Upstream influence of numerically simulated squall-line storms

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SUMMARY

The squall line’s impact on its upstream environment is examined using traditional cloud and simplified (parameterized moisture) models. The study was motivated by the need to explain significant differences between the two dynamical frameworks. In both, the first environmental response to the convection consists of a rapidly propagating gravity wave characterized by deep tropospheric subsidence. This gravity wave accelerates the low-level storm inflow in its wake, with largest effect seen near the surface.

In the more sophisticated model, however, this wave is eventually followed by a shorter vertical wavelength feature, one possessing lower tropospheric ascent. This second gravity wave, absent from the simplified model runs, substantially alters the upstream environment yet again. The gentle low-level uplift establishes the “cool/moist tongue” of air that has been found to stretch well ahead of the storm in simulations. Between the second wave and the main storm updraft, the accelerated inflow is shifted from the ground to the middle troposphere where it helps to push dry air into the convecting region.

This subsequent environmental adjustment responds to the establishment of a small yet persistent area of weak cooling during the early mature phase. Located in the middle troposphere at the upstream edge of the main cloud mass, this cooling is the combined effect of cloud water evaporation and the ascent of subsaturated air. Though the cooling that excites it is of relatively small magnitude (< 2 K), this wave’s effect is dramatic and significant. A crude fix to the moisture parameterization served to establish a qualitatively similar tongue-like feature in the simplified model simulations, bringing the results from the two dynamical frameworks more into line.

KEYWORDS: Squall-lines Parameterized moisture models Multicell storms Gravity waves

1. INTRODUCTION

Traditionally, cloud models have assumed horizontally homogeneous and temporally invariant base state conditions and have initiated convection in relatively crude manners. The former simplifications can be advantageous for dynamical studies as any alterations which ensue in the storm’s surrounding environment must be attributed to the storm (or model) itself. Despite this, traditional cloud models are still very nonlinear and complicated. Much of this is due to the model microphysics, including water substance phase changes, conversion rates, hydrometeor fallspeeds, etc.

Significant simplification could be realized by completely parameterizing moisture and its effects. However the parameterized moisture (PM) model’s usefulness would be severely limited if it cannot generate reasonably realistic circulations. Fovell and Tan (2000; hereafter “FT00”) assessed the performance of a PM model formulation based on, and inspired by, Garner and Thorpe (1992; hereafter “GT”). This PM model made latent heat release conditionally proportional to ascent velocity and used a heat sink of specified size and intensity to generate and maintain the subcloud cold pool.

FT00 showed the PM model can support realistic, sustained “multicellular” behavior in which storm longevity is assured by the sequential generation of short-lived convective cells. Multicell storms are ubiquitous in nature as well as among traditional cloud model simulations. The PM model’s ability to capture this behavior lent support to Fovell and Tan’s (1998; hereafter “FT98”) analysis of the sustained transience which minimized the role of microphysical details in causing the unsteadiness.

Less realistically, the PM model storms were found to exert a dramatic and permanent influence on the lower troposphere on their upstream sides, representing the

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environment into which the storms were propagating. The mature phase low-level storm inflow was considerably intensified in the FT00 and GT simulations, with the largest enhancement found very near the ground. This occurred with both low and moderate CAPE soundings, in multicellular as well as nearly steady cases. In the moderate CAPE case that FT00 highlighted, for example, the surface inflow changed by \(10 \text{ m s}^{-1}\) subsequent to convective initiation (see FT00 Fig. 7). In contrast, typical full physics simulations evince little inflow enhancement in the lowest 2 or 3 km or so (Fovell and Ogura 1988) during maturity.

Certainly, the convection is expected to cause some upstream modification. A fraction of the storm updraft mass overturns to form the forward anvil, and mass continuity dictates some inflow enhancement must accompany that upper tropospheric outflow. However, this influence seemed exaggerated and thus FT00 tried to isolate the apparent flaw in the PM framework. FT00 focused on mechanisms that might account for a PM model tendency to generate excessive anvil outflow strength. They suggested that excess warming in a mature PM storm’s trailing region, a consequence of the PM model assumptions, might be the culprit.

