Direct calculation of entrainment and detrainment in large eddy simulations of boundary layer clouds.

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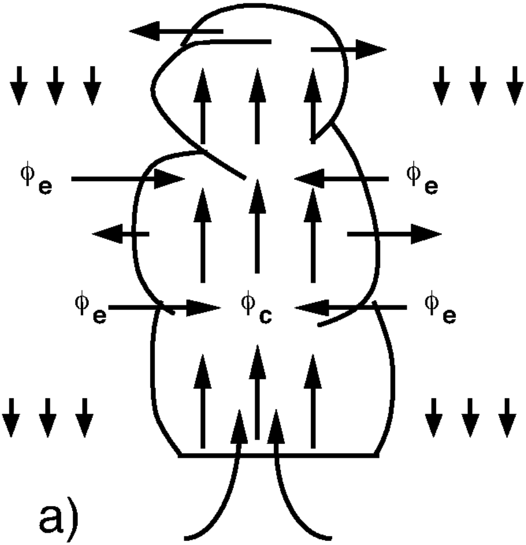
The problem: how do we represent this:

Using this?



Siebesma, 1998

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- Two new approaches have been developed to circumvent this problem:
  - Romps (2010): Time-average the entrainment fluxes over the time needed for an entire grid cell flip from cloud to environment
  - Dawe and Austin (2010): Use spatial interpolation to improve the all or nothing estimate of the cloud volume.

Bulk entrainment – definitions

- Define an averaging operator:

$$\overline{\phi(z)} = \frac{1}{A}\int_0^{L_x}\int_0^{L_y}\phi(x,y,z)dxdy$$
$$\text{where}\quad A = L_x L_y$$

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- Separate the domain into environment and cloud core:

$$\overline{\phi_c} = \phi_c = \frac{1}{A_c} \int \int_{\text{cloud}} \phi(x, y, z) dxdy$$
$$\overline{\phi_e} = \phi_e = \frac{1}{A_e} \int \int_{\text{env}} \phi(x, y, z) dxdy$$
$$a_c = A_c / A \text{ (fractional cloud cover)}$$

Bulk entrainment, cont.

- The entrainment and detrainment rates are:

$$E = -\frac{1}{A}\oint_{\hat{\mathbf{n}}\cdot(\mathbf{u}-\mathbf{u_i})<0} \rho\hat{\mathbf{n}}\cdot(\mathbf{u}-\mathbf{u_i})dl$$

$$D = \frac{1}{A}\oint_{\hat{\mathbf{n}}\cdot(\mathbf{u}-\mathbf{u_i})>0} \rho\hat{\mathbf{n}}\cdot(\mathbf{u}-\mathbf{u_i})dl$$

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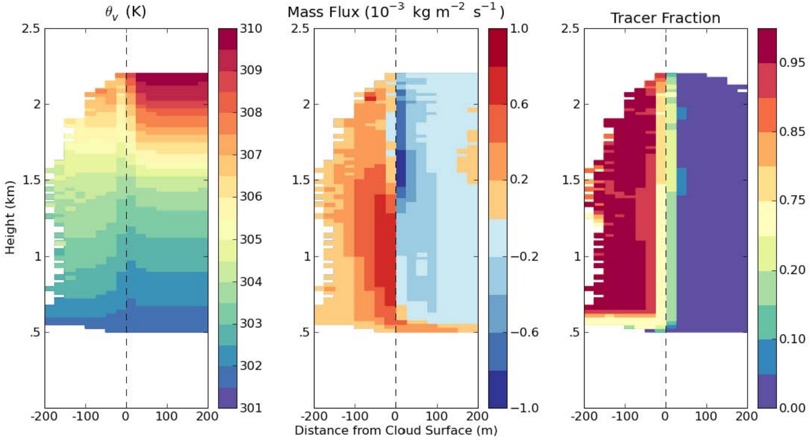
- The bulk plume/environment approximation:

$$E_\phi \phi_e = -\frac{1}{A} \oint_{\hat{\mathbf{n}}\cdot(\mathbf{u}-\mathbf{u_i})<0} \rho \hat{\mathbf{n}} \cdot (\mathbf{u} - \mathbf{u_i}) \phi dl$$

$$D_\phi \phi_c = \frac{1}{A} \oint_{\hat{\mathbf{n}}\cdot(\mathbf{u}-\mathbf{u_i})>0} \rho \hat{\mathbf{n}} \cdot (\mathbf{u} - \mathbf{u_i}) \phi dl$$

to proceed, assume $E_\phi = E$ and $D_\phi = D$, but ...

