Comments on "A Numerical Study of a Marine Subtropical Stratus Cloud Layer and its Stability"

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I have two comments on Moeng and Arakawa (1980, hereafter MA80), who have modeled the downstream evolution of a stratocumulus-topped mixed layer and the breakup of the stratocumulus by cloud-top entrainment instability.

First, the total change in net longwave radiative flux in the lower cloud (Fig. 5), which causes heating, appears about half as large as the total change in net longwave radiative flux near cloud top, which produces cooling. From Fig. 7 we estimate the jump in net longwave radiative flux across cloud top to be about 80 W m^{-2}. Assuming a quasi-steady state in which $\partial/\partial z(\partial h/\partial t) \approx 0$ in the boundary layer, with h the moist static energy, we expect the vertical turbulent flux of h ($\rho \overline{w'h'}$ in Fig. 7) to be linear if ra-

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diative divergence is neglected. Longwave radiative flux divergence in the boundary layer makes the $\rho\overline{w'h'}$ profile significantly nonlinear in MA80, although radiative flux divergence in the boundary layer is usually considered insignificant and ignored in trade-wind, mixed-layer models. By comparing the actual $\rho\overline{w'h'}$ in the upper cloud and an extrapolated value based on the linear, subcloud profile of $\rho\overline{w'h'}$, we find that $\rho\overline{w'h'}$ increases by about 40 W m^{-2} due to radiative warming in the lower cloud, or about half the change due to radiative loss at cloud top. Assuming that the lower cloud and the surface are black bodies, that there is an adiabatic lapse rate in the subcloud layer, and that the air–sea temperature difference is negligible, we can estimate an upper bound on the cloud-base increase in $\rho\overline{w'h'}$ (due to radiative warming) as $\sigma[T_0^4 - (T_0 - \gamma_d z)^4]$, where $\sigma = 5.7 \times 10^{-8}$ W m^{-2} K^{-4}, the surface temperature $T_0 = 295$ K, γ_d is the dry adiabatic lapse rate and z the height. For a cloud base at 750 m, the upper bound on cloud-base radiative warming is 42 W m^{-2}, which suggests the radiation model of MA80 has essentially no absorption in the subcloud layer. If the cloud base were at 1250 m, then the upper bound would be 69 W m^{-2}, comparable to the cloud top change of 80 W m^{-2}; however, for such long path lengths the only transmission should be in the atmospheric window (7–14 μm), which contains possibly 40% of the total longwave radiation, so these upper bounds should be divided by a factor of two and one-half (J. A. Coakley, personal communication, 1981). This would give smaller cloud-base changes in net longwave radiative flux. Also, the cloud-base changes would then be in agreement with values in the literature [for example, see Fravalo *et al.* (1981) and Sommeria (1976)]. Thus, the cloud-base warming of MA80 seems excessive.

If there were cloud-top radiative cooling but no cloud-base heating, then the falling, radiatively-cooled parcels would stir the entire boundary layer and the velocity variances in Fig. 6 would be approximately well-mixed throughout the planetary boundary layer. If there were just the large cloud-base radiative heating of MA80 and no cloud-top cooling, then the radiatively produced convection would be confined to the cloud layer (due to the stabilization across cloud base), as their results show. Thus, I conclude that in MA80 cloud-base radiative heating is possibly even more important than cloud-top cooling in determining their turbulence profiles.

My second comment is that MA80 seem to have used horizontal pressure gradients and hence mean wind speeds which were inappropriately low (by a factor of at least two and as much as six) for the region they modeled. This affects the horizontal evolution of the boundary layer. The planetary boundary layer initially is confined by an inversion across which potential temperature θ and equivalent potential temperature θ_e both increase. As the boundary layer evolves downstream, the surface fluxes of sensible heat and moisture increase θ_e in the boundary layer and the subsidence aloft may decrease θ_e just above the boundary layer. The jump $\Delta\theta_e$ at the inversion decreases and eventually becomes negative, which may lead to cloud-top entrainment instability. Thus, using a more appropriate, higher wind speed would alter the horizontal distance required before entrainment instability could occur (unless other parameters like the divergence were also changed).

In general, MA80 exaggerate the relative importance of cloud-top radiative cooling and cloud-base radiative heating, particularly in their turbulence, but also in their mean budgets. They seriously underestimate surface fluxes and hence, in the lower half of the boundary layer, the production of the total turbulent velocity variance $q^2 = \overline{u^2} + \overline{v^2} + \overline{w^2}$ by buoyancy and shear. Consequently their results are an upper bound on the magnitude of the effect of radiative heating and cooling on boundary layer turbulence.

