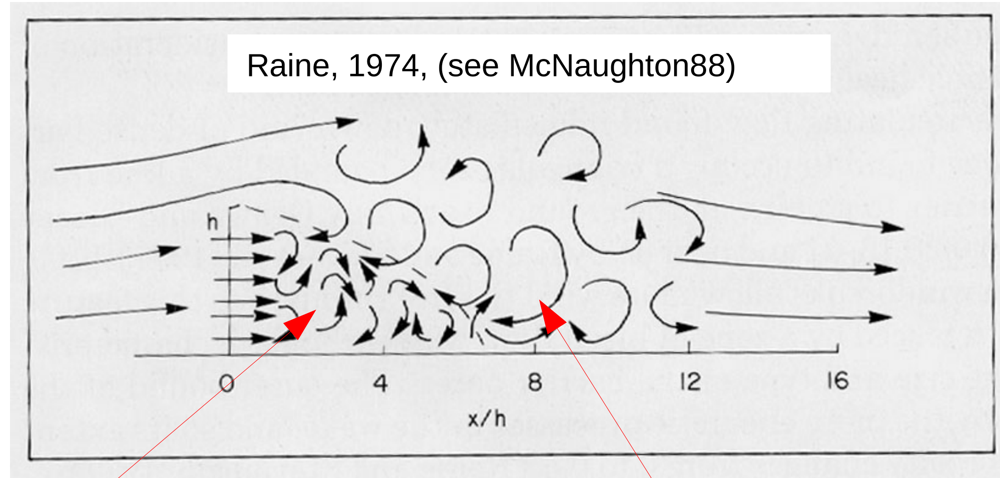
Useful review: pages 17-24 in McNaughton (1988, Agric., Ecosys. & Environ. Vol. 22/23, 17-39)

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



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





"quiet zone" of reduced TKE

- mean wind reduction, turbulent wake

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

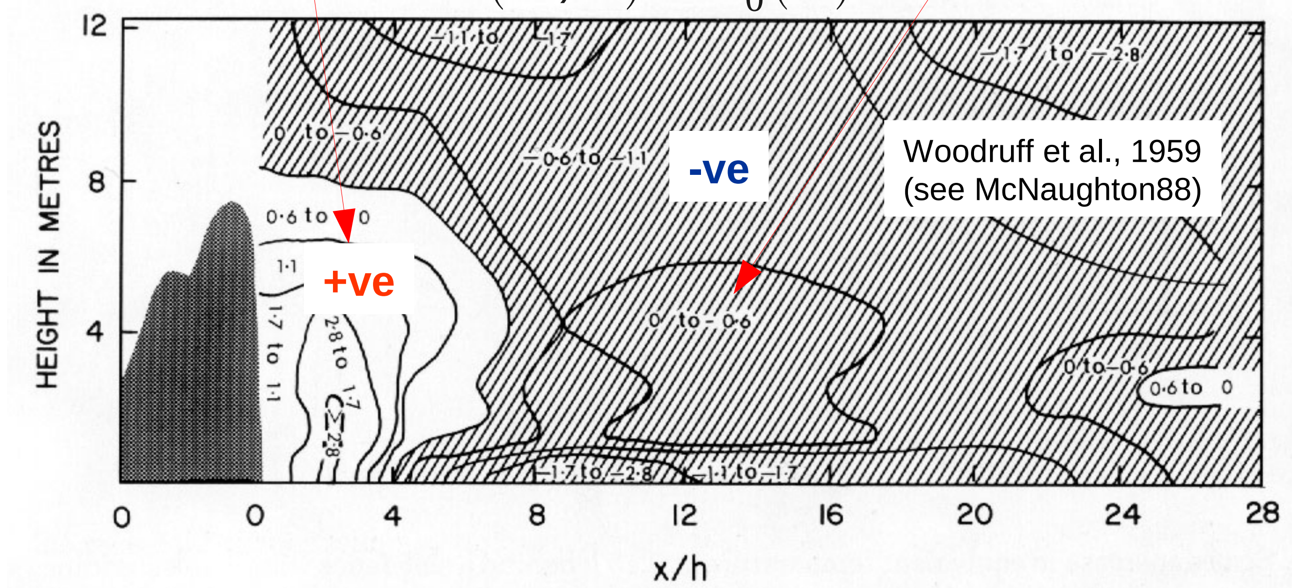
Reduced K_h, so need stronger mean temp gradient to carry Q_H

