Drizzle Production in Stratocumulus

G. Feingold and A. S. Frisch
CIRA/National Oceanic and Atmospheric Administration
Colorado State University
Fort Collins, Colorado

B. Stevens and W. R. Cotton
Colorado State University
Fort Collins, Colorado

Introduction

Although stratocumulus clouds are not prodigious producers of precipitation, the small amounts of drizzle they do produce have an important impact on both cloud macrophysical properties (e.g., spatial coverage, depth and liquid water content) and microphysical properties (e.g., droplet size distributions, effective radii). The radiative effects of stratocumulus are intimately connected to both these macro- and microphysical properties, and it is thus essential that we understand the mechanisms of droplet growth which generate precipitation sized droplets.

Drizzle production is closely related to cloud condensation nucleus (CCN) number and size (see, for example, Albrecht 1989), as well as to cloud dynamics and the ability of clouds to support droplets within their bounds and allow for repeated collision-coalescence cycles (Mason 1952). In order to address both the microphysical and dynamical aspects of drizzle formation (and their close coupling), we have adapted a large eddy simulation (LES) model to include explicit (size-resolving) microphysical treatment of the CCN and droplet spectra (Feingold et al. 1994; Stevens et al. 1995). By directly calculating processes such as droplet growth by condensation and stochastic collection, evaporation, and sedimentation in the LES framework, we are in a position to elucidate the drizzle formation process.

Here we examine only one aspect of drizzle formation, namely, the relationship between it and a measure of the energy of the cloud (in this case root-mean-square velocity). We follow the classic work of Bowen (1950) and Mason (1952), who calculated trajectories of a large collector droplet rising in an updraft and collecting small droplets. Bowen showed that the size of the drop, as it exits the cloud, depends on the updraft velocity and cloud liquid water content. Mason argued that the turbulent motions in stratiform clouds will allow drizzle drops to form by increasing their dwell-time in the cloud. Nicholls (1987) followed this line of thought by studying stochastic drop coalescence within a stochastic turbulence model. This work differs from the aforementioned works in a number of respects. Firstly, we explicitly model droplet growth processes and the mechanisms through which some statistically fortunate droplets grow large enough to initiate gravitational collection. Secondly, given the fact that these droplets do exist, we examine how important the turbulent motion within the cloud is in determining the amount of precipitation produced. We allow full coupling between microphysics and dynamics in a multi-dimensional framework.

Model Description

To expedite processing of a large number of numerical experiments, we run the model in two-dimensions. We have used a single sounding, taken from the research vessel *Malcolm Baldridge*, June 16, 1992, 07:05 GMT, which produced a solid cloud deck and drizzle at the ground.

We examine the sensitivity of drop growth to cloud vertical velocity by artificially modifying the drop terminal velocities. The approach is justified by the fact that the factor of importance is the velocity of the drops relative to the cloud velocities. The net vertical motion of a droplet is given by the sum of the local updraft velocity w and the drop terminal velocity V_T. Thus, the simulation of a more vigorous cloud with, say, $w \pm \in$, is equivalent (in terms of w_{net}) to the simulation of that cloud with the same w but terminal velocity $V_T \mp \in$. Based on this principle, we can simply add or subtract an amount $\in$ to each drop terminal velocity to simulate less or more vigorous clouds. There is, however, a practical problem in the implementation of this approach—when subtracting $\in$ from V_T, we find that for $V_T \leq \in$ the applied fall velocity becomes negative, and droplets fall upwards, eventually reaching the dry above-cloud air where they rapidly evaporate. Thus, we have assumed that all drops with V_T smaller than $\in$ move with the cloud motions. (For symmetry, we have applied this rule for all cases.)

We note that application of corrections to fall velocities has the added benefit of providing insight into the extent to which bulk microphysical parameterizations would misrepresent the drizzle process with an erroneous choice of drop terminal velocity within the bounds $V_T - \in$; $V_T + \in$.