Clouds are surrounded by a cool moist shell



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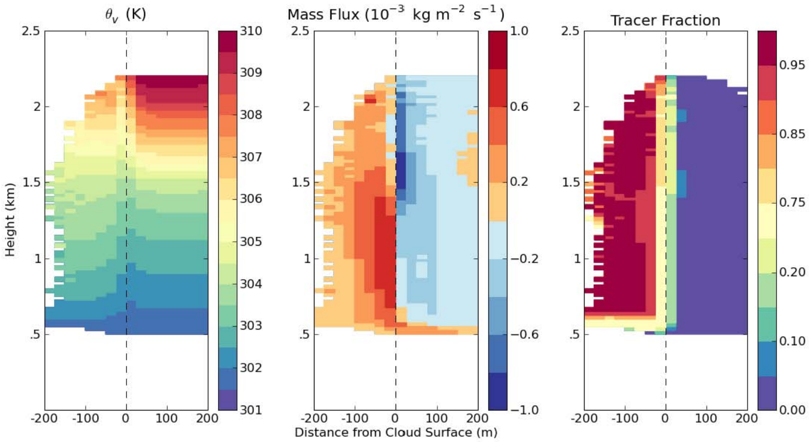

- So entrained/detrained air proprieties differ from ϕ_e, ϕ_c

Calculating bulk E_ϕ and D_ϕ

- Following Siebesma and Cuijpers, mass and tracer continuity yield:

$$E_\phi(\phi_c - \phi_e) = -M_c \frac{\partial \phi_c}{\partial z} - \frac{\partial \rho a \overline{w'\phi'}^c}{\partial z} - \rho a \frac{\partial \phi_c}{\partial t} + a\rho \left(\frac{\partial \bar{\phi}}{\partial t}\right)_{forcing}$$

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- and

$$\begin{aligned} D_\phi(\phi_c - \phi_e) = -M_c \frac{\partial \phi_e}{\partial z} + \frac{\partial \rho (1-a) \overline{w'\phi'}^e}{\partial z} \\ + \rho(1-a) \frac{\partial \phi_e}{\partial t} - \rho(1-a) \left(\frac{\partial \bar{\phi}}{\partial t} \right)_{forcing} \end{aligned}$$

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- where LES is used for the rhs terms. So how do E_ϕ and D_ϕ compare to E and D?

Romps time-averaged direct calculation for E_d and D_d

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- To get the direct $E_d(z)$ and $D_d(z)$:
  - Average $e - d$ over the time required for a grid cell to change state
  - Label positive $e - d$ as e, negative $e - d$ as d
  - Sum e and d over (x, y) to get $E_d(z)$ and and $D_d(z)$.

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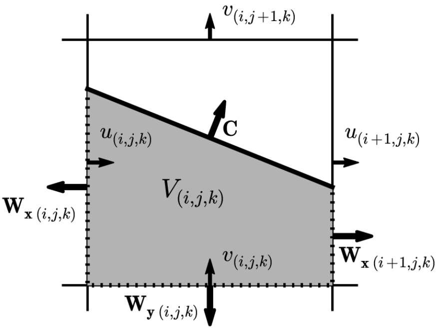
  - Sum e and d over (x, y) to get $E_d(z)$ and and $D_d(z)$.
- Romp's result: E_d and D_d are about a factor of 2 larger than E_ϕ and D_ϕ, and depend on the tracer type

Direct calculations using cloud surface interpolation



- Dawe and Austin (MWR, 2010)

Start with the net entrainment and derive:

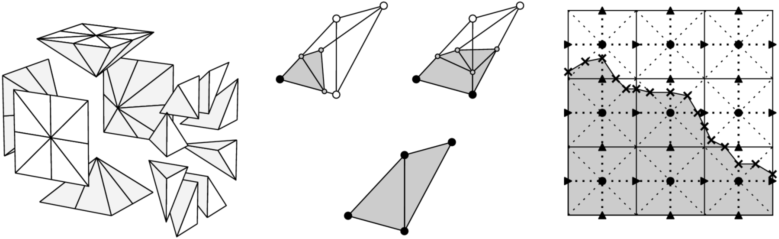
$$E - D = \int_C \rho(\mathbf{u_i} - \mathbf{u}) \cdot d\mathbf{C}$$

$$E - D = \rho \frac{dV}{dt} + \int_W \rho \mathbf{u} \cdot d\mathbf{W}$$

$$E = \max\left(0, \rho \frac{dV}{dt} + \int_W \rho \mathbf{u} \cdot d\mathbf{W}\right)$$

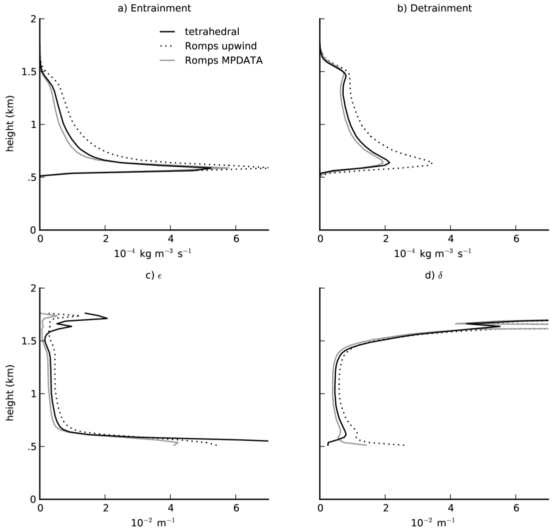
$$D = \max\left(0, -\rho \frac{dV}{dt} - \int_W \rho \mathbf{u} \cdot d\mathbf{W}\right)$$

Use tetrahedral interpolation to get the subgrid volume



- Where the core-environment boundary is determined by interpolating q_v, q_s, T_v and w.

Good agreement between Romps and tetrahedral

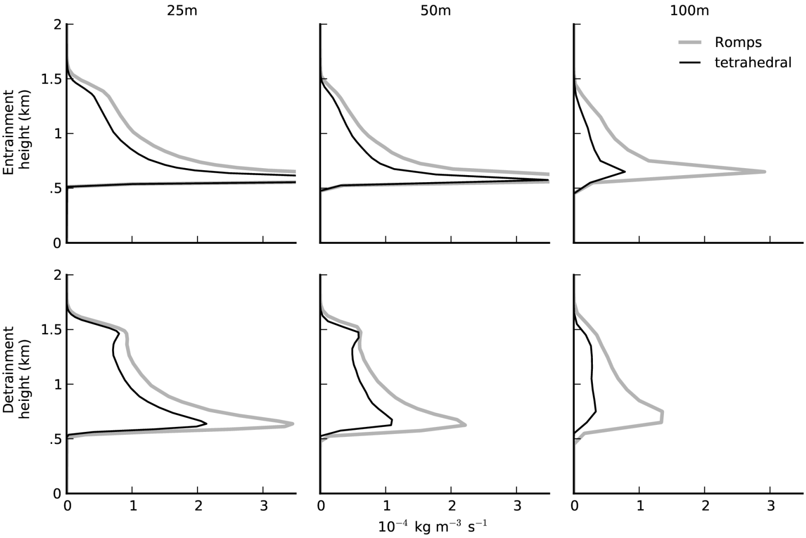


E_d, D_d

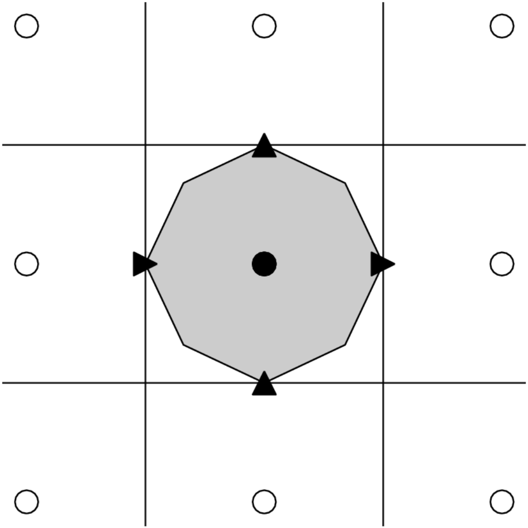
ϵ_d, δ_d

but there's some dependence on numerics (note advection scheme differences)

