Tropical Oceanic Hot Towers: Need They Be Undilute to Transport Energy from the Boundary Layer to the Upper Troposphere Effectively? An Answer Based on Trajectory Analysis of a Simulation of a TOGA COARE Convective System

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ABSTRACT

This paper addresses questions resulting from the authors’ earlier simulation of the 9 February 1993 Tropical Ocean Global Atmosphere Coupled Ocean–Atmosphere Research Experiment (TOGA COARE) squall line, which used updraft trajectories to illustrate how updrafts deposit significant moist static energy (in terms of equivalent potential temperature $\theta_e$) in the upper troposphere, despite dilution and a $\theta_e$ minimum in the midtroposphere. The major conclusion drawn from this earlier work was that the “hot towers” that Riehl and Malkus showed as necessary to maintain the Hadley circulation need not be undilute. It was not possible, however, to document how the energy (or $\theta_e$) increased above the midtroposphere. To address this relevant scientific question, a high-resolution (300 m) simulation was carried out using a standard 3-ICE microphysics scheme (Lin–Farley–Orville).

Detailed along-trajectory information also allows more accurate examination of the forces affecting each parcel’s vertical velocity $W$, their displacement, and the processes impacting $\theta_e$, with focus on parcels reaching the upper troposphere. Below 1 km, pressure gradient acceleration forces parcels upward against negative buoyancy acceleration associated with the sum of (positive) virtual temperature excess and (negative) condensate loading. Above 1 km, the situation reverses, with the buoyancy (and thermal buoyancy) acceleration becoming positive and nearly balancing a negative pressure gradient acceleration, slightly larger in magnitude, leading to a $W$ minimum at midlevels. The $W$ maximum above 8 km and concomitant $\theta_e$ increase between 6 and 8 km are both due to release of latent heat resulting from the enthalpy of freezing of raindrops and riming onto graupel from 5 to 6.5 km and water vapor deposition onto small ice crystals and graupel pellets above that, between 7 and 10 km.

1. Introduction and motivations

In Fierro et al. (2009, hereafter F09), we used a simulation of the 9 February squall line observed during the Tropical Ocean Global Atmosphere Coupled Ocean–Atmosphere Research Experiment (TOGA COARE) to examine whether the updraft cores in a tropical mesoscale
convective system can act as “hot towers” to maintain the upward-moving equatorial branch of the Hadley circulation as described in the “hot tower hypothesis” (HTH) of Riehl and Malkus (1958) and Riehl and Simpson (1979). The HTH postulated that buoyant updrafts from a population of cumulonimbus clouds were able to carry moisture-rich boundary layer air up to the top of the troposphere with little or no dilution by mixing with the drier, lower energy air of the typical tropical environment, which is characterized by a distinct equivalent potential temperature ($\theta_e$) minimum in the midtroposphere generally 10–15 K lower than below cloud base (see Fig. 1c in F09 and later in the text). F09 found that while all parcels exceeding the 10-km level experienced a decrease in $\theta_e$ due to turbulent dilution below 4 km, they all exhibited a gradual increase in $\theta_e$ above that level. Hence, if only the beginning $\theta_e$ value in the boundary layer and the ending $\theta_e$ value in the outflow layer (i.e., above 10 km) were considered, they would give the impression of undilute, $\theta_e$-conserving ascent.

Subsequent research supports the essential aspects of the HTH but has questioned whether “hot towers” sufficient for tropical energy balance necessarily must be “undiluted.” Over a period of decades, numerous penetrations of tropical oceanic convective clouds with research aircraft showed that updrafts were surprisingly weak (LeMone and Zipser 1980; Lucas et al. 1994, and others). Further, a recent review from Zipser (2003) showed evidence that updrafts in marine tropical mesoscale convective systems were almost always diluted.

Observations in COARE, the Global Atmospheric Research Program (GARP) Atlantic Tropical Experiment (GATE), and other field experiments worldwide show that tropical oceanic updrafts often exhibit a vertical velocity (W) profile with a minimum near the freezing level (5 km; e.g., Zipser and LeMone 1980; Lucas et al. 1994; Igau et al. 1999). Above the freezing level, secondary and sometimes primary W maxima are suggested in the previous studies and documented in the statistics of Heymsfield et al. (2010) and the radar observations of specific mesoscale systems whose cloud tops penetrated into the upper troposphere with heights near or above 14 km (e.g., Hildebrand et al. 1996; Hildebrand 1998; Jorgensen et al. 1997; Roux 1998).

Recent modeling studies of this COARE squall line from Fierro et al. (2008, hereafter F08) and F09 suggested that the updraft minimum near the freezing level was due to water loading, with the updraft maxima at higher levels due to water unloading below the freezing level, followed by heating due to freezing and/or riming higher up (e.g., Saunders 1957; Iribarne and Godson 1981; Williams and Renno 1993). In a similar fashion, Zipser (2003) also hypothesized that release of latent heat of fusion and/or accretion of cloud water were important contributors to the acceleration of these updrafts above the freezing level.

The results of F09—that significantly diluted updrafts in an equatorial squall line could transport significant energy to the upper troposphere—offer a way to reconcile this body of work with the HTH. The trajectories that reached cloud top shared a common history: moist static energy ($\theta_e$) decreased from their points of origin to about 5 km (freezing level) because of dilution and then increased again because of freezing processes. But two questions remained unanswered. First, what freezing processes are involved in the $\theta_e$ increase; and second, what processes determine the parcels’ upward trajectories?

To address these questions, we performed more simulations of a typical tropical convective system sampled over the “warm pool” of the equatorial Pacific Ocean in 1993 during TOGA COARE. The squall line examined, which occurred on 9 February 1993, was classified by LeMone et al. (1998) as a “shear parallel convective band,” whose orientation and movement was related to the local vertical wind shear. A complete description of the COARE system evolution, data, instrumentation, flight coordination, and deployment strategies can be found in Petersen et al. (1999) and F09.

Here, we compute tendencies and accelerations along trajectories originating in the inflow air to account for the evolution of $\theta_e$ and W reported in F09. A brief description of the model used, settings, and methodology will be described in the next section. The modeling results will be presented in section 3, followed by the conclusions/discussion in section 4.

