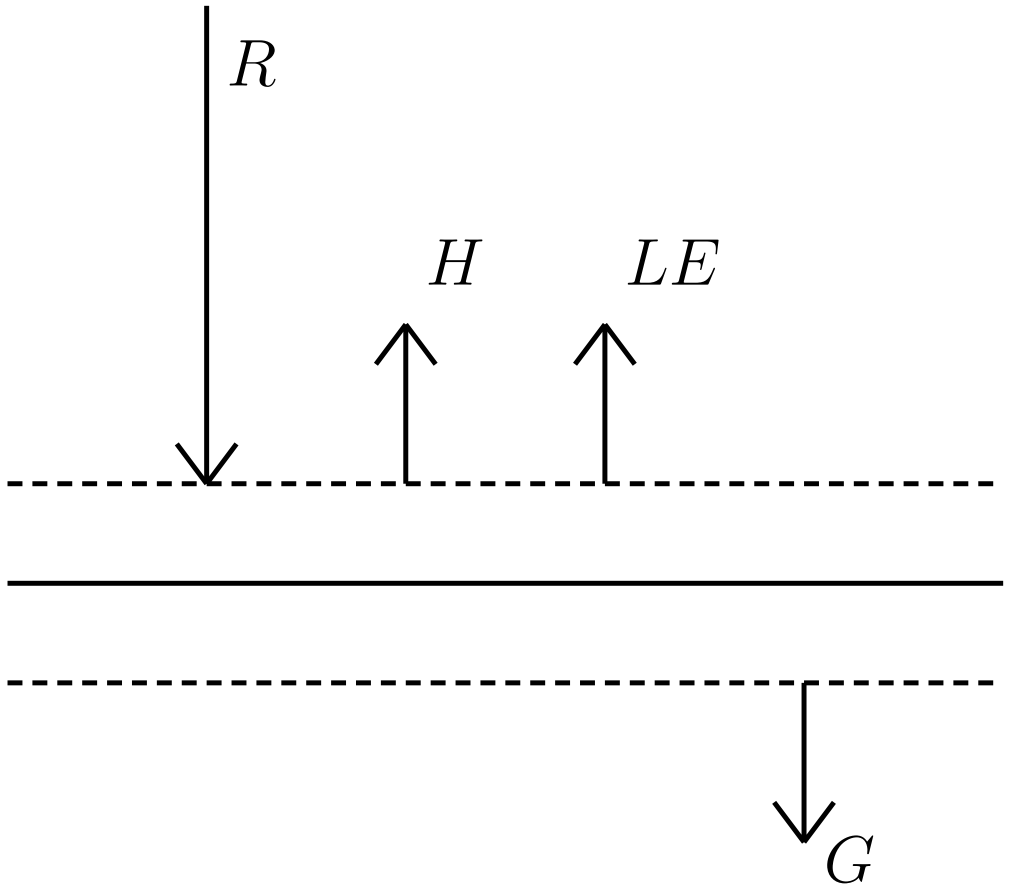
METR 5223: Atmospheric Radiation

Elements of Radiative Convective Modeling

Lecture for Spring 2009

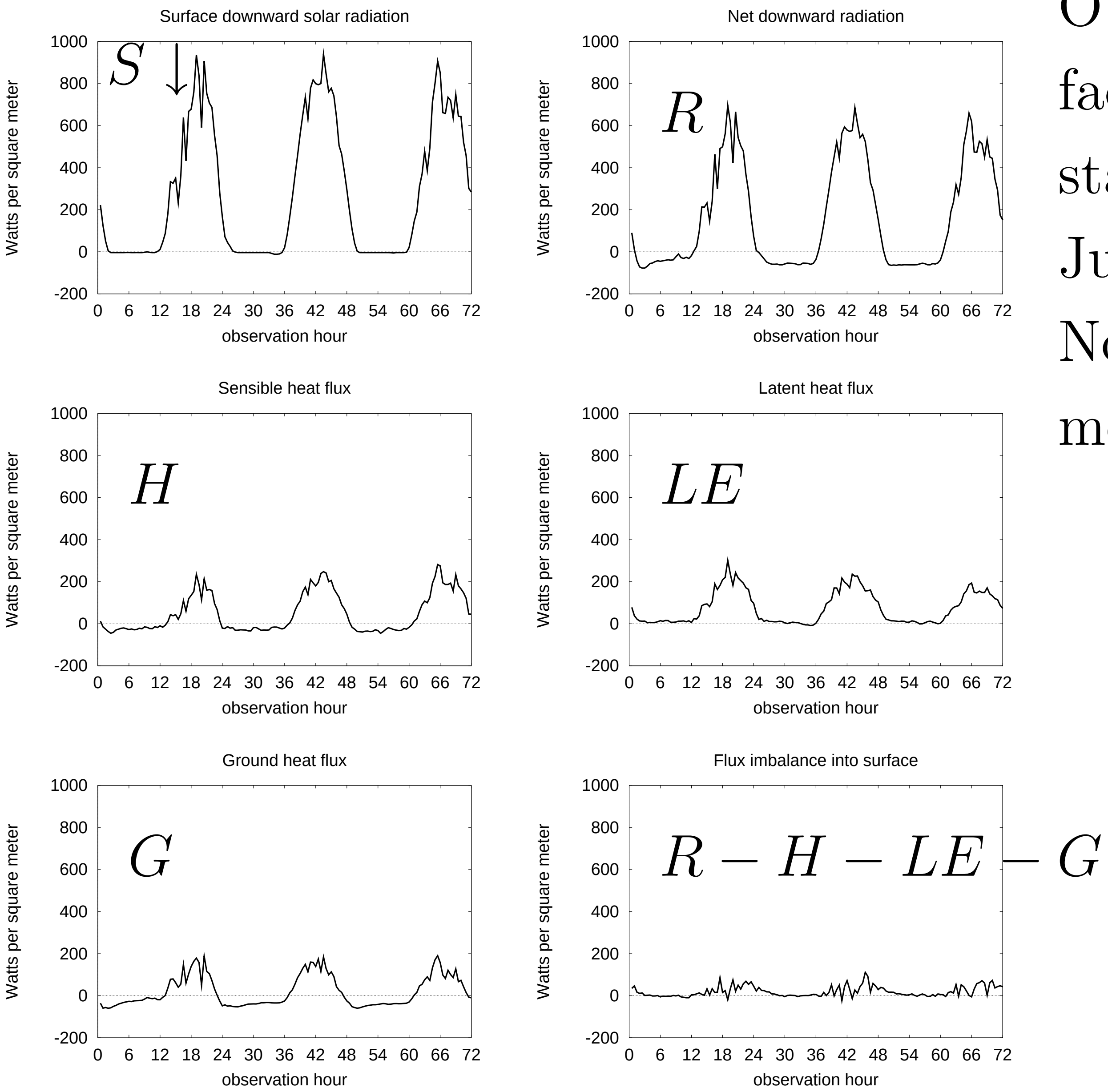
Prof. Brian H. Fiedler

School of Meteorology, University of Oklahoma



Energy balance at surface. R is net absorbed radiation, H is the sensible heat flux, LE is the latent heat flux. G is the heat flux into the ground.

$$R = H + LE + G\ .$$



Observations of surface energy fluxes starting from 0 Z, June 7, 1999 at the Norman, Oklahoma mesonet site.

Net radiation:

$$R = S \downarrow (1 - \alpha_s) + \epsilon (F \downarrow - \sigma T_s^4) \ .$$

Standard bulk aerodynamic formulas used in models:

$$H = c_p \rho C_H U_r \left[T_s - T_a(z_r) \right]$$

$$E = \rho C_E U_r \left[q_s - q_a(z_r) \right]$$

For certain surfaces — an ocean surface, or a dry unvegetated soil — the coefficients are well known.

Consider a dry sand with $\alpha_s = 0.35$, $\epsilon = 0.9$, and $C_H = 2 \times 10^{-3}$.