In this paper, we more closely examine how and why the storm’s upstream environment becomes disturbed by convection. Detailed intercomparisons of simulations made with the PM and full physics models are made, with particular attention paid to how the mature storm structures come about. Further analysis of FT00’s results revealed that the more significant difference between the two models resided in the vertical structure, rather than the magnitude, of the upstream disturbances. The root of this discrepancy is identified and confirmed via a practical “fix” to the PM framework. However, the main intent of this study is to further elucidate the effect convection can exert on its immediate surroundings, and employment of two different model frameworks in juxtaposition serves this end.

2. Comparison of the traditional and parameterized moisture cloud models

In this section, the environmental response to convective heating in the PM and traditional models are described and compared. The storms in this paper propagate towards the east, and the terms “downstream” and “upstream” are synonymous with “west” and “east”, respectively. The upstream side is also downshear relative to the initial vertical shear vector. The horizontal velocity field \(u\) will be presented in domain-relative (\(u_{\text{dom}}\)) and ground-relative (\(u_{\text{grnd}}\)) perspectives and will also be expressed as a perturbation from the initial state (\(u’\)). The domain is translated and thus \(u_{\text{dom}}\) is usually closely approximates storm-relative velocity (\(u_{\text{strm}}\)). The perturbation field is independent of the reference frame. The ground-relative initial wind is taken to be calm at the model surface.

(a) A PM model simulation

The PM framework (see Fig. 1) was described in detail in FT00. Briefly, the chief effects of moisture are parameterized through the inclusion of terms representing vapor-condensation warming and liquid-water evaporation, labeled \(Q^+\) and \(Q^-\), respectively. These terms appear in the model’s perturbation potential temperature (\(\theta’\)) equation as

\[
\frac{d\theta’}{dt} = Q^+ + Q^-.
\]

This perturbation is relative to a base state which is a function of height alone.
The evaporation cooling term $Q^-$ is treated as a shallow, lower-tropospheric cooling zone dimensioned $x_h \times z_h$ in which air is continuously relaxed to the preset potential temperature perturbation $\theta'_c$ with a designated time scale $\tau_c$. The sink’s upstream edge is positioned a small distance $\delta$ behind the storm’s gust front, subsequent to that feature becoming established. As per usual, model domain translation attempts to keep the gust front as stationary as possible, but domain-relative motion can still occur in some cases. The cooling zone’s alignment relative to the front is examined every time step, and shifted if necessary to keep the two properly oriented.

In the PM model latent heat release is made proportional to ascent velocity ($w > 0$) by presuming air rises moist adiabatically within the “cloud”. Designating the slope of the moist adiabat in $(\theta, z)$ space as $\gamma^*$, $Q^+$ is handled as

$$Q^+ = \gamma^* \max(w, w_0)$$

where heretofore $w_0 = 0$. The “unstable region”, the subdomain in which $\gamma^*$ is nonzero, resides between height levels $z_m$ and $z_t$ (Fig. 1). Further, the region stretches laterally rearward to the domain’s downstream boundary but its upstream extent is truncated a small distance $\epsilon$ ahead of the surface gust front’s position. FT00 reported this truncation had virtually no effect on their simulations; we explain this result presently. As in FT00, both $\delta$ and $\epsilon$ are taken to be 5 km.

Figures 2 and 3 present typical multicellular PM model simulations, emphasizing the organizational period during which the environmental alteration develops. These examples employed FT00’s low and moderate instability soundings, having 400 and 1800 J kg$^{-1}$ of CAPE, respectively (see Fig. 1 and FT00’s Fig. 4). Note the ambient stability of the low CAPE sounding is also quite small; the moderate CAPE case, based on FT98’s initial environment, possesses a realistic background state.