2. Model description and methodology

The model used in this study is the same as in F08 and F09, namely the Straka Atmospheric Model (SAM); Straka and Anderson 1993; Straka and Mansell 2005). The SAM is in essence a standard nonhydrostatic compressible cloud model in terms of numerics and cloud physics (F08). To offset the associated greater computational needs and the limitation of the SAM code structure used, which is set up to run only on a single processor [i.e., non–message passing interface (non-MPI)], we replaced the original single-moment 10-class bulk ice (10-ICE) scheme of F09 with the less computationally intensive single-moment 3-ICE microphysics scheme of Gilmore et al. (2004). The main difference between the 10-ICE and 3-ICE code is two more sizes of hail, two additional sizes of graupel, and two additional crystal habits. Since the heating source and sink terms for all the additional graupel (hail) and ice crystals are treated the same as in the 3-ICE
scheme, it is expected that the results presented herein would hold for the 10-ICE scheme. The five discrete bulk hydrometeor categories of the 3-ICE scheme also follow an inverse exponential size distribution for precipitating particles and a monodisperse size distribution for cloud particles. The five categories are cloud droplets, rain, cloud ice, snow, and graupel/hail as a single category. The model predicts the mixing ratio of each category, while total concentration is diagnosed from mixing ratio and assumed particle size. A constant cloud condensation nuclei (CCN) concentration of 100 cm$^{-3}$ was prescribed through the domain, which is a typical value for clean maritime air (e.g., Rogers and Yau 1989). Activation/nucleation of CCN is dependent on the local supersaturation following Soong and Ogura (1973). The reader is invited to refer to F08 and/or F09 for more details behind the SAM model numerics and physics.

The procedures follow and extend those in F09. Trajectories initially prescribed within a finite volume are computed using a forward-upstream technique at each time step. The trajectory data (Table 1) are saved at each time step for postprocessing, resulting in trajectories with the time resolution of the model computational time step (here 2 and 4 s for the 300- and 750-m simulations, respectively). This procedure guarantees simulated trajectories with high accuracy. As in F09, pseudoadiabatic equivalent potential temperature (labeled $\theta_e$ for simplicity) is computed using the iterative formula proposed by Emanuel (1994), which is more appropriate for the tropical environment:

$$\theta_e = \theta \exp \left[ L_v(T_{LCL}) q(T_{LCL}, P_{LCL}) \right] \frac{L_v(T_{LCL}) q(T_{LCL}, P_{LCL})}{c_p T_{LCL}},$$  \hspace{1cm} (1)

assuming

$$L_v(T_{LCL}) = L_{v0} + (c_{pv} - c_i) T (\degree C),$$

where $L_{v0} = 2.5 \times 10^6$ J kg$^{-1}$ and the specific heat capacity for water vapor and liquid water are respectively set to constant values, namely $c_{pv} = 1870$ J kg$^{-1}$ K$^{-1}$ for vapor and $c_i = 4186$ J kg$^{-1}$ K$^{-1}$ for liquid. Equation (1) is solved using Newton’s method, generally in 3–5 iterations with an accuracy on the order of $10^{-3}$ K. This value of $\theta_e$ is conserved for moist and dry processes (Emanuel 1994) under condensation and evaporation.

The acceleration terms were computed following Houze (1993) and Braun (2002), where total (or net) buoyancy acceleration $BA$ is defined as

$$BA = g \left[ \frac{\theta'}{\theta} + 0.608 q'_v + (\kappa - 1) \frac{p'}{p} - \sum_n q'_n \right].$$ \hspace{1cm} (2)

with $\kappa = R_v/c_v$, primes denoting perturbation quantities (from the homogeneous hydrostatic base state in overbars), $n$ denoting the particle type, and other symbols having their usual meanings. The terms are, from left to right, buoyancy changes owing to perturbation potential temperature, perturbation vapor mixing ratio, perturbation pressure, and hydrometeor loading. The buoyancy acceleration owing to loading by rain, for example, is just $-g q_v$. The perturbation pressure results from converting density to temperature. Its presence arises from the fact that the SAM model predicts pressure whereas other cloud models (e.g., Klemp and Wilhelmson 1978) use the Exner formulation for pressure. Typically, the $p'$ term is small compared to the potential temperature term (e.g., Schlesinger 1973).

The Lagrangian vertical acceleration $dw/dt$ for each parcel is then given by

$$\frac{dw}{dt} \approx BA + PGA + mixing,$$ \hspace{1cm} (3)

where $PGA = - (1/\rho_{air})(dp'/dz)$ is the vertical acceleration due to the vertical perturbation pressure gradient, $\rho_{air}$ being the density of air. Note that the second term in (3) is quite different from the third term in (2).

Heating and cooling associated with phase transitions from water to vapor (and vice versa) directly translate into changes in $\theta$ and do not affect $\theta_e$. Hence, in addition to
mixing effects, changes in $\theta_e$ are by definition attributed to heating/cooling from freezing, melting, deposition, and sublimation involving here the five particle types specified, with the last two adjusted to include only the impact of the “ice-to-water” or “water-to-ice” portion of the transition as shown below. Those four microphysics terms are estimated from the model output along each trajectory.

An evaluation of the “mixing” term in (3) from model output is difficult. For this reason, we employ the simplest approach whereby mixing is estimated by comparing side by side the simulated $\theta_e$ vertical profile and the vertical profiles of the aforementioned four microphysical terms.

More precisely, freezing and melting in the model takes into account the following processes. For rain, these are rain contact freezing and accretion (riming) upon contact with ice crystals, graupel, or snow and Bigg heterogeneous freezing. The cloud droplet freezing terms are riming/accretion onto ice crystals, graupel, and snow; conversion of cloud droplets to ice or snow through the ice crystal, graupel, and snow; and homogeneous freezing of cloud droplets at temperatures colder than $-40^\circ$C. For melting, only three terms are needed, accounting for melting of graupel, snow, and ice crystals.