Example calculation:

Assume $S\downarrow = 400$ W m^{-2}, $G = 0$, $F\downarrow = 300$ W m^{-2}, $T_a = 20$ C, $\rho = 1.2$ kg m^{-3}, and $U_r = 10$ m s^{-1}.

Find T_s in $R = H + LE + G$ or:

$$S\downarrow(1-\alpha_s) + \epsilon(F\downarrow - \sigma T_s^4) = c_p \rho C_H U_r \left[T_s - T_a(z_r)\right]$$

(The sand is dry so $q_s = q_a(z_r)$ and $E = 0$.)

The answer is $T_s = 25.2$ C, which is 5.2 C greater than the air temperature.

With $U_r = 1$ m s^{-1}, $T_s = 38.97$ C.

Also changing to $\epsilon = 1$, $T_s = 36.39$ C.

Thus our calculations seem to tolerate errors, or lack of knowledge, of the precise value of ϵ when compared with the big changes in T_s caused by the the natural variability in the wind speed. If you are contemplating walking across the pavement barefoot, you should be more concerned about knowledge of the wind speed, rather than the emissivity.

Radiative heating rates in the atmosphere

Let F be a net vertical irradiance:

$$F \equiv F\uparrow - F\downarrow$$

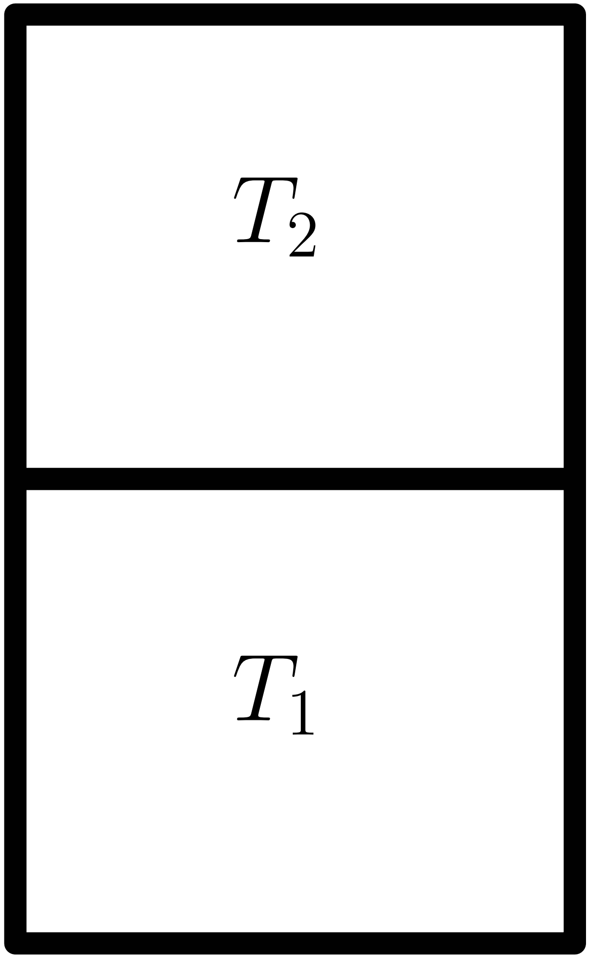
The convergence of this flux will cause a heating:

$$\rho c_p \left(\frac{dT}{dt} \right)_{\mathrm{RAD}} = -\frac{dF}{dz}$$

or

$$\left(\frac{dT}{dt} \right)_{\mathrm{RAD}} = \frac{g}{c_p} \frac{dF}{dp}$$

Convective heating rates in a model atmosphere



Suppose

$$\frac{T_2 - T_1}{z_2 - z_1} < \left.\frac{dT}{dz}\right|_{\text{crit}}$$

Find ΔT such that $T_2^{\text{new}} = T_2^{\text{old}} + \Delta T$ and $T_1^{\text{new}} = T_1^{\text{old}} - \Delta T$ makes

$$\frac{T_2 - T_1}{z_2 - z_1} = \left.\frac{dT}{dz}\right|_{\text{crit}}$$

Suppose the adjustment process produces ΔT in a time step Δt, the convective heating rate is:

$$\left(\frac{dT}{dt}\right)_{\mathrm{CON}} = \frac{\Delta T}{\Delta t}$$

In radiative equilibrium,

$$\left(\frac{dT}{dt}\right)_{\mathrm{RAD}} = 0$$

In radiative-convective equilibrium,

$$\left(\frac{dT}{dt}\right)_{\mathrm{RAD}} + \left(\frac{dT}{dt}\right)_{\mathrm{CON}} = 0$$