Enhanced K_h

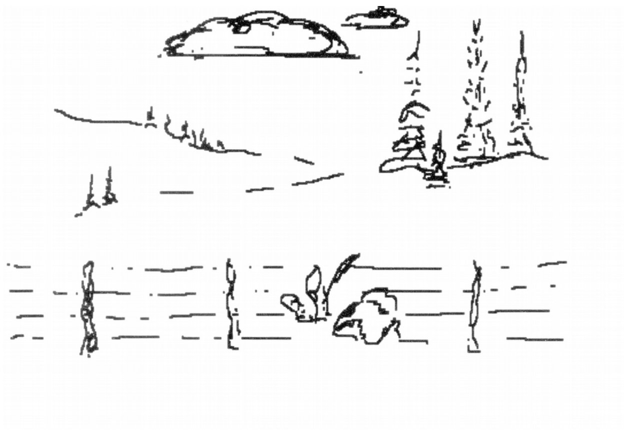
- altered scalar fields

midday summertime anomaly in mean temperature [°C],

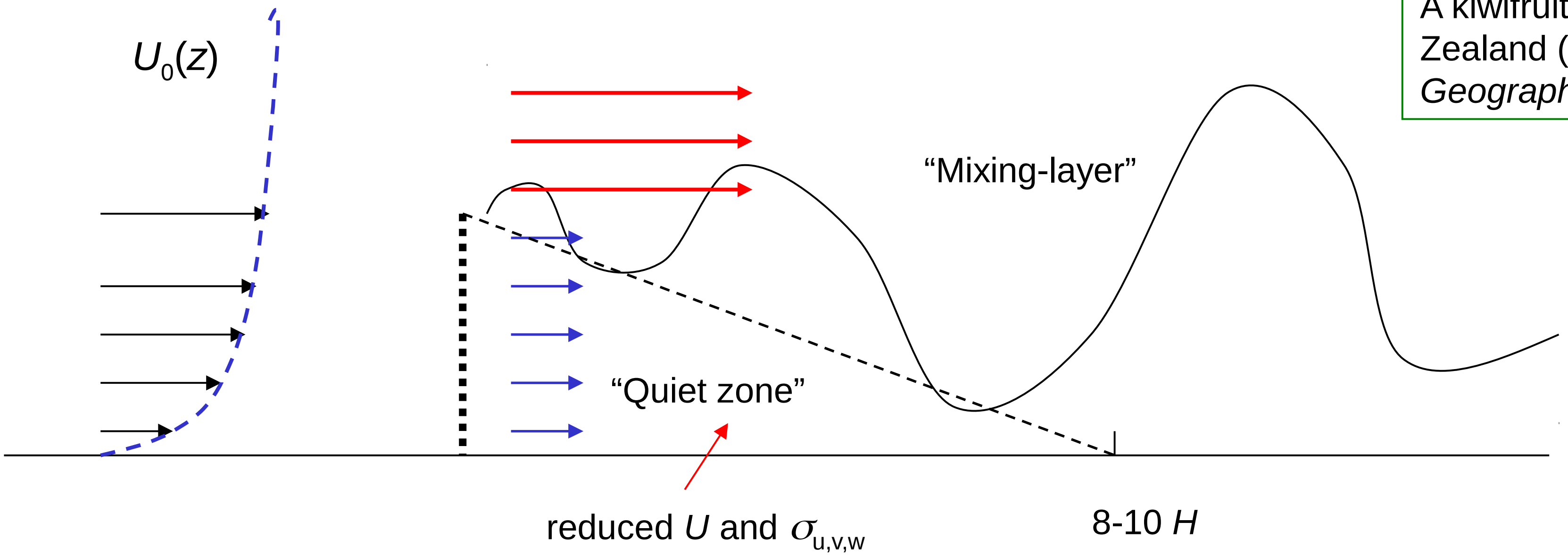
$$\overline{T}(x,z)-\overline{T}_0(z)$$



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
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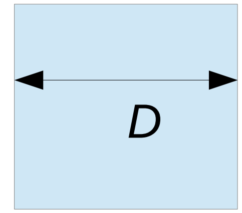