Deposition (vapor to ice) and sublimation (ice to vapor) of crystals involve the computed particle mass changes with attendant latent heating changes in $\theta$ through the saturation adjustment scheme (Tao et al. 1989). Deposition includes diffusion of vapor onto graupel, ice crystals, or snow as well as nucleation of crystals, while sublimation includes transitions to vapor from ice in any form. Because deposition and sublimation involve latent heat/enthalpy of vaporization $L_v$, which only affects changes of $\theta$ and not $\theta_e$, the heating rate from those two processes were multiplied by the scaling factor $L_f/(L_v + L_f)$, where $L_f$ is the latent heat/enthalpy of fusion ($=336 \times 10^3$ J kg$^{-1}$). For the sake of simplicity, variations of $L_v$ and $L_f$ with temperature were not accounted for during the computation of the aforementioned four microphysics heating rates affecting $\theta_e$.

The squall line is initialized in the same manner as in F08 and F09 and therefore will be described only briefly. The initial thermodynamic environment and wind field are represented by a slightly modified version of the sounding launched from the R/V Vickers (Fig. 1a). This modified sounding exhibits attributes consistent with maritime tropical environments such as low lifting condensation level ($\sim600$ m), an equilibrium level near 14.5 km, weak convective inhibition, and a freezing level below 5 km AGL (i.e., 4.5 km). The reader is invited to consult F09 for a more detailed and comprehensive description of the sounding and corresponding wind profile.

The dimensions of the domain for all simulations are kept identical to that in F08 and F09: namely, with $(X, Y, Z) = (132, 210, 24)$ km. The convection is triggered via an ellipsoidal cold thermal anomaly with dimensions of $20 \text{ km} \times 130 \text{ km} \times 2 \text{ km}$ centered at $z = 1.5$ km, which was placed at $X = 105$ km and $Y = 54$ km, which mimics a gust front as in Trier et al. (1996, 1997).

To separate the impact of horizontal grid spacing from the impact of using the simpler cloud physics scheme, two high-resolution simulations are conducted: the first with grid spacing identical to F09 (horizontal 750-m, vertical 200-m) but with the simpler 3-ICE microphysics. The second simulation uses the same vertical grid spacing and simpler 3-ICE scheme but with a finer 300-m horizontal grid spacing. The nonacoustic computational time step is set to $dt = 4$ and 2 s for the 750- and 300-m runs, respectively. In both cases, the simulation is run for a total cloud time of 6 h.

The choice of 300 m was motivated by previous work of Bryan et al. (2003) and by observations, which show that the characteristic length scale of about 90% of the tropical maritime updraft and downdraft cores in GATE were smaller than 2 km (e.g., LeMone and Zipser 1980), and TOGA COARE (Igau et al. 1999), balanced against the increased cost of running on an even finer mesh. Fortunately, it is likely that many of the smaller cores never reached the upper troposphere. Choice of an optimal horizontal and/or vertical grid spacing for simulating precipitating convection remains an active area of research (e.g., Bryan and Morrison 2012). We suggest that if the general shape of $\theta_e$ along the trajectories reaching the upper troposphere changes little among simulations, we can use the 300-m run to explain them.

While most aspects of the treatment of the trajectories are identical for the three runs, their origins and numbers differ slightly. The time interval for the forward integration of the trajectories is the same as F09 for both runs: The trajectories, one per grid point, are inserted at hour 0 of cloud time at a distance of about 20 km ahead of the simulated gust front position and are ended at 4 h 50 min, at which time most parcels are either advected rearwards into the upper-tropospheric portion of the stratiform precipitation region and eventually into the anvil, or into the mesoscale downdraft region behind the leading edge of the convective line. A starting time of 2 h is chosen because the simulated squall line reaches maturity at about 1 h 30 min. The HTH assumes that parcels originate near or below cloud base; consequently, parcels were started at the lowest four model levels at $z = 100, 300, 500,$ and 700 m. The prescribed volume from which the parcels originate in the 750-m 3-ICE simulation is set to a slightly larger size than in F09 with dimensions $(X, Y, Z) = (12, 30, 0.6)$ km, which corresponds to $17 \times 41 \times 4 = 2788$
trajectories. For the fine-resolution 300-m 3-ICE run, the trajectories start from a volume with dimensions of \((X, Y, Z) = (3, 22.5, 0.6) \text{ km}\), corresponding to \(11 \times 76 \times 4 = 3344\) trajectories or about 20% more than in the 750-m 3-ICE run. In both simulations, the trajectories originate, as before, ahead of the convective line. For the 300-m run this corresponds to the grid points within the area delineated by \((X_1, X_2) \times (Y_1, Y_2) = (240, 250) \times (225, 300) = (72, 75) \times (67.5, 90) \text{ km}\) (highlighted in Fig. 2a by a black rectangle).

To maintain consistency with F09, the modeled trajectories in both runs are divided into the same batches according to their maximum altitude: batch A (0–6 km), B (6–10 km), C (10–14 km), and D (>14 km). Those threshold levels were selected for the following reasons: 6 km is well representative of the lowest level of the mixed-phase region in tropical convection (Petersen et al. 1999), 10 km is the mean height of the simulated higher updraft maximum, and 14 km is close to the cloud heights observed in COARE.

3. Model results

a. Comparison of the 300-m simulated squall-line evolution with F09

The two simulated squall lines follow an evolution similar to F09; hence, only the results from the 300-m simulation are described and compared here. At about \(t = 20–30 \text{ min}\), the line initiates from a rapid strong burst of isolated strong convective cells (maximum updraft nearing 25 m s\(^{-1}\)) along a line parallel to the long axis of the originating ellipsoidal cold anomaly, but normal to the
low-level shear (not shown). This short-lived burst is followed by its progressive collapse enhancing the low-level cold pool, which allows a squall line to form by 1 h 30 min (not shown). By $t = 2 \, \text{h} \, 15 \, \text{min}$ (Fig. 2a), 15 min after the parcels were released ahead of the gust front, the squall line is well developed. Cloud tops, represented here by the 5-dBZ echo contour, often reach an altitude of 14 km (Fig. 3a). Although hydrometeor mixing ratios were not measured in COARE, the simulated values are shown and briefly described here in order to facilitate the interpretations and subsequent analysis of the microphysical tendencies and accelerations shown later.