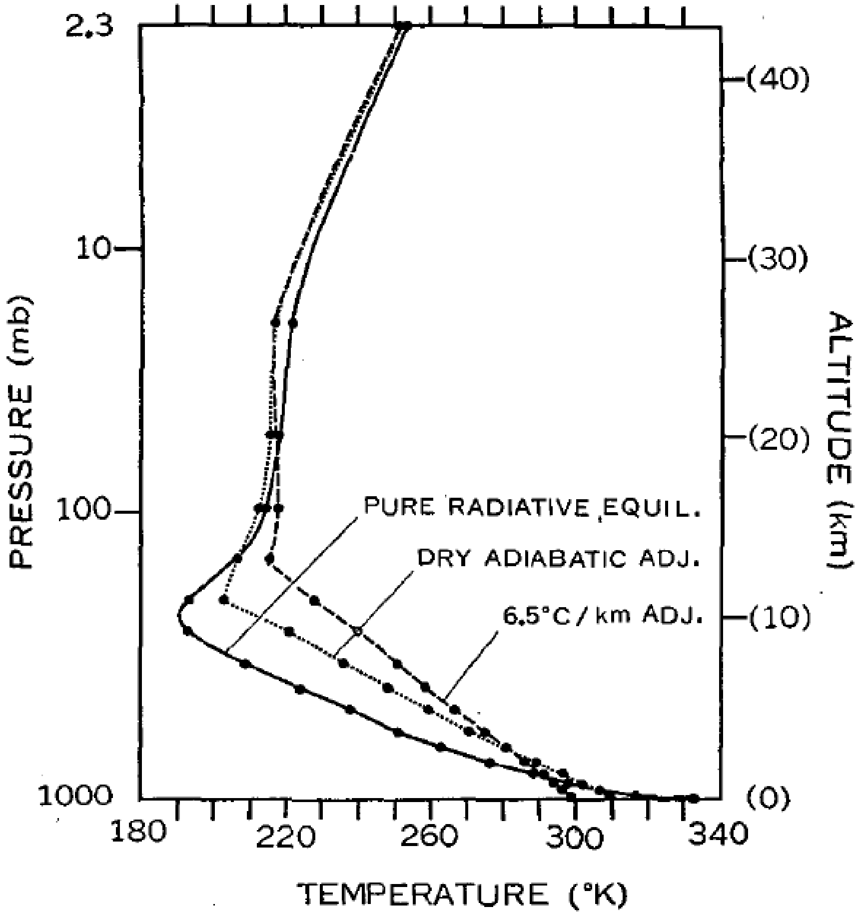


FIG. 4. The dashed, dotted, and solid lines show the thermal equilibrium with a critical lapse rate of 6.5 deg km^{-1}, a dry-adiabatic critical lapse rate (10 deg km^{-1}), and pure radiative equilibrium.

Radiative equilibrium and radiative convective equilibrium in the classic paper of Manabe and Strickler (1964).

