

RESEARCH ARTICLE

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Special Section:

Fast Physics in Climate Models: Parameterization, Evaluation and Observation

Key Points:

- New ARM scanning radars can highlight differences in numerical simulations
- Simulations constrained to observations captured key aspects of the cloud event
- More complete observations are required to constrain 3-D cloud behavior

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Insights from modeling and observational evaluation of a precipitating continental cumulus event observed during the MC3E field campaign

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Abstract A case of shallow cumulus and precipitating cumulus congestus sampled at the Atmospheric Radiation Measurement Program Southern Great Plains supersite during the Midlatitude Continental Convective Clouds Experiment is analyzed using a multisensor observational approach and numerical simulation. Observations from a new radar suite surrounding the facility are used to characterize the evolving statistical behavior of the precipitating cloud system. This is accomplished using distributions of different measures of cloud geometry and precipitation properties. Large-eddy simulation (LES) with size-resolved (bin) microphysics is employed to determine the forcings most important in producing the salient aspects of the cloud system captured in the radar observations. Our emphasis is on assessing the importance of time-varying versus steady state large-scale forcing on the model's ability to reproduce the evolutionary behavior of the cloud system. Additional consideration is given to how the characteristic spatial scale and homogeneity of the forcing imposed on the simulation influences the evolution of cloud system properties. Results indicate that several new scanning radar estimates such as distributions of cloud top are useful to differentiate the value of time-varying (or at least temporally well-matched) forcing on LES solution fidelity.

1. Introduction

Much of the uncertainty in climate model projections stems from our limited understanding of cloud and precipitation processes and the parameterization of these processes in global climate models (GCMs). While representing deep convective clouds in GCMs remains an ongoing problem [Stevens and Bony, 2013], the treatment of shallow clouds such as stratocumulus and trade cumulus also plays a crucial role in climate system modeling. Because of their high albedo, broad areal coverage, and high radiating temperature (i.e., low, warm cloud tops), stratocumulus produce a climatologically important cooling effect on the climate system [Hartmann *et al.*, 1992; Klein and Hartmann, 1993]. The moistening and cooling from shallow cumulus balance the warming and drying from large-scale subsidence. Turbulent fluxes from shallow cumulus clouds thus help determine the steady state depth and thermodynamic properties of the trade wind boundary layer [Riehl *et al.*, 1951; Stevens, 2007; Rauber *et al.*, 2007] and perhaps precondition the lower atmosphere for deeper convective events [e.g., Johnson *et al.*, 1999; Waite and Khouider, 2010].

Representing shallow clouds and their transitions in GCMs is a major challenge [Bony and Dufresne, 2005; Wyant *et al.*, 2007; Medeiros *et al.*, 2008]. Typically, results from cloud-resolving models (CRMs) or large-eddy simulation (LES) models serve as benchmarks for developing and tuning single-column models (SCMs). These SCMs in turn often form the basis of parameterizations in GCMs. Thus, LES is positioned as a key tool for investigating cloud system processes (e.g., aerosol indirect effects). For example, LES intercomparisons under the GCSS/GASS (GEWEX (Global Energy and Water Exchanges Project) Cloud Systems Study/Global Atmospheric System Studies) framework have been fruitful [e.g., Moeng *et al.*, 1996; Bretherton *et al.*, 1999; Stevens *et al.*, 2001; Brown *et al.*, 2002; Siebesma *et al.*, 2003; Stevens *et al.*, 2005; Ackerman *et al.*, 2009; van Zanten *et al.*, 2011], and simulations from these intercomparisons have served as a benchmark for SCMs and associated parameterization development [e.g., Lenderink *et al.*, 2004; Zhu *et al.*, 2005; Wyant *et al.*, 2007]. However, this GCSS/GASS paradigm investigates model pathologies largely using intercomparisons of simulations, placing less emphasis on evaluating LES solutions against observational constraints. One noteworthy exception to this approach was the

Dynamics and Chemistry of Marine Stratocumulus [Stevens *et al.*, 2003] model intercomparison [Stevens *et al.*, 2005], in which model representations of entrainment rate were compared to innovative new aircraft observations of cloud top entrainment [Faloona *et al.*, 2005].