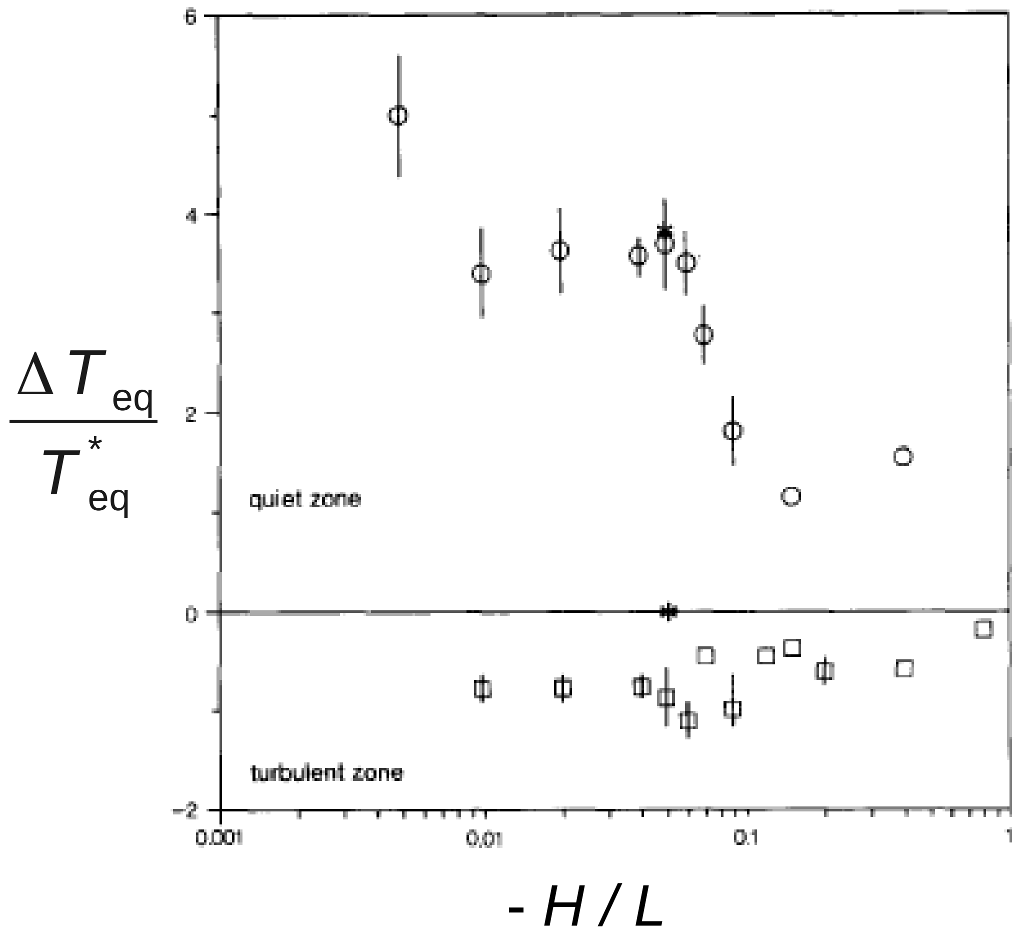
THE MICROCLIMATE IN THE CENTRE OF SMALL SQUARE SHELTERED PLOTS

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

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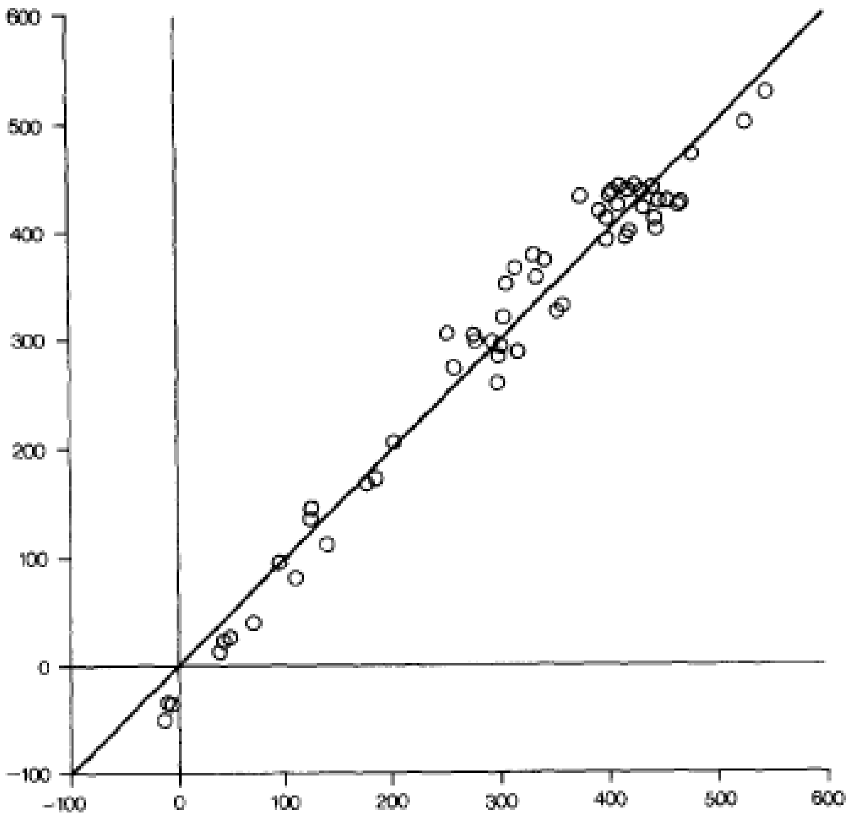
THE MICROCLIMATE IN THE CENTRE OF SMALL SQUARE SHELTERED PLOTS

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

$Q^* - Q_G$



$Q_H + Q_E$

• fitted MO profiles to measured profiles of *U*, *T*, *Q* (humidity) and inferred $u*$, $T*$, $q*$ 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 φ), 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

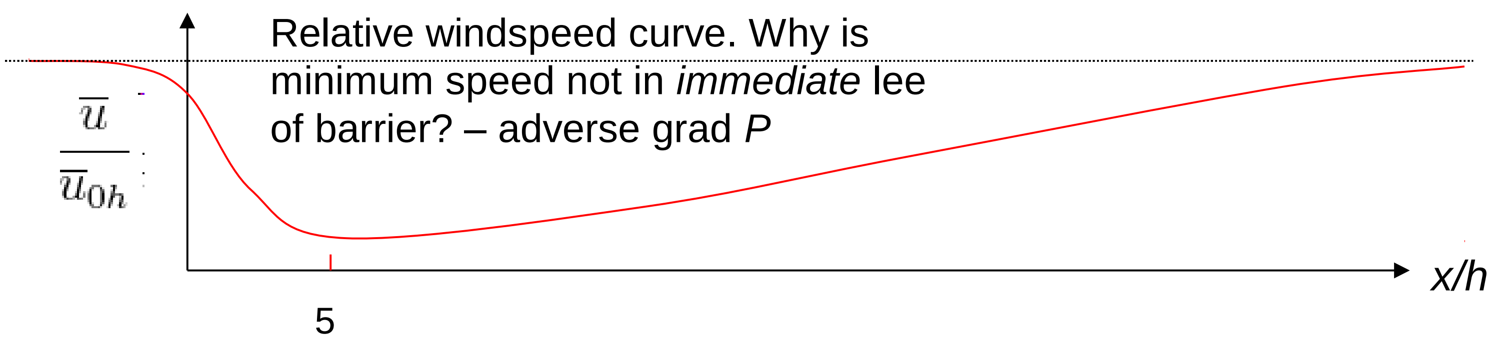
$$\frac{\Delta \overline{u}}{\overline{u}_{0h}} \quad \text{or} \quad \frac{\Delta \overline{u}(x,z)}{\overline{u}_0(z)}$$

(is this strongly height-dependent?)

$$\frac{k}{\overline{u}_{0h}^2} \qquad \frac{k}{u_{*0}^2}$$

"resistance coefficient" (defined over)