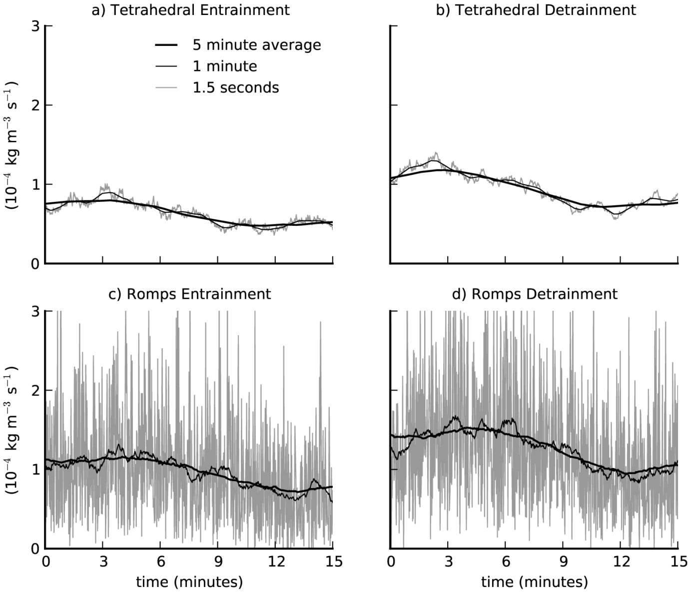
The tetrahedral technique requires high resolution . . .



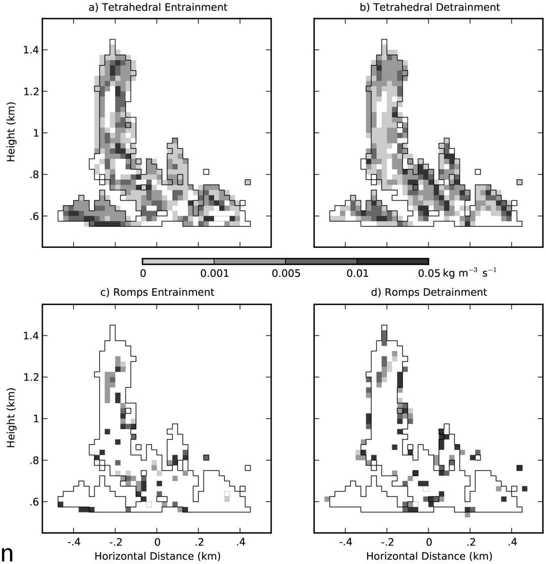
Because the interpolation biases single gridcell cloud area