At $2 \, \text{h} \, 15 \, \text{min}$, the graupel mixing ratios often reach 5 g kg$^{-1}$ between $z = 5$ and 7 km AGL at the convective line’s leading edge, with the snow mixing ratio ranging between 0.5 and 1 g kg$^{-1}$ above 10 km AGL (Fig. 3d). An hour later, the squall-line convection still maintains its strength and exhibits a well-defined stratiform precipitation region behind its leading edge characterized by weaker up- and downdrafts. After about 4 h, the convection at the squall-line leading edge undergoes gradual weakening. At the same time, the simulated squall line’s stratiform region expands in areal coverage (Figs. 2c and 3c.f).

Comparing the simulated convective line with observations from Petersen et al. (1999), it is clear that the model produces a stronger-than-usual vertical development of the maritime convection. As stated in F08, this is likely attributed to the artificial and idealized triggering mechanism used to generate this squall line. The stratiform region is mainly composed of snow above 10 km AGL and a modest amount of graupel with mixing ratio rarely exceeding 2 g kg$^{-1}$ (Figs. 3b,c,e,f). As shown in F08, the model tends to overestimate the amount of graupel, particularly in the convective line with mixing ratios often near or exceeding 2 g kg$^{-1}$. While there were few microphysical measurements in COARE, more recent field campaigns in the 1990s that included aircraft penetrations of tropical clouds in the southwestern Amazon [the Tropical Rainfall Measuring Mission (TRMM)–Large-Scale Biosphere–Atmosphere Experiment in Amazonia (LBA)] and near Kwajalein [the Kwajalein Experiment (KWAJEX)] revealed that only updrafts greater than 5 m s$^{-1}$ contained appreciable amounts of supercooled water above the freezing level, which was mainly confined below the −12°C level (Stith et al. 2002). They found that above this level (~12°C) the great majority of the sampled clouds were glaciated with only traces of supercooled water existing.

**Fig. 2.** Horizontal ($X$–$Y$) cross section of simulated radar reflectivity (dBZ) at $z = 1$ km AGL for the 300-m run, at three different model times: (a) 2 h 15 min, (b) 3 h 20 min, and (c) 4 h 25 min, corresponding to the formation stage, mature stage, and onset of the decaying stage, respectively. Legends for color and shadings are shown on the right side of the figure. A thick black line shows the −1-K contour of potential temperature perturbation, which depicts the location of the outflow boundary at the surface. A small black rectangle highlights the location where the trajectories were released at 2 h.
up to $-18^\circ$C in a few of the cases. This finding would be consistent with limited riming confined within a relatively shallow layer within the mixed-phase region of tropical updrafts and hence small amounts of graupel. They also found that tropical anvils were mainly composed of ice crystals and aggregates, which the model seem to represent with a reasonable degree of realism with simulated snow mixing ratios ranging between 0.5 and 1 g kg$^{-1}$ (Figs. 3d–f).

The most relevant differences between the simulated squall lines are in their maximum $W$: the simulated maximum $W$ values at 750 m (20–25 m s$^{-1}$) are slightly stronger than at 300 m (15–20 m s$^{-1}$) and are, by virtue of the minimum allowed resolvable wavelength at a given horizontal grid spacing, wider by a factor of 2 or more (Weisman et al. 1997). Larger simulated maximum $W$ values within a continental squall line were also reported for grid spacings near 1 km compared to 250 m in a recent study from Bryan and Morrison (2012).

b. Comparison of the simulated squall-line trajectory statistics

Before carrying out the analysis along trajectories of the aforementioned four microphysics rates, (2), and (3) to determine the origins of the $\theta_e$ and $W$ tendencies, it is relevant to first compare the general trajectory statistics of F09 with the ones obtained in the above simulations. As noted in the foregoing, the trajectories are divided into four groups according to the maximum altitude reached, referred to as batches A ($\leq 6$ km), B (6–10 km), C (10–14 km), and D (>14 km). For brevity, we focus on the extreme batches A and D.

Comparing the present 750-m 3-ICE run to the 750-m 10-ICE run (F09) shows that the microphysics scheme affects the fraction of trajectories in each batch. In the 750-m 3-ICE simulation, 10.2%, 15.2%, 53.3%, and 21.3% of the 2788 parcels were distributed in batches A, B, C, and D, respectively, compared to 17.3%, 20.1%, 60%, and 2.7%
in F09. In F09, 62.7% of the parcels were able to exceed an altitude of 10 km compared to 74.6% here. The updrafts using 3-ICE are stronger than those for 10-ICE by about 5 m s\(^{-1}\), especially above the freezing level. Furthermore, the 3-ICE convective line at \( t = 2 \) h exhibits a more solid appearance than the 10-ICE run of F09.

This difference highlights the importance of the microphysics. The primary difference between the 3-ICE and 10-ICE scheme of the SAM model lies in the assumptions made regarding the ice-phase/rimed heavier precipitating particles that interact in a nonlinear fashion with the cloud-scale dynamics and kinematics. In the 3-ICE scheme, the graupel/hail category falls into one species with a single terminal fall speed, drag coefficient, density, and intercept parameter, which are here set to values representative of the small graupel expected in tropical oceanic convection due to the generally modest amounts of supercooled water observed above the freezing level (e.g., Stith et al. 2002). The graupel effective density was set to 400 kg m\(^{-3}\) and its intercept parameter to \( 4 \times 10^6 \) m\(^{-4}\). For rain (snow) the intercepts were set to \( 8 \times 10^6 \) m\(^{-4}\) (\( 3 \times 10^6 \) m\(^{-4}\)). This would produce particulate-loading histories different from the 10-ICE scheme, which allows a broader terminal fall speed spectrum associated with three densities of graupel and two sizes of hail.