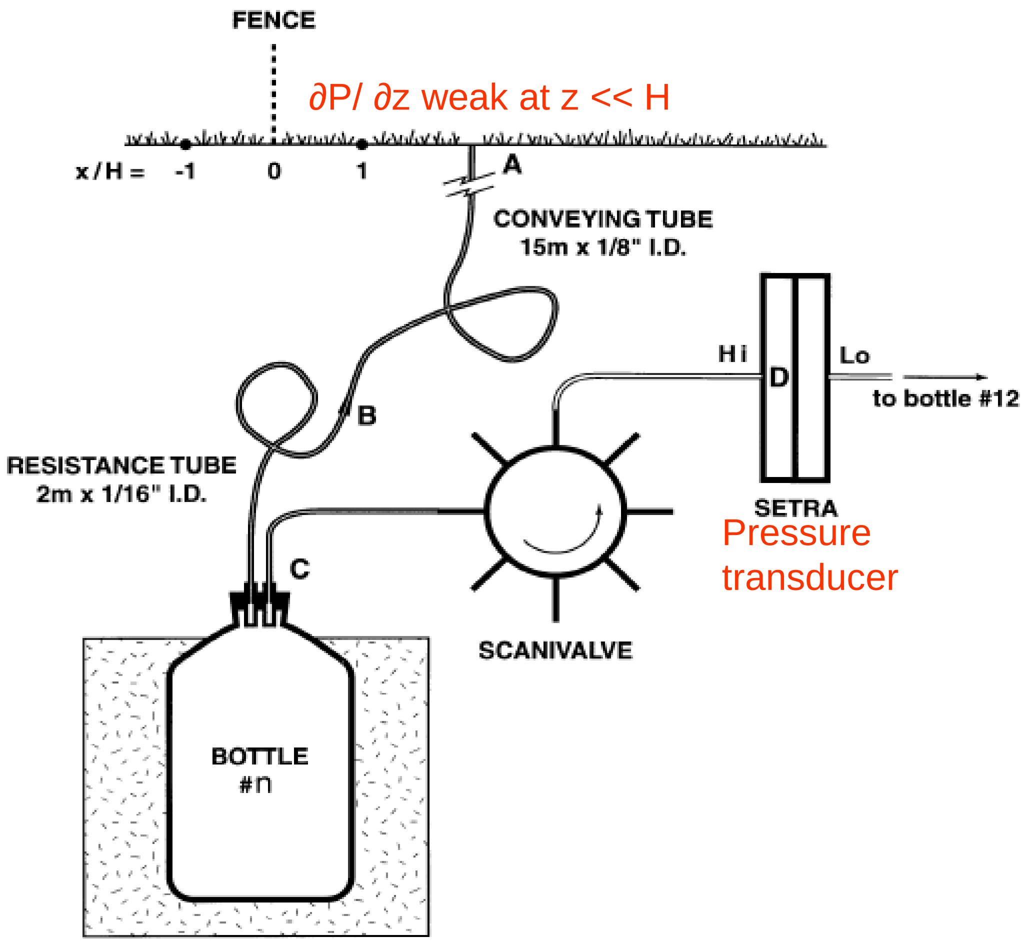
as function of: $\frac{x}{h}, \frac{z}{h}, \frac{h}{L}, \frac{L}{\delta}, \frac{h}{z_0}, \phi, k_r, \frac{\overline{u}_{0h}\, h}{\nu}, \ldots$?



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

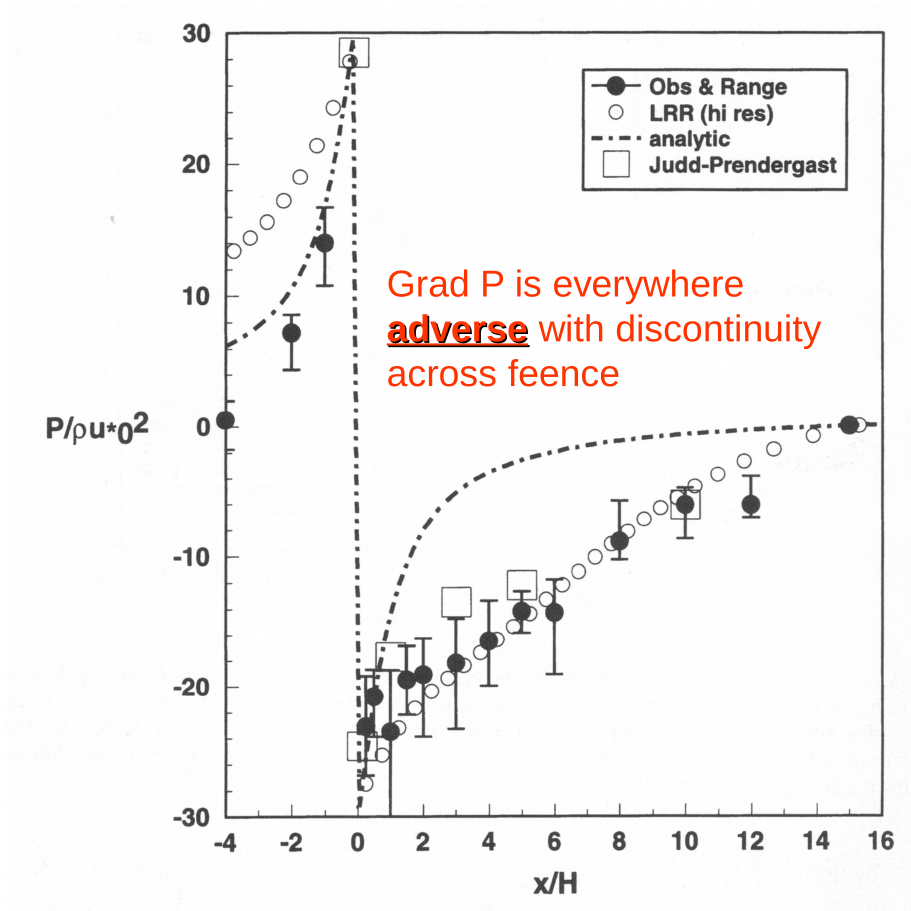
$$\frac{\partial\, \overline{u'w'}}{\partial z}$$

 Wilson, *Boundary-Layer Meteorology* **85:** 327–358, 1997.