Middle tropospheric warming, manifested as positive potential temperature perturbations ($\theta'$), is seen spreading outward from center of convective activity in both simulations. The induced horizontal velocity perturbations ($u'$) are directed away (towards) the convective region above (below) the level of maximum warming. On the upstream (east)
Figure 2. Low CAPE PM model simulation, having moderate low-level shear of $3.15 \times 10^{-3}$ s$^{-1}$ below 3 km. (a)-(d) show perturbation horizontal velocity ($u'$; 2 m s$^{-1}$ contours) and perturbation potential temperature ($\theta'$; shaded); (e)-(h) present domain-relative horizontal velocity ($u_{dom}$; 3 m s$^{-1}$ contours) and vertical velocity ($w$; shaded). Only a portion of the model domain is shown, and horizontal distance is measured from the west edge of the fine grid region; see FT00. Domain was translated eastward at 11.5 m s$^{-1}$. 
side of each case, the largest outflow is located at the level at which the temperature perturbations vanish (about 9-10 km). Note that the strongest inflow is located just above the model’s free-slip bottom surface.

The vertical velocity ($w$) fields in Figs. 2 and 3 reveal that the leading edge of the environmental adjustment, GT’s “storm-front”, is marked by deep subsidence. Focusing on the upstream environment, the subsidence is seen to propagate eastward away from the convection, leaving in its wake a permanently altered flow field. Note the dashed contours of domain relative horizontal velocity ($u_{dom}$) are “lifted” by the disturbance’s passage, never to settle back to their original positions. In between the convection and the storm-front, transience quickly disappears. While the environmental response is quicker to develop, has larger amplitude and faster propagation in the moderate CAPE/realistic stability case, there is substantial qualitative resemblance between the two environments.

Figure 3 presents a vertical profile of the ground-relative wind ($u_{grnd}$) taken 10 km ahead of the surface gust front position for the low CAPE case, along with the initial profile. Note that easterly perturbations, representing enhanced storm inflow, are present below 5 km, with the largest values at the ground. This profile was taken at $t = 10800$ sec; the simulation had become statistically steady by this time, at least in the vicinity of the convection (and behind the spreading storm-front). The moderate CAPE run will be examined further in Section 4.

(b) A traditional cloud model simulation

Now we turn to a rather typical multicellular simulation made with a traditional cloud model, specifically the ARPS model (Xue et al. 2000). The simulation, being FT98’s
two-dimensional case, started with a moderate CAPE ($\approx 2500 \text{ J kg}^{-1}$) environment with a vertical shear of $3 \times 10^{-3} \text{ s}^{-1}$ in a 2.5 km deep layer. Thus, the initial ground-relative flow above the shear layer was 7.5 m s$^{-1}$. The domain speed was 12 m s$^{-1}$ during the time period depicted. Please refer to FT98 for additional information.

At first, the environment’s response to the convection is rather similar to that seen in the PM model. Figure 5 presents $w$ and $u_{dom}$ fields sampling the early portion of the model storm’s organizational period. As in the PM model simulation, the initiation of convection provokes rapidly spreading subsidence waves. In this case, the upstream-propagating feature’s domain-relative motion was about 33 m s$^{-1}$, making for a 45 m s$^{-1}$ ground-relative signal*. The ground-relative speed for the downstream side’s wave was about 30 m s$^{-1}$. As shown in Fig. 6, which presents $\theta'$ and $u'$ fields for these same three times, the subsidence waves are again marked by warming. In the upstream-propagating feature’s wake are westerly and easterly horizontal wind perturbations, present above and below the location of maximum warming, respectively.

The principal difference between the PM and traditional model simulations at this point concerns transience in the model storm upstream environment. The full physics storm’s main updraft underwent dramatic fluctuations between vertical and downshear (upstream) tilted orientations, exciting a sequence of similar, semi-discrete wave-like features (see especially Fig. 5c) that propagated through the upstream environment. This was possibly suppressed in the PM model by the nature of the moisture param-

* This was not the fastest moving signal in this early period; much faster (but also much smaller amplitude) features are evident in $u_{dom}$ contours’ kinks.
Figure 5. Early organizational period of the full physics model simulation. Shown are $u_{dom}$ (2.5 m s$^{-1}$ contours) and $w$ (shaded) for a portion of the model domain. Upstream and downstream movement of the initial subsidence wave is tracked and labeled with ground-relative propagation speed. Horizontal distance is measured from the west boundary.