Tetrahedral fluxes can be used for E_d/D_d snapshots



E_d/D_d spatial variability, 1 minute average

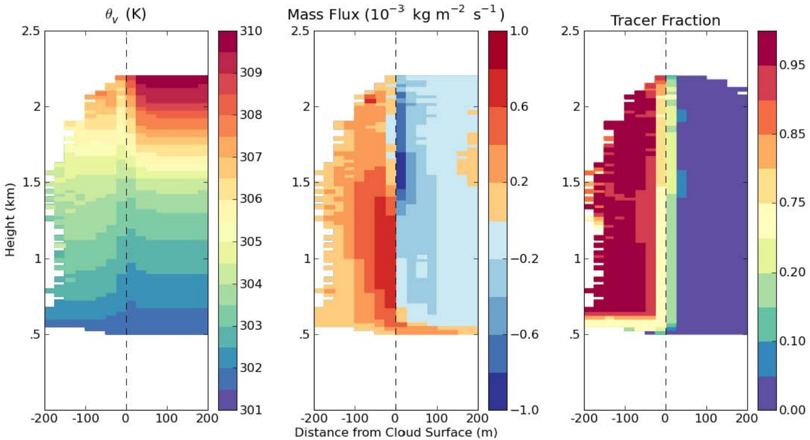


Tetrahedral

Romps

n

Linking direct and bulk entrainment rates

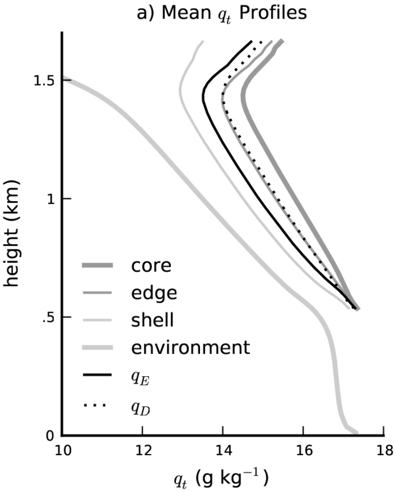


Core: $w > 0$, $\Delta T_v > 0$, $q_l > 0$

Edge: outermost core gridcell

Shell: Innermost environment gridcell

converting E_d to E_ϕ



Define shell and edge tracer values q_{shell} and q_{edge}.
These values will differ from q_c and q_e, the mean cloud core and environment vapor values
Can show that:

$$E_\phi = E_d - E_d \frac{q_{shell} - q_e}{q_c - q_e}$$

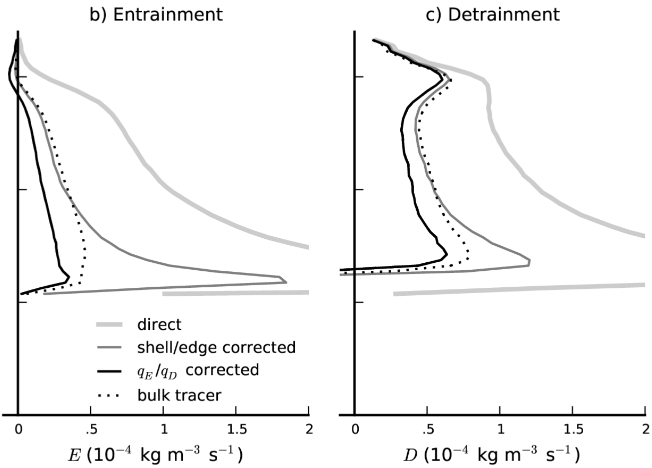
$$-D_d \frac{q_c - q_{edge}}{q_c - q_e}$$

Alternatively, use conditional averages to include Reynolds correlations

$$q_E = (E\phi)_d / E_d$$

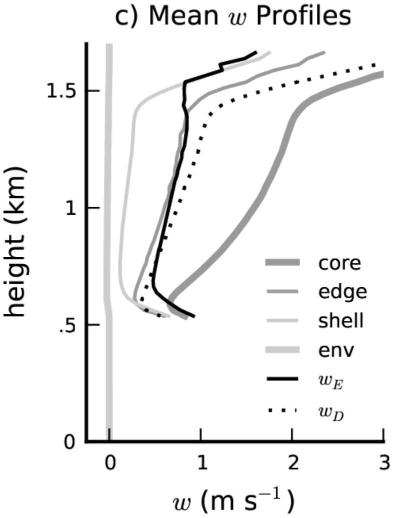
$$q_D = (D\phi)_d / D_d$$

Corrected fluxes restore bulk tracer profile



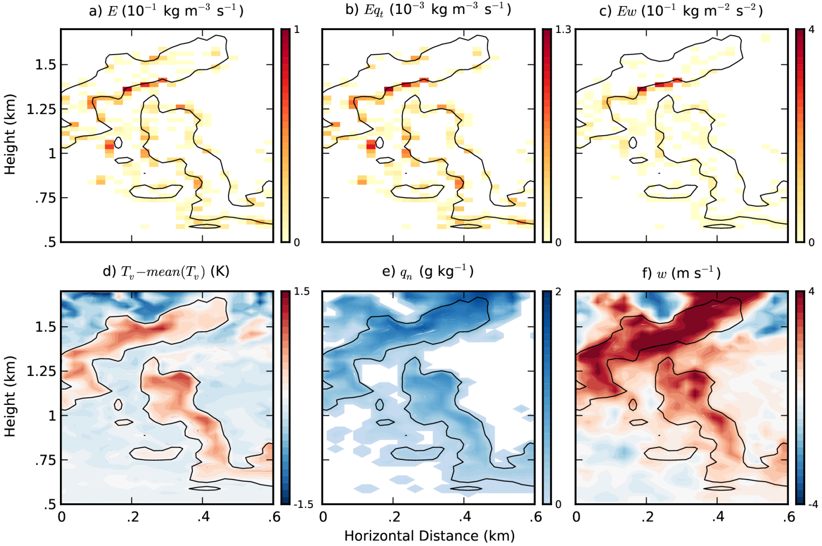
The q_E / q_D underestimate of E and D arises from differencing two large quantities: $q_c(E_d - D_d)$ and $(Eq)_d - (Dq)_d$.