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:

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

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

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

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

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

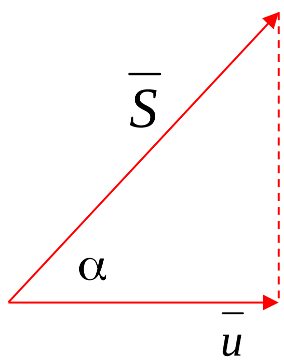
zero by symmetry restriction

of minor importance

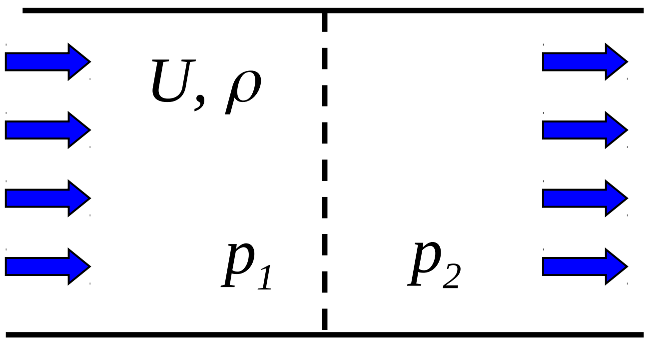
Drag is proportional to projection of $\left(\overline{S}\right)^2$ onto x-axis, where

$\overline{s}^2 = \overline{u}^2 + \overline{v}^2 + \overline{w}^2$

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



Definition of "resistance coefficient" with respect to a uniform stream forced through blocking porous screen



$$k_r = \frac{p_1 - p_2}{\rho U^2}$$

(indep of U for large U)

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

Step fnctn

Earliest attempts to apply fluid mechanics to windbreak (Tani, 1958; Kaiser, 1959)

- treat windbreak as a source of mean velocity deficit $\Delta\overline{u}$

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

(kinematic pressure)

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

neglect

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

Substitute
$$\begin{aligned}\overline{u} &= \overline{u}_0 + k_r\,\Delta\overline{u}\\ \overline{w} &= k_r\Delta\overline{w}\\ \overline{p} &= k_r\,\Delta\overline{p}\end{aligned}$$

"Perturbation expansion" in small parameter k_r

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

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

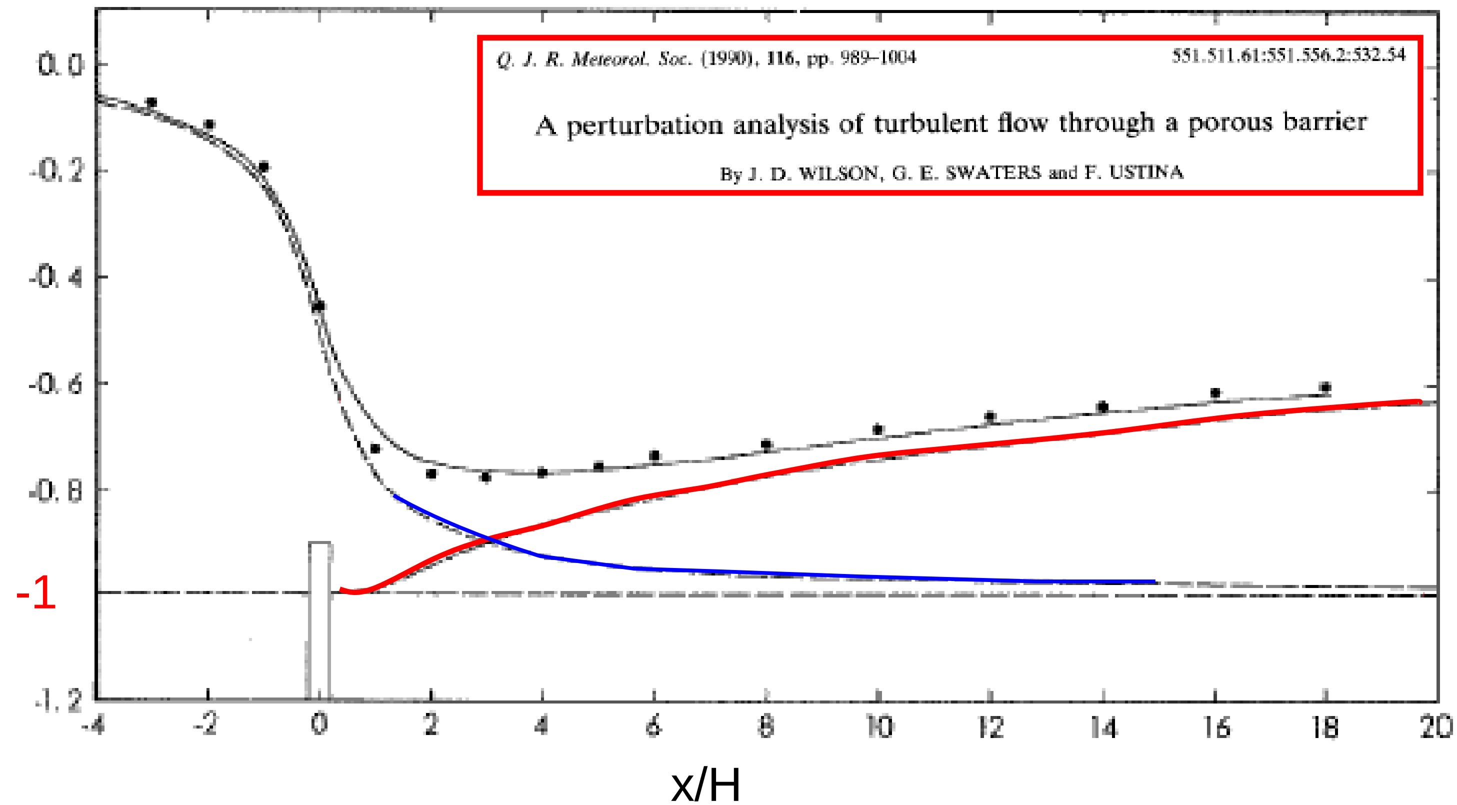
Further simplifications: $\Delta\overline{w} = 0,\ \partial\overline{p}/\partial x = 0,\ K = K_0 = \text{const.},\ \overline{u}_0 = \text{const.}$