The main point is that the ground-relative perturbations at this time are easterly and westerly in the lower and upper troposphere, respectively, with the largest inflow enhancements found very near the surface.

Figure 7 shows $\theta'$ and $u'$ fields for the period 3900-4500 sec. By 3900 sec (Fig. 7a), the subcloud cold pool was well formed and the storm’s rising front-to-rear airflow had already acquired the characteristic upshear (downstream) tilt superposed with multicel-
Figure 6. As in Fig. 5, but showing $u'$ (2 m s$^{-1}$ contours) and $\theta'$. 

lular transience it would retain for the balance of the simulation. This figure roughly spans one cell generation period and focuses more narrowly on the storm’s leading edge. The deep waves generated earlier have already moved upstream out of the subdomain depicted although warming and westerly $u'$ perturbations still occupy much of the upper troposphere. Large easterly perturbations are still present upstream of the convection but, importantly, the level of maximum inflow enhancement has shifted to the middle troposphere ($z \approx 5$ km). The near-surface perturbations just ahead of the storm are now rather small.

Note also the presence in all three panels of a tongue-like feature representing small but sustained amount of midtropospheric cooling stretching across the storm’s leading
edge (284 ≤ x ≤ 292 km). This feature will be designated the cool/moist tongue since it will presently be shown that its air has also been moistened relative to the initial state. Both the local cooling and nearby enhanced inflow are disturbed as the cell passes through its life cycle but appear to change in unison. The maximum inflow perturbation remains rooted just above this tongue. Though not shown, the cool/moist tongue also occurs in less geometrically constricted 3D simulations as well.

Figure 8 provides a closeup view of the storm’s leading edge, at a time one minute prior to that of Fig. 7b. The $\theta'$ and $w$ fields (Fig. 8a) show the cool/moist tongue forming a “cap” hugging the top of the developing cloud and its updraft. This air has been moistened as well: Fig. 8b reveals positive water vapor perturbation ($q'_v$) values

Figure 7. As in Fig. 7 but for a period straddling the organizational and mature phases.
throughout the tongue area.

The bottom two panels, along with Fig. 8b’s equivalent potential temperature (θ_e) distribution, suggest that subsaturated air has been rising in the vicinity of the developing cloud and has been chilled by both adiabatic expansion and cloud water evaporation. The latter, along with the general uplift, accounts for the moistening. Figure 8c shows the combination of the vertical potential temperature advection (VPTA), expressed as

\[ VPTA = -u \frac{\partial \theta}{\partial z} \]  

and the cloud water diabatic term, \( S(q_c) \), for this instant of time*. The net effect of vertical advection and cloud-vapor phase changes is cooling through the tongue area. In Fig. 8d the \( S(q_c) \) field is shown in isolation. Mixing of subsaturated air into the cloud continues to erode the cloudy area, resulting in part of the cooling seen.

The midtropospheric cooling continues to spread horizontally in the upstream direction over the next several hours. The times shown in Fig. 9 were chosen to represent roughly the same point within the multicellular repeat cycle. The leading edge of the cool/moist tongue is marked by a weak, propagating updraft (see Fig. 9a). The enhanced midtropospheric inflow also spreads, in lockstep with the cooling; the correlation between these features will be explored in the next section. As these features widen, the upstream environment in the immediate vicinity of the convection settles into an essentially statistically steady mature state.