Figure 4, which compares the statistics for the 750- and 300-m 3-ICE runs, also reveals interesting differences in trajectory statistics. For the 300-m run, 12.5%, 40.8%, 39.9%, and 6.8% of the 3344 parcels were distributed in batches A, B, C, and D, respectively, with “only” 46.7% of the parcels able to exceed a maximum altitude of 10 km during their journey. This 30% reduction in the percentage of parcels reaching higher levels is not surprising, considering that the 300-m run produced overall weaker updrafts and, most importantly, smaller and finer-scale updraft cores as also reported in a recent modeling work of a continental squall line by Bryan and Morrison (2012) (see section 3c).

Looking at the 300- and 750-m runs in more detail, the 750-m run tends to have more parcels originating from lower altitudes reaching a maximum altitude of 14 km and higher (batch D), in agreement with F09. However, this relationship does not hold for the 300-m run, which shows remarkably evenly distributed count statistics for batch D. Similarly, the 750-m run further shows that parcels reaching lower maximum altitudes (namely batches A and B) generally originate from the higher initiation levels (i.e., 500 and 700 m), as in F09, whereas the 300-m simulation shows no preference in altitude of origin for those batches.

c. Analysis of \( \theta_e \) along trajectories for both the 300- and 750-m simulation and comparisons with F09

Despite the differences in parcel numbers in each batch, the present simulations qualitatively reproduce the \( \theta_e \) behavior of interest in F09, namely that the simulated trajectories experience a decrease in \( \theta_e \) below 2–5 km (see Fig. 8 in F09), followed by an increase higher up. It was speculated in F09 that this evolution resulted from mixing of lower-\( \theta_e \) air dominating at lower levels and heating due to ice processes dominating higher up. These tendencies will be analyzed in more detail in this section.

Horizontal and vertical cross sections of trajectories of the two batches analyzed in this section, namely the extreme cases A and D, are shown for the 300-m run in Fig. 5. For the remainder of the analysis, only the parcels originating from an altitude of \( z = 100 \) m will be analyzed for the purpose of simplicity. The present results for parcels originating from \( z = 100 \) and 300 m in batches C and D are qualitatively similar to those in F09 (with the same applying to batches A and B together).

The trajectories in the figure are similar to those of F09 (see their Figs. 5 and 6, not shown for batch A). Both
groups of parcels move from their release points a few tens of kilometers ahead of the squall line, upward over the gust front, which is about 1 km deep at its leading edge (not shown). Those in batch D are later carried aloft in the strong convective updrafts at the leading edge of the line, with $W$ ranging from 3 to 10 m s$^{-1}$ especially below $z = 5$ km (Figs. 3d–f and 5b; see also Fig. 5a in F09). Upon reaching cloud top, near 14–15 km, those parcels are advected rearward into the upper-level stratiform anvil, which is mainly composed of ice crystals and snow particles (Figs. 3e,f; see also Fig. 5 in F09). The trajectories in batch A follow an evolution similar to those in F09, with a great majority of those parcels being advected back to lower levels once behind the leading edge of the gust front, which is consistent with the presence of downdrafts produced by the evaporative cooling and melting (not shown) of weak to moderate precipitation (Fig. 3).

Plots of $\theta_e$ as a function of time for batches A and D for the lowest level of origin (100 m), presented in Fig. 6, also show behavior broadly similar to F09. As they are being advected behind the leading edge of the line, the batch A parcels in the 300-m run experience an average decrease in $\theta_e$ of about 10 K (thick black line in Fig. 6) while moving downward (see later in the section) with $W$ between 0 and $-2$ m s$^{-1}$ (Fig. 3d–f). In stark contrast, $\theta_e$ decreases only by about 2 K for the batch D trajectories below the freezing level ($\sim$5 km AGL) and then increases higher up. The large decrease in $\theta_e$ for the batch D parcels during the earlier part of their ascent (from about 350 K to values as low as 340 K; Fig. 6a) was attributed to mixing with the much lower-$\theta_e$ air of the environment, which reached values as low as 336 K between 1.2 and 7.5 km (Fig. 1c in F09 and Fig. 8 herein). Later in their journey, most batch D trajectories completely recover from this $\theta_e$ decrease by showing $\theta_e$ values back near 350 K.

The consistency between the results of the 300-m 3-ICE simulation and the 750-m 10-ICE simulation in F09 provides confidence that the $\theta_e$ tendency and acceleration analysis along trajectories presented in the following two sections (sections 3d and 3e) is applicable to F09.
While the evolution of $u_e$ along the trajectories is similar for the 750-m run, the magnitudes of the changes differ slightly. The batch D trajectories end up with $u_e$ higher by about 2 K (Figs. 6a,b), with a smaller dip in $u_e$ at midlevels, consistent with a smaller passage time because of its larger $W$ (see Fig. 7a,b). The average magnitude of the $u_e$ decrease for batch A in the 750-m run exceeds that for the 300-m run by about 3 K.

Comparing the individual $u_e$ traces in Fig. 6 for parcels originating at 100 m to the average of the parcels originating at all levels in batches A and D (highlighted by the thick black line), it is confirmed that the overall $u_e$ evolution of all the parcels is similar. Hence we remain confident that selecting the parcels originating at $z = 100$ m is adequate for the present analysis.

The corresponding $W$ time series in Fig. 7 for both 3-ICE simulations also shows some degree of consistency with F09, namely $W$ in batch A is overall weaker than in batch D, with batch A parcels spending more time in downdrafts (compare Figs. 7a,b with Figs. 7c,d). However, there are several noteworthy differences between the 300- and 750-m 3-ICE simulations for batch D parcels. As mentioned earlier, $W$ often exceeds 20–25 m s$^{-1}$ for the 750-m run, compared to about 15–20 m s$^{-1}$ in the 300-m run and with less time variability (Figs. 7a,b). In the 750-m run, nearly all parcels reach cloud top in about 30 min. In contrast, the batch D parcels in the 300-m run take longer, slowing down or even descending briefly 1 or several times between about 7 and 8.5 km AGL before reaching their maximum altitude (see section 3d).