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 $dQ=k_r\, u_0^2\, dz$

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

$$\frac{1}{k_r}\;\frac{\Delta\overline{u}}{\overline{u}_0}$$



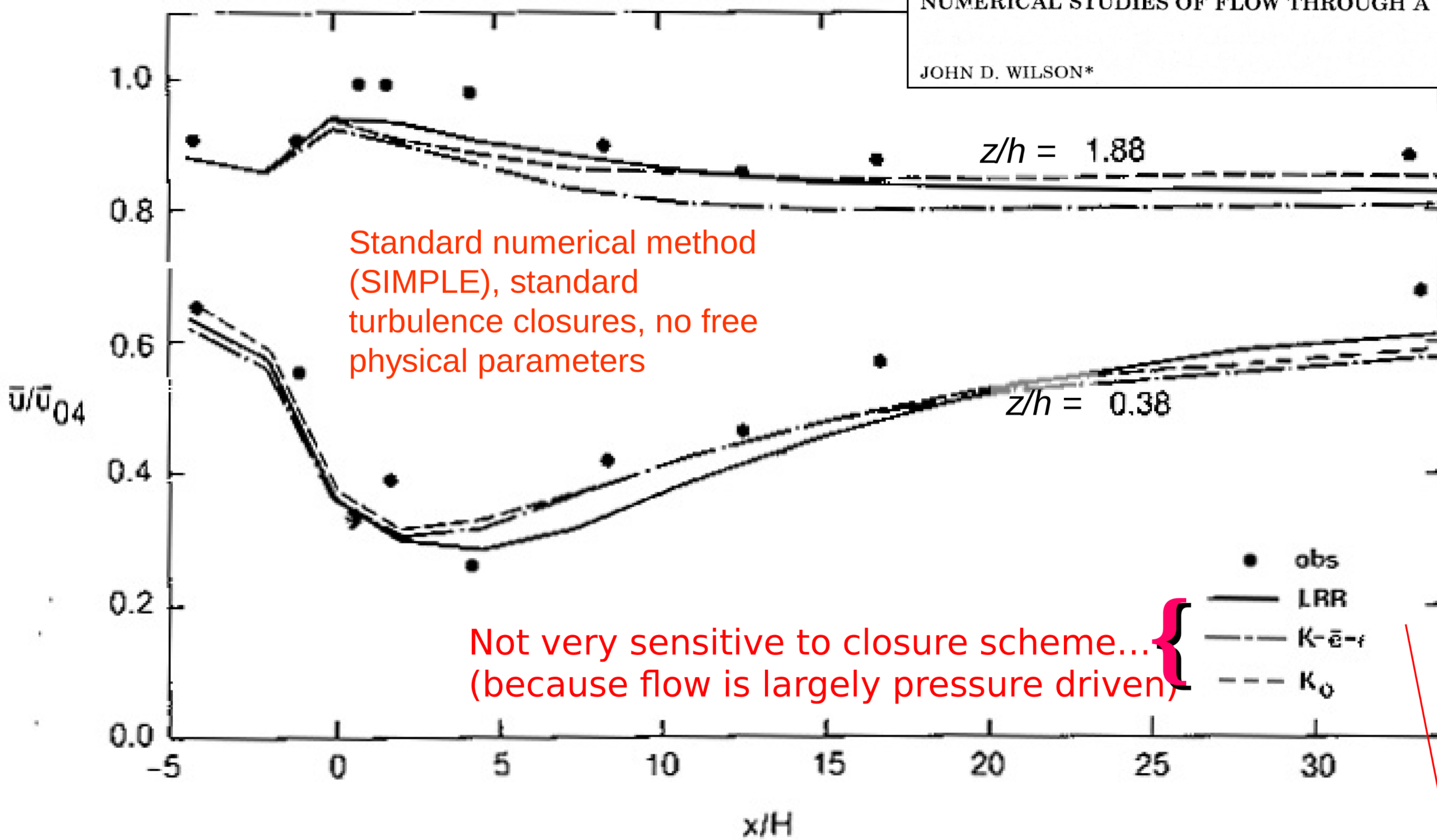
Kaiser's solution necessarily places minimum velocity at the barrier (source of momentum deficit) – unrealistic. Contrast with later analytic solutions that retain grad P. The dashed line – no recovery – neglects $\partial\overline{u'w'}/\partial z$

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)