The vertical profile of \( v_{grnd} \) characterizing the upstream environment during this mature phase is shown in Fig. 10. In pointed contrast to the PM model results, the easterly wind perturbations are centered around 5 km, and only very minor alteration to the storm’s immediate low-level inflow can be noted. A very similar profile was found in Fovell and Ogura’s (1988; Fig. 13) moderate intensity squall-line; elevated upstream inflow also appears in the midlatitude cases simulated by Schmidt and Cotton (1990; Fig. 11) and Nachamkin and Cotton (2000; Fig. 10) as well as in Lafore and Moncrieff’s (1989; Fig. 4) tropical setting. A similar upstream response is even noted in very strong, nearly erect storms such as Fovell and Ogura’s (1989) most highly sheared cases (not shown); FT98’s 3D storm during maturity was also strongly similar (also not shown).

3. Interpretation of these results

(a) Results from analytic investigations

Nicholls et al. (1991, hereafter NPC) investigated the environmental response to maintained heat sources in an idealized setting. Their examination extended previous studies by Lin and Smith (1986) and Bretherton and Smolarkiewicz (1989), among others, and was itself further amplified by Mapes (1993). Using analytical solutions of the 2D linearized and hydrostatic Boussinesq equations, NPC demonstrated that a vertically oriented source immediately provokes subsidence of depth comparable to the source which then rapidly propagates away. This feature, clearly GT’s storm-front, is considered a gravity wave. It may be followed by a sequence of successively slower moving gravity waves which in combination constitute the aggregate environmental adjustment to convection.

The top two panels of Fig. 11 qualitatively depict the adjustment of an initially quiescent environment to a symmetric and vertically oriented heat source, based on

* This combination comprises most of what FT98 termed “TVPT” (see their Fig. 9), except for rainwater evaporation which is excluded here to highlight the cloud water evaporation contribution. There is no rainwater for \( x > 281 \) km anyway.
Figure 8. Close-up view of the storm leading edge at 4140 sec. (a) $w$ (2 m s$^{-1}$ contours) and $\theta'$ (shaded); (b) perturbation water vapor mixing ratio ($q'_c$; 1 g kg$^{-1}$ contours) and equivalent potential temperature ($\theta_e$; shaded); (c) vertical potential temperature advection (VPTA) and cloud water diabatic heating/cooling [$S(q_c)$] combination (0.0075 K s$^{-1}$ contours); (d) $S(q_c)$ (0.005 K s$^{-1}$ contours). Cloud outline (0.1 g kg$^{-1}$ mixing ratio) shown as grey broken curve is shown on panels (a), (b) and (d). Cool/moist tongue superposed on panels (b)-(d). Thin contours superposed on shaded fields in panels (a)-(c) to highlight negative values.
information presented in NPC Figs. 3 and 5, and Mapes’ Fig. 3. When the heating function possessed a single vertical mode (Fig. 11a), being a half-sine profile of depth $H$, this subsidence wave was the sole disturbance excited in the surrounding environment (at least in their analytic, linear and rigidly capped solution). This response had a vertical wavelength $L_z$ of $2H$ and its intrinsic phase speed was well approximated by

$$c = \frac{NL_z}{2\pi},$$

where $N$ is the tropospheric Brunt-Vaisalla frequency, and $\pi = 3.14159\ldots$. As this feature progressed away from the source, upper tropospheric outflow (inflow) in the upper (lower) troposphere remained in its positively buoyant wake stretching between the storm-front
and the heat source. The largest enhanced flow towards the source was at the surface.

When other vertical modes were incorporated into this heating profile, additional environmental responses were excited. Specifically, a second vertical mode having heating and cooling each of depth $H/2$ was found to provoke a wave-like response having half the vertical wavelength and phase speed of the initial wave, the latter being given by (4). Thus, this second feature propagated through the wake of the initial subsidence wave, modifying its predecessor’s environmental adjustment. NPC arranged their second mode to augment the upper tropospheric heating of the first mode while introducing diabatic cooling into the lower troposphere. Mapes (1993) showed that a qualitatively similar response could be obtained using source profiles having only positive values. In this scenario (depicted in Fig. 11b) the lower tropospheric cooling still occurs, but it originates from adiabatic expansion cooling locally exceeding the applied heating.