In particular, the leading edge of the simulated squall line using 300-m grid spacing is composed of several smaller updraft and downdraft entities (Figs. 3d–f) rather than of a more solid positive updraft contour as simulated more consistently in the 750-m run (not shown) in line with recent modeling work from Bryan and Morrison (2012). Hence, it is expected that most of the parcels reaching the highest levels (namely parcels in batches C and D) in the 300-m simulation will experience periods of strong decelerations resulting in near-zero or even negative $W$ during their ascent. An example of this behavior will be analyzed in detail in the
next section for one representative trajectory (highlighted by a thick black dashed line in Figs. 6 and 7 and a solid gray line in Fig. 5).

The behavior of batches B and C are intermediate as expected. Batch B updrafts are overall similar to those of batch A except for the addition of quick passages within the convective line (characterized by the strongest updrafts), as seen more consistently in batch C and especially in batch D. As in F09, the 300-m simulation showed a larger degree of dilution for parcels in batch C than in batch D (Fig. 8). In this figure, one can clearly see that if only the starting (near the surface) and ending $\theta_e$ values (i.e., in the upper troposphere for batch D) are considered, the impression of $\theta_e$ conservation, and hence “undilute ascent,” is given. Clearly the profile shows that these batch D parcels experience dilution at low levels between 500 m and 6 km AGL. The alert reader will also notice along-trajectory increases in $\theta_e$ even below the freezing level, something impossible if the parcel is mixing with the environmental values in the figure; rather, the increases should result from interactions with adjacent parcels within the cloud that have higher $\theta_e$ values.

d. Analysis of acceleration terms from total buoyancy equation along trajectories for the 300-m simulation

As shown in F08 and F09, the lower-level updraft maximum in the convective line is a very common feature. Citing the work of Trier et al. (1997), they associated the low-level updraft in their simulation with an upward perturbation pressure gradient force and speculated that the increase in $W$ higher up was a consequence of water unloading and buoyancy accelerations. To verify this speculation, it is logical to focus primarily on parcels reaching cloud top (i.e., near the top of the upward branch of the Hadley cell). Hence only results for parcels pertaining to batch D will be shown in the remainder of the paper, with particular emphasis on a representative batch D trajectory (trajectory number 36).

To look at the $W$ behavior in greater detail and for the sake of clarity, we plot the terms on the right-hand side of (3) in Fig. 9 for batch D trajectory 36. In the figure, total buoyancy acceleration was further separated into thermal buoyancy (TB) and hydrometeor loading following (2). For this trajectory, TB is the dominant positive
acceleration term from about \( z = 1 \) km all the way up to about 13 km near cloud top (Figs. 3a–c), with the largest values (near 0.1 m s\(^{-2}\)) in the 5–9-km layer (Fig. 9a). For reference, this value is roughly equivalent to a virtual-temperature excess of 3 K. Large positive TB is typical along the trajectories in batch D from \( z = 1 \) km upward, as are moderate to large \( W \) at times exceeding 15 m s\(^{-1}\) (Fig. 7), consistent with moderate convective available potential energy (CAPE, Fig. 1a). Note that in this simulation it took only about 30 min for the parcels in batch D to rise from about 1 to 10 km (Figs. 9a,b), meaning that, in spite of some peak \( W \) values of 15 m s\(^{-1}\), the average \( W \) during the ascent, significantly, is only 5 m s\(^{-2}\). Below 1 km, the only positive acceleration term is the pressure gradient (PGA), which is sufficiently large to overcome the negative BA and enable parcels in batch D to start rising in the convective line. Between 1 and 7 km, PGA is negative with magnitudes comparable to or even larger to that of BA, resulting in a net deceleration to near zero \( W \) between 5 and 7.5 km. For this parcel, the positive momentum gained below this layer allows a monotonic rise to 14 km. It is typical for batch D parcels to reach near zero, or even weak negative \( W \) values, near 7.5 km, before increasing their ascent rate back to 10 m s\(^{-1}\) or more in the second part of their ascent toward near cloud top.

To better understand this behavior, the locations of the parcel are plotted on vertical cross sections of \( W \) in Fig. 10. Figure 10b reveals that the deceleration of this parcel (and nearly all parcels in batch D) is associated with a downdraft near the 7.5–8-km level. Once through this region, the parcel accelerates upward to continue its ascent to near cloud top (Figs. 10c,d). Both Figs. 5b and 10 confirm that the parcels in batch D are very likely to have their ascent interrupted by brief periods of near-zero or even negative \( W \), in contrast to the 750-m simulation, which mainly showed updrafts through the whole depth of its leading convective line (see F09). As mentioned earlier, this difference in parcel behavior is attributed to the higher level of detail being resolved at a grid spacing of 300 m. In fact, Bryan et al. (2003) and Bryan and Morrison (2012) showed that squall lines exhibited rather noticeable differences in their convective structures when simulated with a horizontal grid spacing varying between 250 m and 1 km. Their 1-km simulation revealed a more solid leading convective line versus, at 250 m, a more “diffuse,” turbulent-looking (and more realistic) convective pattern as seen herein. While comparisons to observations have been encouraging for coarser models (e.g., Trier et al. 1997; F09), research continues on
determining the optimal resolution(s) for simulating mesoscale convective systems.

The vertical profile of TB and BA along this trajectory is similar to those for other batch D trajectories (not shown) and shows overall consistency with those in Fig. 13 of Trier et al. (1997). However, their simulated maximum TB and BA are about 25% smaller. Consistent with this, simulated $W$ values along their trajectories are also lower by about 5 m s$^{-1}$ (their Fig. 15).

Figure 11 shows the contribution of each of the predicted species to the “total loading” term in Fig. 9. Not surprisingly, the dominant loading term below about 5 km is from rain, with magnitudes reaching about 0.05 m s$^{-2}$ at about 4 km, or rainwater mixing ratio of 5 g kg$^{-1}$ (corresponding to an approximate rain rate of 100 mm h$^{-1}$), with a comparatively small but noticeable contribution from cloud water (generally less than 1 g kg$^{-1}$). Higher up, accelerations owing to rain loading decrease to zero, with ice particles becoming the dominant contribution. Consistent with Figs. 3d–f, accelerations owing to graupel loading are largest in the mixed-phase region between 5 and 7 km, with magnitudes reaching over 0.02 m s$^{-2}$ (corresponding to 2 g kg$^{-1}$ of graupel). Accelerations owing to loading of the sum of graupel, snow, and ice crystals show similar magnitudes between 7.5 km and cloud top, near 14 km.