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

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

Closures :

$$(1) \quad K_0 = k_v u_{*0} z,$$

$$(2) \quad k - \epsilon$$

$$(3) \quad \text{LRR second order closure}$$

Numerical solution of mtm eqns

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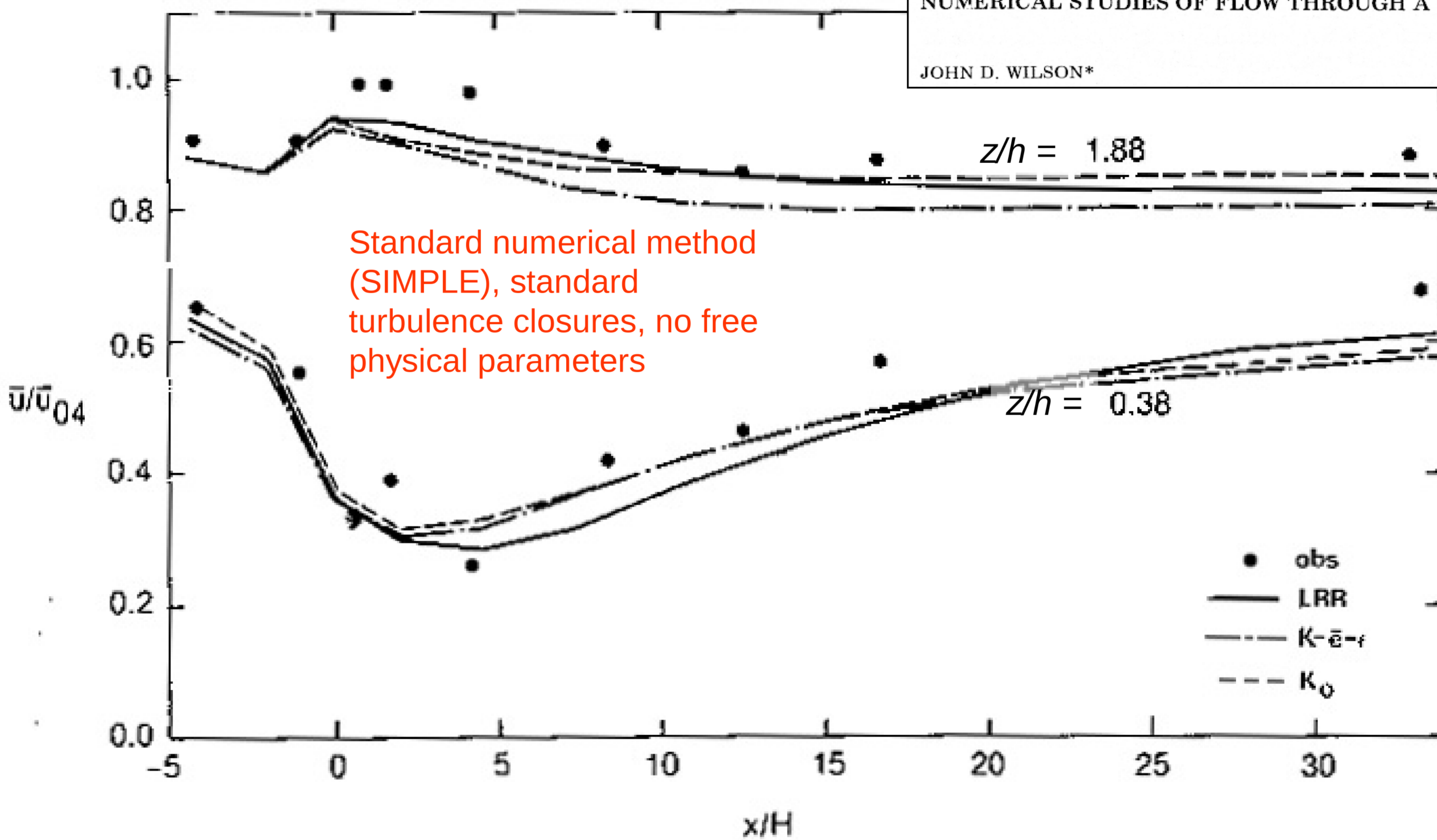
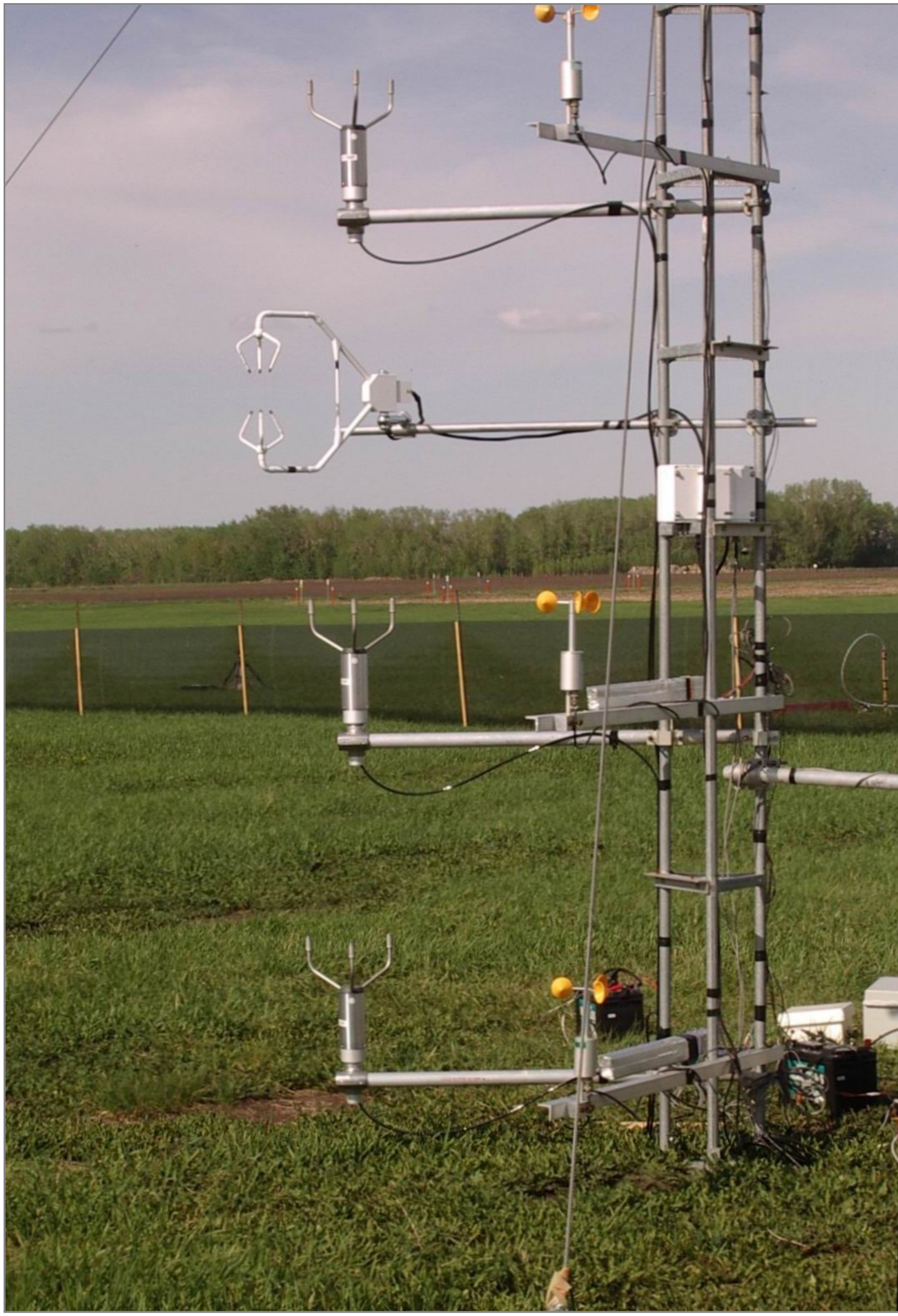