However provoked, the shorter wavelength feature possessed ascent confined to the lower troposphere. In its own wake, the previously established upper tropospheric outflow became stronger and more concentrated. More importantly, the enhanced inflow was shifted up to the middle troposphere. As shown in Fig. 11b, the strongest inflow was located where $\theta' = 0$, at the top of the adiabatically cooled layer.

These schematics were based on linear analytic solutions made using a vertically oriented heat source in a calm, constant $N$ environment and a rigidly topped domain. Figure 11c presents a nonlinear numerical solution for a somewhat more realistic situa-
Environmental response to maintained heat sources. Top two panels qualitatively depict response to symmetric heating functions with (a) one and (b) two vertical modes, drawn from Nicholls et al. (1991) and Mapes (1993); only upstream side is drawn. Panel (c) presents numerical result for a less idealized situation. Heating functions shown at right; for (c) the function is spatially averaged over the vertically tilted source region.

Figure 11. Environmental response to maintained heat sources. Top two panels qualitatively depict response to symmetric heating functions with (a) one and (b) two vertical modes, drawn from Nicholls et al. (1991) and Mapes (1993); only upstream side is drawn. Panel (c) presents numerical result for a less idealized situation. Heating functions shown at right; for (c) the function is spatially averaged over the vertically tilted source region.

Even with these complications, the essential tropospheric response remains unchanged. It is noted that the addition of a cold pool for $x < 270$ km, with maximum cooling at the ground, could have eliminated the shallow layer of source-relative outflow solution, made using a nonlinear and nonhydrostatic model. The initial conditions come from FT98, so the atmosphere is sheared as well as multiply layered, with an appreciably deep stratosphere. The heat source has been tilted towards the west with height, mimicking the orientation of the storm’s characteristic front-to-rear flow. The heating function itself represented a compromise between FT00’s low- and high-CAPE environments. Perturbations from the initial horizontal wind field (i.e., $u'$) are shown.
that developed in the lower troposphere just ahead of the main updraft. This would have made the solution shown even more realistic with respect to the lower tropospheric upstream response, an expectation consistent with Pandya and Durrán’s (1996) Fig. 20.

(b) Application to the traditional cloud model simulation

Figure 12 presents a Hovmöller diagram of $\Delta u$, defined here as the difference between $u'$ values at the 5.4 and 0.1 km height levels, for the FO98 ARPS simulation. The 5.4 km level was chosen not only because it roughly corresponds to the height at which $u' = 0$ immediately following the initiation of convection (see Fig. 7) but also it is the level at which the $u'$ inflow maximum subsequently occurs. Thus, a positive $\Delta u$ occurs when the induced inflow maximum is close to the ground, and negative values obtain with elevated inflow. The reference frame is translating eastward at 12 m s$^{-1}$. During maturity, the model storm propagation speed is 15 m s$^{-1}$ (3 m s$^{-1}$) relative to the ground (domain). All speeds labeled on the figure are ground-relative.

The deep subsidence waves, the environment’s first sizable reaction to the convection, are seen to spread quickly in both directions away from the initial disturbance. Again (see Fig. 5), the ground-relative propagation speeds were 45 and 30 m s$^{-1}$ for the eastward and westward moving features, respectively, values which correspond to a intrinsic phase speed of 37.5 m s$^{-1}$. (Above the 2.5 km deep shear layer, initial $u_{gnd}$ is 7.5 m s$^{-1}$.) Referring to (4), this intrinsic speed is consistent with a gravity wave having a vertical wavelength roughly twice the tropospheric depth $H$ of 12 km$^*$. During this period, the storm updraft is powerful and extends vertically through nearly the entire troposphere.

The $\Delta u$ values between the initial upstream wave and the convection are positive, reflecting the fact that the inflow enhancement is located in the lower troposphere. Progressively slower moving features, representing waves having shallower vertical structures, spread out in this original wave’s wake. After an hour, negative $\Delta u$ values appear in the vicinity of the convection and subsequently spread eastward. This represents the permanent shift of the enhanced inflow to the middle troposphere.