Results of the Bowen Model Without Dynamical Feedback

We ran our base case assuming that all droplets with r < 0 µm ($\in$= 50 cm s^{-1}) move with the air motions, while drops larger than 60 µm fall at their correct V_T. This value of $\in$ is somewhat arbitrary, but not restrictive in the sense that results vary only quantitatively for different choices of $\in$. We then repeated this run but subtracted or added an $\in$ of 50 cm s^{-1} from V_T ($V_T - \in$, or $V_T + \in$) for r > 60 µm. These cases represent a more (or less) vigorous cloud with an enhanced (or diminished) ability to maintain droplets in the cloud. To isolate the effect of the drop terminal velocity relative to cloud updraft velocity, we imposed on the $V_T - \in$ and $V_T + \in$ runs the identical cloud vertical velocities as produced for the V_T run. The differences in drizzle production between these cases are summarized by Figure 1. Drizzle rate (R) for the $V_T - \in$ run is more than double that for the V_T run, while for the $V_T - \in$ run, it is less than half of that in the V_T run. Analysis of radar reflectivity shows that the precipitation reaching the ground in the $V_T + \in$ run is in the form of larger drops, while in the case of the $V_T + \in$ run, it is in the form of small drops. When drops have reduced fall velocities (equivalent to more vigorous clouds), re-circulation in the cloud produces more drizzle, comprising larger drops. The converse is true for the case where drops have enhanced fall velocities (equivalent to less vigorous clouds).

Results of Bowen Model with Dynamical Feedback

We repeated the experiments $V_T - \in$ and $V_T + \in$ in Section 3a but this time allowed feedback to the dynamics (i.e., w was prognosed correctly). Figure 2 shows the same trends for R observed in Figure 1, although the degree of enhancement in R for the $V_T - \in$ run is reduced. Therefore the dominant effect is the fall velocity of the drops relative to cloud vertical velocity, and feedbacks are not sufficient to erase this effect. In Figure 3, we show

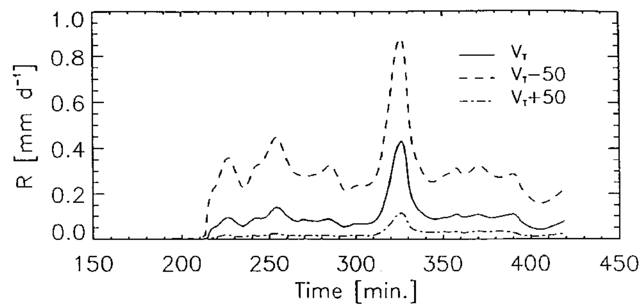

Figure 1. Temporal evolution of R [mm d^{-1}] for the V_T, $V_T - \in$, $V_T + \in$ cases: fixed dynamics.
profiles of the

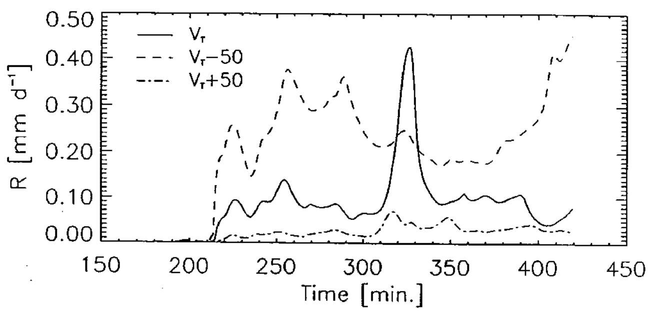

Figure 2. Temporal evolution of R [mm d^{-1}] for the V_T, $V_T - \in$, $V_T + \in$ cases: with feedback to dynamics.

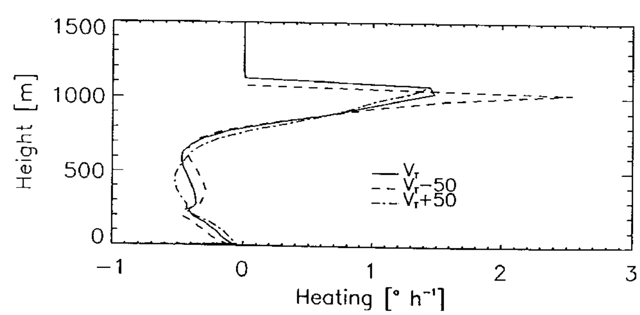

Figure 3. Profiles of heating associated with the drizzle flux for the V_T, $V_T - \in$ and $V_T + \in$ cases. The $V_T - \in$ run shows stabilization relative to the V_T run, while the $V_T + \in$ run shows relative destabilization.

heating associated with the drizzle (calculated from the divergence of the drizzle flux). The profiles show that in the case of the $V_T + \in$ run, most of the cooling associated with evaporation was confined to the region below cloud base (200-500 m), while the $V_T - \in$ run shows cooling spread over the depth of the sub-cloud layer. Relative to the V_T run, the $V_T + \in$ run produces stronger destabilization of the boundary layer and tends to produce deeper circulations. In contrast the V_T - $\in$ run produces a relative stabilization of the boundary layer; cooling in the region just below base is weaker than in the V_T run, while cooling near the ground is stronger. The resulting circulations tend to be weaker, and confined to two levels. We note that it is the *distribution* of this cooling with height that is critical to generating the stronger deeper circulations.

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