The primary simulations in MA80 have two spatial dimensions, height z and horizontal distance y, where $y = 0$ at about 34°N, 135°W, and y increases following an air trajectory toward Hawaii (Fig. 40 in Neiburger *et al.*, 1961, shows an appropriate trajectory). In Fig. 3 of MA80, $y = 0$ on the right side corresponds to 34°N, 135°W (I will call it point A) and the left side is 1000 km downwind (point B). The wind speed is a maximum of about 3.5 m s^{-1} near the surface at point B and decreases vertically and horizontally moving away from point B; at point A, the horizontal velocity $V = 1$ m s^{-1}.

Wind speeds of 1–3.5 m s^{-1} are abnormally low for the North Pacific trade winds. I suspect that they referred to Fig. 35 in Neiburger *et al.* (1961), which is reproduced as Fig. 1 in Schubert *et al.* (1979). These figures show the magnitude of resultant or vector averaged winds which are much lower than average wind speeds. Fig. 4 in Schubert *et al.* (1979) shows that the average wind speed at 10 m varies from 6 to 7 m s^{-1} along the trajectory being modeled by MA80. One can also arrive at comparable estimates of the geostrophic wind (6–8 m s^{-1}) using the sea surface pressure patterns shown as Fig. 4 in Schubert *et al.* (1979) or Fig. 32 in Neiburger *et al.* (1961). Thus, at point A the winds of MA80 are about a factor of six too low and at point B about a factor of two low.

MA80's abnormally low wind speeds will probably not seriously modify the downstream evolution of mean temperature and humidity profiles due to surface fluxes alone. To show this, consider the downstream evolution of mixed-layer potential temperature due to the surface flux of potential temperature $\overline{w\theta_0}$, with

$$\overline{w\theta_0} = -C_{DH}V_{10}(\theta_{10} - \theta_0), \qquad (1)$$

where C_{DH} is a transfer coefficient for sensible heat flux, V_{10} is the wind speed at $z = 10$ m and θ_{10} and θ_0 are the potential temperatures at 10 m and the surface. The advective change in mixed layer θ due to $\overline{w\theta_0}$ alone in a horizontal distance Δy is

$$V\frac{\theta_B - \theta_A}{\Delta y} = \frac{\overline{w\theta_0}}{z_i}, \quad (2)$$

where z_i is the inversion height and θ_B and θ_A are the mixed layer θ's at points B and A. Substituting (1) in (2) yields

$$\theta_B - \theta_A = \frac{C_{DH}(\theta_0 - \theta_{10})\Delta y}{z_i}.$$

Thus, to first order, the effect of wind speed cancels for C_{DH} independent of wind speed. While a larger wind speed implies a larger surface flux, the larger flux has proportionately less time to act as the air column moves a fixed distance Δy downwind.

While the change (in a fixed horizontal distance) in mixed layer θ and mixing ratio due to surface fluxes is unaffected by wind speeds, these surface-flux-induced changes would be relatively more important if a higher, more appropriate wind speed were used, compared to changes due to other processes not dependent on wind speed. For example, radiative cooling of the mixed layer in a fixed horizontal distance would be decreased by a factor of three or four if a more appropriate wind speed were used.

To estimate the effect on turbulence quantities of using a larger, more appropriate wind speed, we let $V_{10} = 7$ m s^{-1} and the transfer coefficient for momentum flux $C_{DM} = 1.4 \times 10^{-3}$. Then the friction velocity $u_* \approx 0.26$ m s^{-1}. Using a lower boundary condition for neutral conditions of $q^2 = 7.5u_*^2$ (Wyngaard, 1975), shear production would produce an additional 0.5 m^2 s^{-2} at the surface, which would decrease with height but would strongly affect the q^2 profile shown in Fig. 6 of MA80. The surface sensible and latent heat fluxes would increase proportionally to the wind speed since the air–sea mean gradients are, to a first approximation, independent of wind speed. The larger surface buoyancy flux would strongly increase buoyant production of q^2 in the lower half of the boundary layer. The additional q^2 in the lower boundary layer could make the profile of q^2 in MA80's Fig. 6 roughly constant in the boundary layer. Likewise, increasing surface energy fluxes by a factor of two to six will substantially modify the flux profiles of MA80's Fig. 7.

In summary, by using inappropriately low wind speeds, MA80 are not accurately modeling the downstream breakup of stratus to cumulus in the North Pacific trades. Their low wind-speeds have artificially increased the relative importance of cloud-top radiative cooling and cloud-base warming in both the mixed layer potential temperature or equivalent potential temperature budgets and also in the mixed layer budget of q^2. In particular, their estimate of radiative warming at cloud base seems unrealistically large.

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