Based on the foregoing, we interpret this second feature as representing the environment’s response to the essentially permanent adiabatic cooling that appears on the convective updraft’s upstream side (as seen in Figs. 7 and 9). The ground-relative phase speed of this wave’s leading edge is about 22 m s$^{-1}$. While the presence of the lower tropospheric shear layer is likely a complicating factor in this case, it is reasonable to presume this value represents an intrinsic phase speed of $\approx 16$ m s$^{-1}$, corresponding to a source with a vertical half-wavelength of about 5 km, about the depth over which the adiabatic cooling extends in the vicinity of the storm updraft. Behind this second principal feature, which slows somewhat as it spreads outward, the upstream environment settles into a statistically steady state.

The upstream propagation of this wave’s updraft spreads the midtropospheric cooling upstream and moistens the lower troposphere. The presence of a cooled, moistened lower tropospheric layer extending well upstream of the storm during maturity has been noted in squall-line simulations in the past (e.g., Fig. 11 of Fovell and Ogura (1988), Lafore and Moncrieff’s (1989) Fig. 21a and FT98’s Fig. 15). Observations may be more scarce, but at least Hoxit et al.’s (1976) analysis of a Nebraska squall line evinced a clear pre-squall elevated moist cool/moist tongue (their Fig. 4). FT98 remarked on the role of the elevated moisture in the triggering of new convective cells. The “daughter clouds” they identified in the storm’s immediate upstream environment resided in air brought to

* Incidentally, the dependence in (4) on $N$ explains why the signal propagated much more slowly in the FT00’s low CAPE, low stability sounding.
saturation by cooling and moistening within the tongue.

4. Evaluation and revision of the PM framework

In effect, the upstream environment in the more sophisticated model experiences two separate, successive adjustments to the establishment of convection, the first in response to the initial burst of deep tropospheric heating, and the second occurring once the principal storm updraft starts canting upshear. The PM model storms also tilt upshear, so it is not the tilt itself that is crucial. Rather, the subsequent adjustment proceeds once
small yet sustained midtropospheric cooling appears in the vicinity of the convection, on the upstream side of the principal storm updraft. Like the first, the subsequent adjustment substantially and permanently modifies the upstream environment, this time effectively displacing the enhanced storm inflow to the middle troposphere.

This second adjustment is heretofore missing in the PM model owing to its inability to establish the localized cooling. In the PM model, all rising motion within the unstable region is presumed to be saturated, thus generating local warming, independent of the origin of the parcels being displaced. Thus, as presently formulated, the PM model is unable to detect which parcels are likely candidates for saturation and warming generation upon ascent, and which would fail to do so. Moreover, there isn’t any real cloud water to evaporate.

We have modified the moisture parameterization so that it roughly captures this small but crucial localized cooling. In so doing, we exploit our ability to control this cooling’s presence in the PM model, thereby further supporting for our analysis. Since the cooling typically appears on the periphery of the cell updrafts where vertical velocities are upward though small (Fig. 8), revising \( w_0 \), the threshold vertical velocity for triggering parameterized warming in (2), to be a small positive number (rather than exactly zero) seems to represent the simplest and most straightforward solution. This is an attempt to crudely mimic both adiabatic cooling of rising subsaturated air as well as the evaporation cooling owing to cloud water detrained into that air.

Figure 13 demonstrates the modification’s influence when FT00’s moderate CAPE environment is employed. The upper panel shows \( \theta' \) and \( u' \) fields as in Fig. 3 but at a later time, during the statistically steady mature phase. The bottom panel presents the modified PM model storm fields occurring during this same period. The empirically determined \( w_0 \) value used was 1.75 m s\(^{-1}\). This value was large enough to have a demonstrable effect without impairing the storm’s viability. Increasing \( w_0 \) effectively renders some fraction of the convective instability inaccessible to the storm. We were unable to find a threshold value that permitted the much more fragile low CAPE environment to continue supporting deep convection.