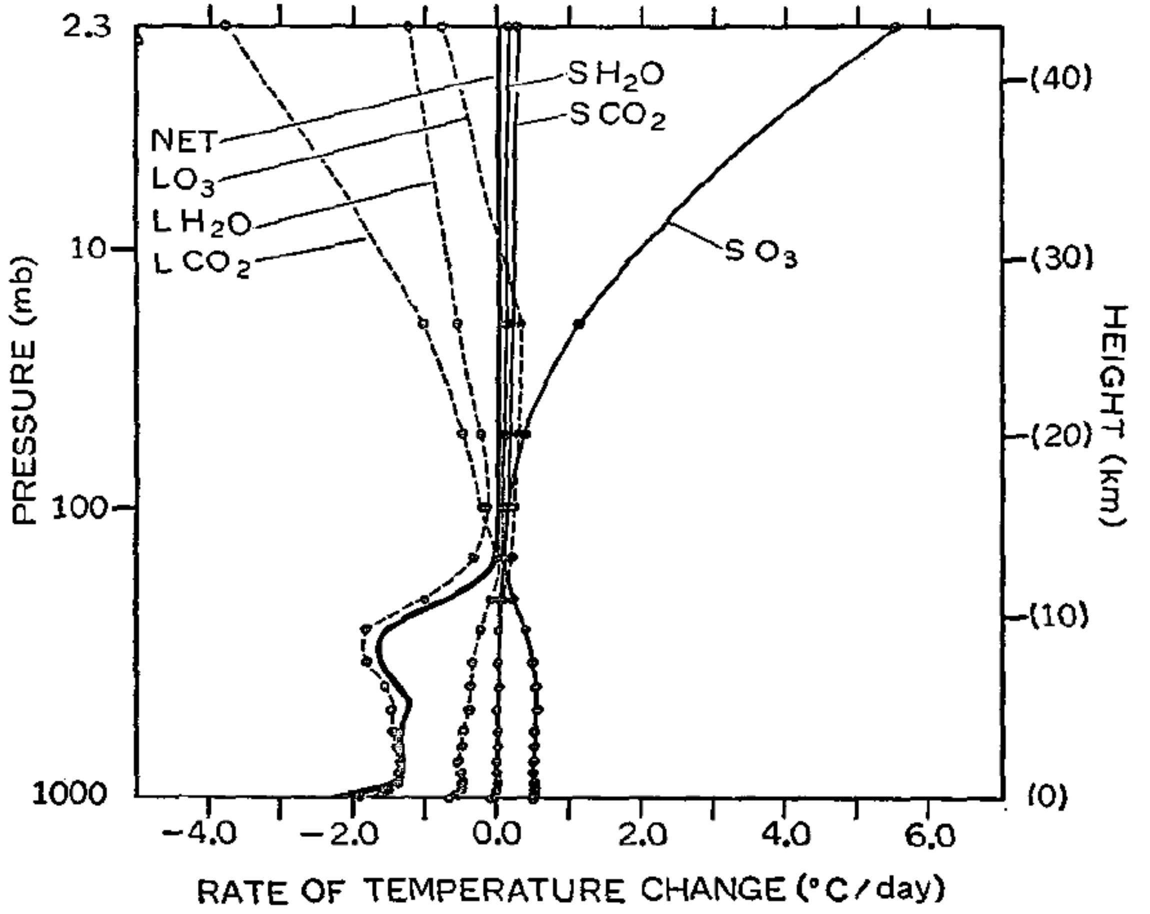


FIG. 8c. Vertical distributions of the radiative heat balance components for the thermal equilibrium of a clear atmosphere. Refer to Fig. 6b for further explanation.

Contributions to $\frac{dT}{dt}$ in radiative-convective equilibrium. (Manabe and Strickler, 1964).