Minimum mean wind speed occurs at about *5H* downwind of the barrier, and the fractional reduction in wind speed at that point is:

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

Windbreak experiment at Ellerslie



11 cup anemometers
8 two-D sonic anemometers
2 three-D sonic anem/thermometers (16 Hz)
wind vane
2 thermocouple ΔTs

34 wind signals, 4 T signals, 3 dataloggers

Mean speed… effect of stratification (L) in perpendicular flow

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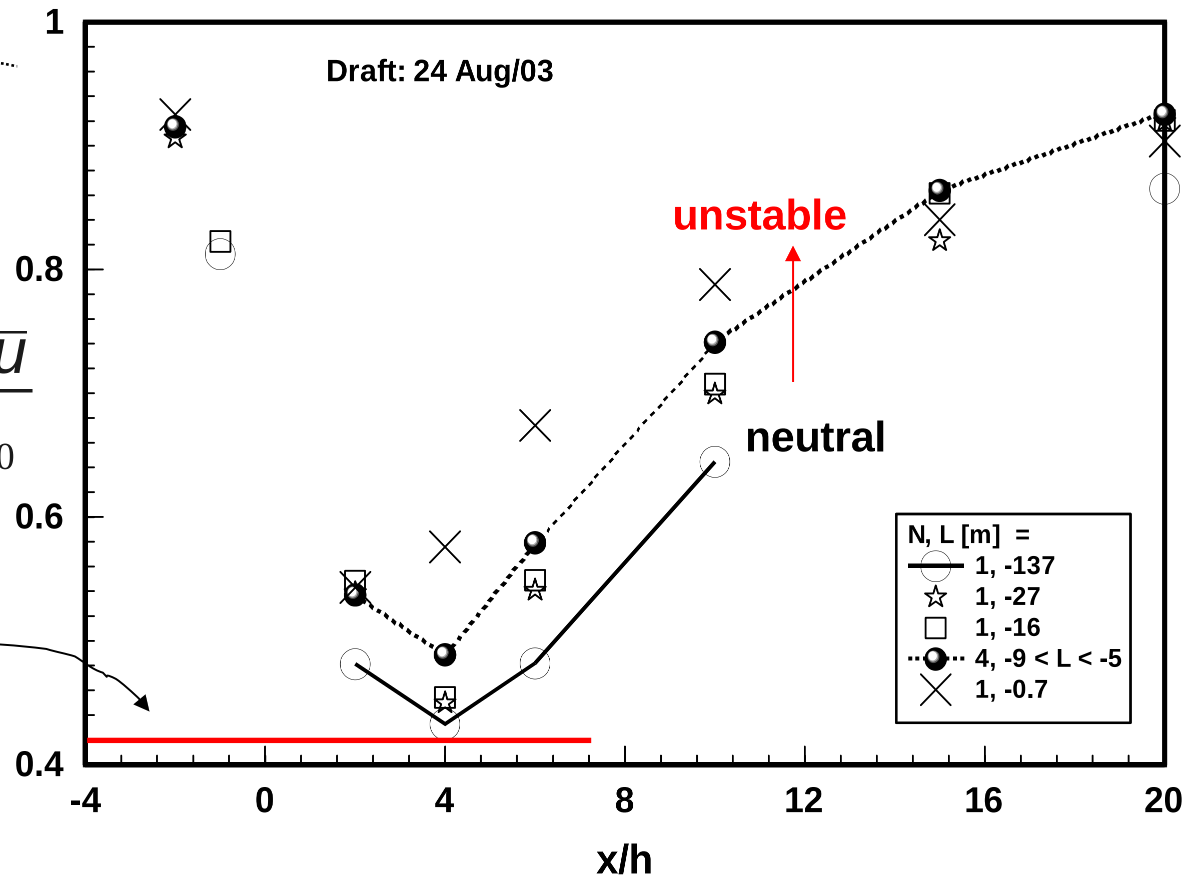
Oblique, Stratified Winds about a Shelter Fence. Part II: Comparison of Measurements with Numerical Models

JOHN D. WILSON

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

"relative windspeed" $\dfrac{\Delta \bar{u}}{\bar{u}_0}$



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

Mean speed… effect of obliquity in neutrally-stratified winds