Note the modified PM model simulation sports a shallow tongue of chilled air, in roughly the expected position. This localized cooling has resulted in an elevation of the enhanced inflow and a generally more realistic simulation overall, a conclusion also supported by the vertical \( u_{\text{grand}} \) profiles of Fig. 14. There are some differences between these two PM simulations. Raising the triggering threshold reduces the total amount of tropospheric warming in the domain, both within and surrounding the convective updraft. With less heating aloft, the cold pool cooling parameter (\( \theta'_c \)) had to be reduced to obtain roughly the same storm speed of \( \approx 15 \) m s\(^{-1}\). These differences do not appear to invalidate our conclusions.

The present analysis explains why unstable region truncation had virtually no effect on FT00’s simulations. Truncation was intended to prevent parameterized warming in the upstream environment where little if any saturated ascent would be expected. No significant ascent occurred there anyway, at least not in the FT00 cases. In the modified PM framework, however, unstable region truncation may help the upstream environment properly respond to elevated cooling near the convection.

5. Discussion and conclusions

The impact of organized convection on its upstream environment has been examined using a traditional cloud model as well as a dramatically simplified parameterized moisture (PM) model owing to Garner and Thorpe (1992; “GT”) and Fovell and Tan (2000;
The study was originally undertaken to explain the differences in mature phase upstream inflow characteristics noted by FT00.

In both dynamical frameworks, the initial burst of convective heating excited a pair of rapidly propagating gravity waves that were characterized by deep tropospheric subsidence and left the environment surrounding the convection positively buoyant. In the wake of the upstream wave, enhanced flow towards (away from) the convection in the lower (upper) troposphere was noted. The lower tropospheric inflow was intensified by a nonnegligible amount. In the explicit moisture model, this increased the rate at which moist air from the upstream environment was being fed into the storm, at least during the organizational period.

The model storms in both frameworks eventually settled into extended mature phases possessing upshear tilted main updrafts modulated by multicellular behavior. In the more sophisticated model, however, the onset of maturity coincided with the development of a second and markedly different alteration of the storm’s upstream environment. This was also attributed to a propagating gravity wave, this time excited by a shallow, weak and yet persistent layer of midtropospheric cooling which appeared on the upstream side of the main storm updraft around this time.

This secondary wave propagated through the initial wave’s wake, substantially altering the upstream environment yet again. It was characterized by lower tropospheric
ascent which served to cool and moisten part of the storm inflow, forming the “cool/moist tongue”. Importantly, the second wave also caused the upstream inflow enhancement to shift from the surface to the middle troposphere. Although ubiquitous in two- and three-dimensional squall-line simulations made with full physics cloud models, this second gravity wave adjustment was completely missing from the original PM model runs.

Analysis suggested that two processes contribute to the establishment and maintenance of the cool/moist tongue: adiabatic expansion of rising subsaturated air and evaporation of cloud droplets detrained from the storm updraft. Neither of these processes were captured in the PM models of FT00 and GT, the former having been absent since all ascent in the unstable region was presumed already saturated. Introducing a minimum updraft speed for parameterized warming generation resulted in a similar tongue-like feature appearing in the PM model simulations as well. With this crude fix, the PM model storms possessed much more realistic upstream structures during the mature phase.

The magnitude of the cooling in the midtropospheric tongue is not large, especially when compared to the positive and negative temperature perturbations associated with the convective updraft and subcloud cold pool. Its impact on the storm inflow is impressive nonetheless. The cool/moist tongue’s associated upstream horizontal velocity perturbations affect the midlevel vertical wind shear in contact with the cold pool and (in the full physics model) encourages the entrainment of dry, stable midtropospheric air into the main storm updraft. Simultaneously, the tongue cooling and moistening established by the gravity wave response helps bring air above the mixed layer closer to saturation. The “daughter clouds” that can appear in this elevated layer were found to play a role in triggering new cell development by Fovell and Tan (1998).

Figure 14. As in Fig. 4 but for the modified PM simulation shown in Fig. 13b, for comparison with Fig. 10.
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References