Not surprisingly, deceleration owing to rain loading is by far the most prominent loading term, resulting from enhanced coalescence processes in tropical maritime clouds (e.g., Rogers and Yau 1989) operating on very moist air (dewpoint $= 25^\circ$C; Fig. 1a). To evaluate the influence of rainwater loading on $W$ above the freezing level, the above 300-m simulation was rerun with the loading acceleration term for rain alone removed. It showed simulated maximum $W$ more than twice as large as the present 300-m values with loading of rain active (often reaching 45–50 m s$^{-1}$ at $z = 8–12$ km; not shown), which is about 10 m s$^{-1}$ less than the maximum $W$ value predicted by parcel theory ($\sim 59$ m s$^{-1}$).

A surprising aspect of Fig. 11 (and Fig. 9) is that this parcel (and nearly all parcels in batch D) have rain mixing ratios of 1.5–2 g kg$^{-1}$ during the initial phase of their ascent at altitudes as low as at $z = 100$ m (indicated...
by rain loading accelerations from 0.015 to 0.02 m s$^{-2}$ below 500 m; Fig. 11a). In other words, the simulated parcels begin their ascent in the rain, just behind the leading edge of the gust front (Figs. 2 and 3). Conventional thinking would favor those parcels descending to the ground owing to rain loading. However, the combination of Figs. 9 and 12 reveals that once the parcels are behind the gust front (Figs. 12a,b; region of relatively lower $u_e$), the combination of large positive pressure perturbations near the ground (near 100 Pa; Figs. 12c,d) and negative pressure perturbations (as low as $-60$ to $-80$ Pa at 1.5 km) above it results in a large negative vertical pressure gradient and hence a large negative PGA in that layer (Fig. 9a), large enough to counteract that of the negative BA produced by rain loading and cooler air present.

Because Fig. 9 shows only one trajectory in batch D, we examined the acceleration profiles for each of the 200+ batch D trajectories, and most of them show similar behavior (not shown). Similar profiles for $W$, also not shown here, reveal that the levels with the most parcels experiencing weak to negative $W$ values are between 7.5 and 8.5 km, although about a dozen parcels decelerate at several different altitudes during their ascent. The interrupted ascent of the batch D parcels is illustrated well in Fig. 5b between $X = 50$ and $X = 70$ km.

e. Analysis of $\theta_e$ changes along trajectories for the 300-m simulation

Figure 13 shows vertical profiles and time series of $d\theta_e/dt$ resulting from freezing, sublimation, melting, and deposition, with appropriate corrections made for deposition and sublimation (see section 2) for the same representative trajectory as in Fig. 9. Heights corresponding to the $0^\circ$ and $-40^\circ$C isotherms are indicated in the figure for both typical batch D trajectories and the environment for trajectory 36. For reference, the freezing-level buoyancy in Fig. 9 corresponds to roughly a 3-K temperature excess or a freezing height about 500 m higher than the environment, while the buoyancy at 10 km corresponds to a roughly 2-K temperature excess and a $-40^\circ$C height of around 200 m above the environment of this parcel, assuming a moist-adiabatic lapse rate.

Before continuing the analysis, it is worth reminding the alert reader that in the figure “freezing” takes into account rain contact freezing and accretion (riming) upon contact with ice crystals, graupel, or snow; Bigg heterogeneous freezing; homogeneous freezing of cloud droplets at temperatures colder than $-40^\circ$C; riming/accretion onto ice crystals, graupel, and snow; and conversion of cloud droplets to ice or snow through the Werner–Bergeron–Findeisen process. The melting term includes melting of graupel, snow, and ice crystals. Deposition includes diffusion of vapor on graupel, ice crystals, or snow, as well as nucleation of crystals, while sublimation includes transitions to vapor from ice in any form.

From the figure, both freezing and adjusted deposition terms contribute significantly to the $\theta_e$ increase between 6 and 10 km AGL. From Fig. 9a, this is also the zone of updraft enhancement from release of latent heat of fusion. This effect has been recognized by many, including inter alia Braham (1952), Saunders (1957), Iribarne and Godson (1981), Williams and Renno (1993), Trier et al. (1997), Petersen et al. (1999), and Zipser (2003). Deposition is maximum between 7.5 and 10 km with heating rates near 20 K h$^{-1}$ while freezing is maximized in a shallow layer above the freezing level between 5.5 and 6 km with heating rates near 300 K h$^{-1}$ at 5.5 km (highlighted by a small arrow in Figs. 13a,b). This rapid jump in
freezing above 5 km is mainly explained by two known limitations of the 3-ICE (Lin–Farley–Orville) scheme: (i) freezing of raindrops and contact freezing is assumed to occur instantaneously when the ambient temperature is at and below freezing and (ii) there is significant immersion or Bigg freezing, the magnitude of which is assumed to increase exponentially with height.

Because this parcel (and most parcels in batch D) nearly stops climbing or even descends near 7.5 km, it experiences noticeable cooling from sublimation, with rates of about −15 K h⁻¹. Once the parcel reaccelerates and starts rising above that level, the heating rate of deposition is able to exceed that of the cooling of sublimation (Fig. 13a).

Not surprisingly, the enthalpy of deposition is maximized where moderate graupel mixing ratio (>2 g kg⁻¹) is present, which is between 6 and 10 km (Figs. 12c,d, 13a, and 14a). Freezing includes accretion/riming onto the three predicted ice species, heterogeneous freezing, and the Bergeron process, the last of which is maximized between 5 and 6 km because rain mixing ratios are still appreciable (Fig. 14a).

Above about 6–6.5 km the simulated clouds become quickly glaciated (Figs. 13a and 14a) in agreement with observations within tropical clouds (e.g., Petersen et al. 1999; Stith et al. 2002). Note that in batch D the heating rate from enthalpy of freezing goes to zero at about 7.5 km.