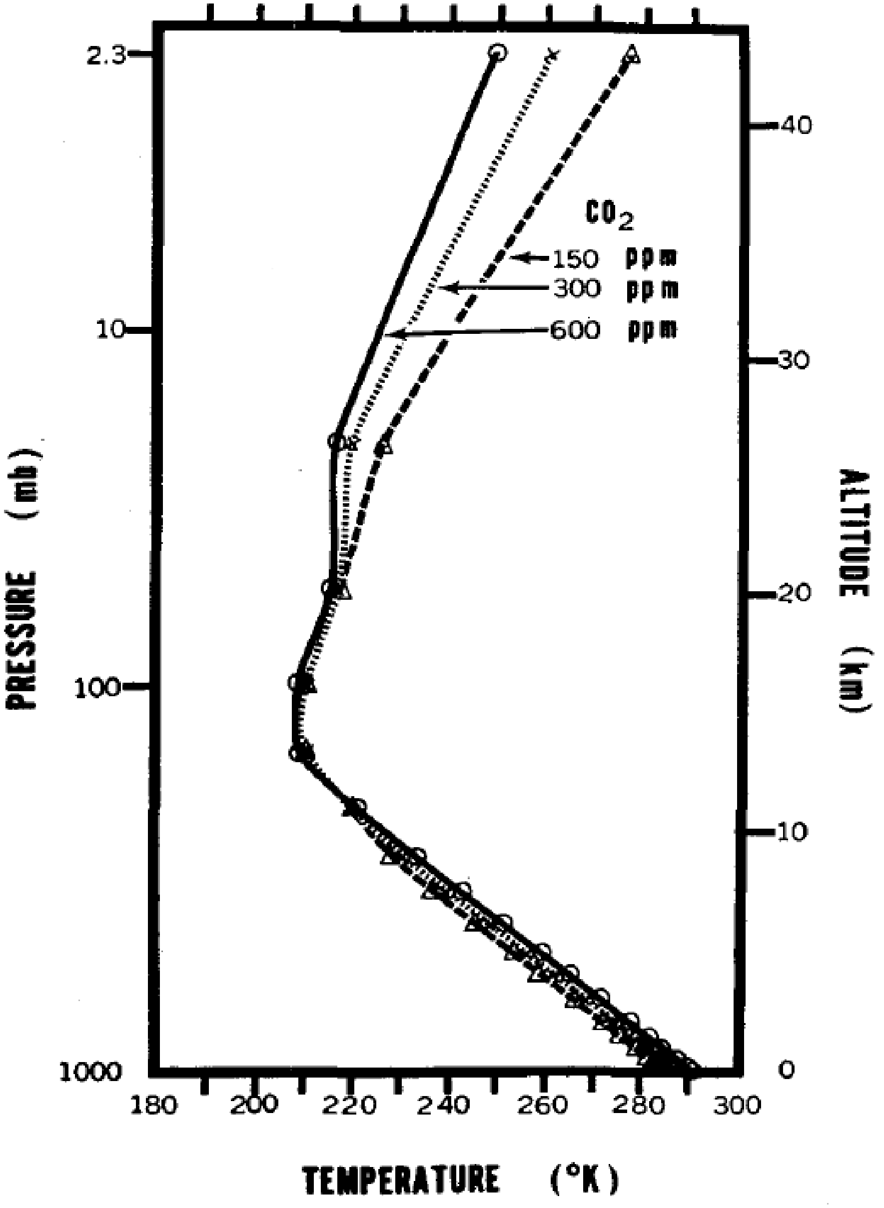


FIG. 16. Vertical distributions of temperature in radiative convective equilibrium for various values of CO_2 content.

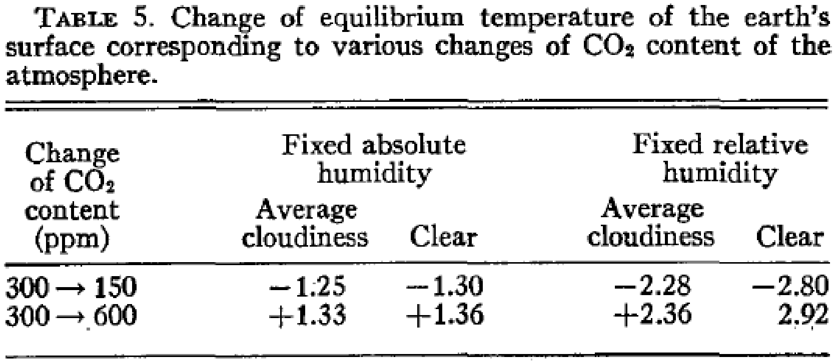


TABLE 5. Change of equilibrium temperature of the earth's surface corresponding to various changes of CO_2 content of the atmosphere.

Change of CO_2 content (ppm)	Fixed absolute humidity		Fixed relative humidity	
	Average cloudiness	Clear	Average cloudiness	Clear
300 → 150	−1.25	−1.30	−2.28	−2.80
300 → 600	+1.33	+1.36	+2.36	2.92

Radiative-convective equilibrium with various amount of CO_2. (Manabe and Wetherald, 1967).

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