Vertical momentum



When we modify the entrainment calculation to account for negative and positive w, we find w_E, the Reynold's correlated mean entrained vertical velocity > 0 and nearly as large as w_D.

Why is is the cloud entraining positive w?



Updrafts produce newly buoyant parcels well above cloudbase.

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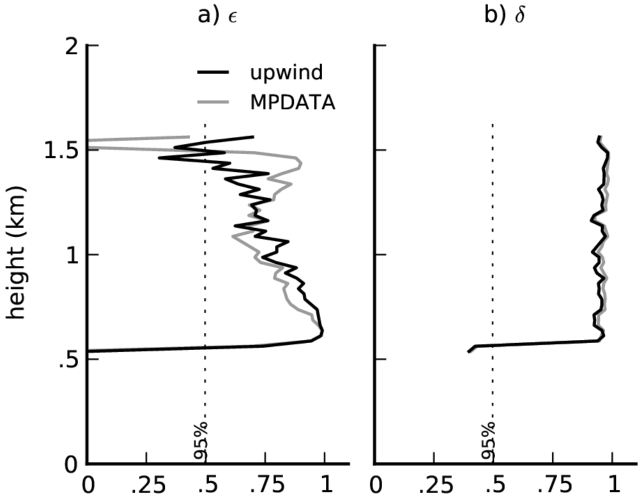
- Tetrahedral interpolation can be applied to individual clouds, and rapidly changing boundary layers, either over a cloud life cycle or in a snapshot.
- The interpolation technique is applicable in general to any flow through a material surface in a three-dimensional model.

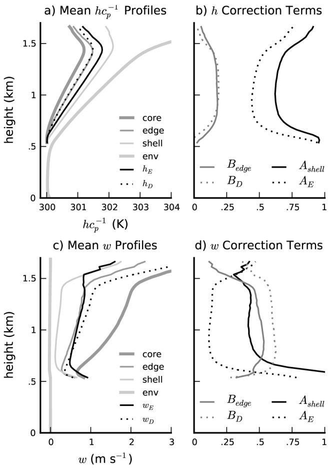
Thanks to:

- Marat for providing SAM
- Support from the Canadian Foundation for Climate and Atmospheric Science

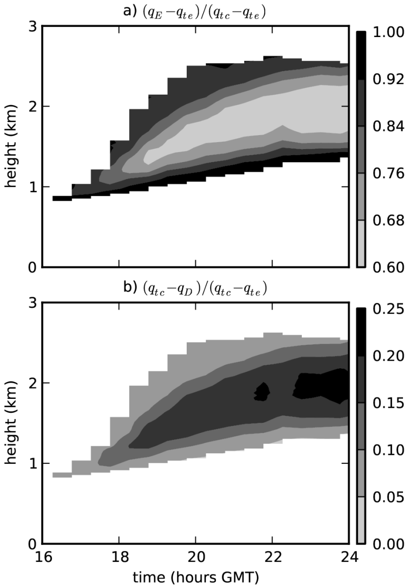
Romps - tetrahedral correlation



static energy



ARM diurnal



a) $(q_E - q_{te})/(q_{tc} - q_{te})$
b) $(q_{tc} - q_D)/(q_{tc} - q_{te})$
height (km)
time (hours GMT)
0
1
2
3
16
18
20
22
24
1.00
0.92
0.84
0.76
0.68
0.60
0.25
0.20
0.15
0.10
0.05
0.00