FIG. 12. As in Fig. 10, but for (a),(b) θ_e (K) in shaded contours and the 4 m s⁻¹ vertical velocity contour (dark gray lines) and (c),(d) perturbation pressure (Pa) in shaded contours and the 1 and 2 g kg⁻¹ graupel mixing ratio contours (dark gray lines). Legends for color and shadings are shown at the bottom of each corresponding row. Note that to better focus on the cold pool/gust front, the vertical axis only extends to a height of 12 km.
This is because cloud mixing ratio between 6.5 and 7.5 km does not decrease as abruptly as rain mixing ratio to near-zero values above the freezing level, resulting in cloud liquid water contents that are still sufficient for substantial freezing to occur, mainly in the form of riming of graupel (Figs. 13a and 14a). As before, vertical profiles of the above microphysical tendencies for each of the 200+ parcels of batch D were created (not shown) and compared side by side and confirmed that the above results for this single representative trajectory holds for nearly all the trajectories in batch D.

The results herein are in agreement with past modeling work from Tao et al. (2001) and Johnson et al. (2002), who showed that freezing/riming were key in the increase above about 6 km AGL for tropical marine convection. For continental convection, however, freezing is expected to play an even more dominant role, as the larger CCN concentration allows for more supercooled water to reach higher altitudes, spreading the release of heat from enthalpy of freezing over a deeper layer. In tropical maritime clouds, observations show that the lower CCN concentration results in more efficient coalescence beneath the freezing level preventing accumulation of condensate above, consistent with the simulated rain mixing ratio and freezing rate profile of this COARE simulation. In addition, it is the slower updrafts in tropical maritime clouds, comparable to the fall speed of raindrops in the 0° to −10°C layer (z = 5–6 km) that leads...
to their long residence time in this layer, permitting their freezing and the accompanying large fusion heating rate (Fig. 13).

4. Discussion and conclusions

The Riehl and Malkus (1958; see also Riehl and Simpson 1979) HTH postulated that deep undiluted convective clouds were transporting energy-rich boundary layer air up to the high troposphere, providing a significant contribution to the vertical (and hence latitudinal) heat transport within the equatorial upward branch of the Hadley cell. While the first study (F09) confirmed that buoyant tropical oceanic clouds participating in the vertical transport of high-$\theta_v$ boundary layer air to the outflow layer in the upper troposphere experience some degree of dilution below the freezing level from entrainment and mixing, a number of relevant scientific questions remained unanswered. Among those questions was which microphysical processes were responsible for the simulated $\theta_v$ increase above the freezing level.

Toward this goal, we conducted an idealized simulation of the same well-documented typical tropical maritime convective system from the TOGA COARE field program (9 February 1993) as in F09, but using a finer horizontal grid spacing to resolve the well-documented smaller scale of tropical oceanic updrafts. A simpler microphysical scheme was employed to enable calculation of microphysical tendencies and accelerations along trajectories at a temporal resolution of the model advective time step given the single-thread code structure of the version of the SAM model used here.

Results show multiple trajectories with an increase in $\theta_v$ aloft, often associated with a second updraft maximum above 6–8 km. The $\theta_v$ increase can be mainly attributed to heating from enthalpy of freezing between 5 and 6.5 km and enthalpy of deposition between 7 and 10 km, while the updraft maximum above 6–8 km resulted primarily from thermal buoyancy excess.

It is also confirmed that hydrometeor loading has a significant impact on the updraft acceleration, especially below the freezing level where rain mixing ratios are appreciable because of enhanced coalescence processes. In fact, removing the effect of rain loading alone in the vertical momentum equation resulted in an updraft speed above the freezing level more than twice as large as presently simulated.

An interesting result of this simulation was the finding of the relatively large contribution of the PGA above the freezing level, which nearly balanced the positive thermal buoyancy acceleration. Williams and Stanfill (2002) suggested that because the observed updraft speeds above the freezing level in tropical buoyant clouds were much less than what was predicted by standard parcel theory (in this simulation $\sim 15$ vs $\sim 60$ m s$^{-1}$), the difference is likely due to dilution by turbulent entrainment of lower ambient $\theta_v$ air. Among many others, Xu and Emanuel (1989) argued that the dominant balance above the freezing level in tropical clouds was between thermal buoyancy and the gravitational loading of condensate, in which case the CAPE largely vanishes in reversible ascent, thereby erasing conditional instability in the tropical atmosphere. However, other authors (e.g., Trier et al. 1997) argued that that pressure fields in mesoscale convective systems affect updraft vertical velocities through their depths. Our results are consistent with those of Trier et al. with the dominant balance (aloft) between thermal buoyancy and the PGA.

The large negative PGA above the freezing level owes its existence to a persistent large negative pressure anomaly (on the order of 1 hPa) near the 4.5–5-km level (Figs. 12c,d). Looking at Figs. 13a and 3d–f, this mesoscale (i.e., diameter on the order of tens of kilometers) low is largely attributed to the hydrostatic effect of warming, in turn due to fusion heating during freezing of raindrops and accretion/riming of supercooled water onto frozen drops (consistent with graupel mixing ratios nearing 2 g kg$^{-1}$). This idea is not new and has been found to be quite important in the maintenance of mesoscale convective systems (e.g., Maddox 1980).

Now that we have provided strong evidence that convective cores in tropical mesoscale systems can transport the needed energy to the top of the troposphere, the more difficult question remains as to whether they actually do. This would require an updated version of the earlier Riehl and Malkus (1958) and Riehl and Simpson (1979) studies.

A second challenge for forthcoming studies would be to determine how the results obtained here for this maritime tropical squall line would change for a typical continental squall line in the midlatitudes.

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Special Note

Dr. Joanne Simpson strongly encouraged the authors to follow up F09 by a new set of model runs, using trajectory calculations to learn which microphysical terms restored $\theta_e$ to the high values found in the boundary layer. She made many specific suggestions for the conduct of this research during the last years of her life, motivated to understand just how real clouds may differ from the undilute ascent that her previous work indicated. While she could not review the final product, we believe that it is appropriate to recognize her contributions with coauthorship.

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