One explanation for a deemphasis on observations is that shallow clouds are difficult to reliably sample, and our existing observational capabilities are often incompatible with high-resolution LES models. Although shallow cloud observations are collected over a handful of ground-based supersites [Stokes and Schwartz, 1994; Ackerman and Stokes, 2003; Illingworth *et al.*, 2007; Mather and Voyles, 2013], comprehensively sampling elusive targets such as shallow cumulus requires merging a wide array of active and passive cloud sensors, together with support from aircraft [e.g., Vogelmann *et al.*, 2012]. Newer approaches employing Doppler radar spectra can give dynamical and microphysical insights into shallow clouds, but until recently these techniques have been computationally prohibitive [e.g., Kollias *et al.*, 2011; Luke and Kollias, 2013]. Adding to these challenges, the majority of high-resolution data sets have been restricted to single-column or “pencil beam” measurements from millimeter-wavelength radars that must represent transient cloud and precipitation behaviors across scales from typical LES/CRM domain sizes ($\sim 4\text{--}40\text{ km}^2$) to GCM (grid box) scales ($\sim 100\text{ km}^2$).

This study documents our preliminary effort toward a more natural coupling of LES modeling and cloud-observing capabilities for one particular ground-based supersite. This site is the Oklahoma Southern Great Plains (SGP) location operated by the Department of Energy (DOE) Atmospheric Radiation Measurement (ARM) program centered at Lamont, Oklahoma. The site recently complemented its single-column measurements with new scanning radar that may better suit the sampling of LES-scale domains. We consider an application of LES with size-resolved (bin) microphysics as one avenue for capitalizing on the fine details sampled by these new radar facilities. Observations from radar systems should compare favorably with bin microphysical models that are better capable of reproducing radar measurements from drop-size distributions and high-resolution turbulent flow fields [e.g., Kollias *et al.*, 2011]. An overarching objective is to determine whether scanning radars provide useful model constraints for documenting the time-evolving structure of clouds, precipitation, and associated processes around this ARM site [Mather and Voyles, 2013; Kollias *et al.*, 2014]. An additional emphasis is to evaluate the sensitivity of cloud and precipitation properties to differences in the spatial scale and temporal details of the large-scale forcing.

We investigate the sensitivity of LES solutions to changes in spatial and temporal large-scale forcing for a challenging shallow cumulus and congestus event. We emphasize the transient nature of the forcing (i.e., time-varying versus steady state) and its importance on simulation outcome. This is accomplished by assessing the extent to which ensemble members reproduce salient observed macrophysical cloud field properties over the ARM facility [e.g., Xu *et al.*, 2002, 2005; Xie *et al.*, 2005; Mechem *et al.*, 2010]. The counterargument that a model must first reproduce microphysical processes in order to faithfully represent macroscale properties is also valid. Although evaluating LES performance against macroscale cloud properties is often the easier task and the approach this study adopts, improvements to LES macroscale and microscale cloud property representations are intrinsically iterative.

The paper is outlined as follows. A synoptic overview of the event and the associated model forcing considerations is presented in section 2. The event is analyzed using a multisensor observational approach and numerical simulation as described in sections 3 and 4. A discussion on the sensitivity of LES macroscale properties as evaluated against ARM observations is provided in section 5. Key findings are summarized in section 6.

2. The 25 May 2011 Congestus Event: Synoptic Overview and Cloud Forcing

This study focuses on a strongly forced case of shallow cumulus transitioning to precipitating cumulus congestus. The clouds were observed on 25 May 2011 during the DOE ARM and National Aeronautics and Space Administration (NASA) Global Precipitation Measurement mission campaign, the Midlatitude Continental Convective Clouds Experiment (MC3E) [Jensen *et al.*, 2010]. Clouds developed following a frontal passage and a period of deeper convection connected with a mature low-pressure system over central Kansas (Figure 1). Shallow clouds formed over the SGP site by 1500 UTC (0900 CST). The clouds surrounding the SGP facility gradually transitioned to scattered cumulus congestus by midday (1800 UTC; 1200 CST) and isolated, comingled, precipitating congestus thereafter. Winds at cloud level naturally shifted from southwest to northwest as the low-pressure system progressed eastward by the afternoon (late UTC) hours.

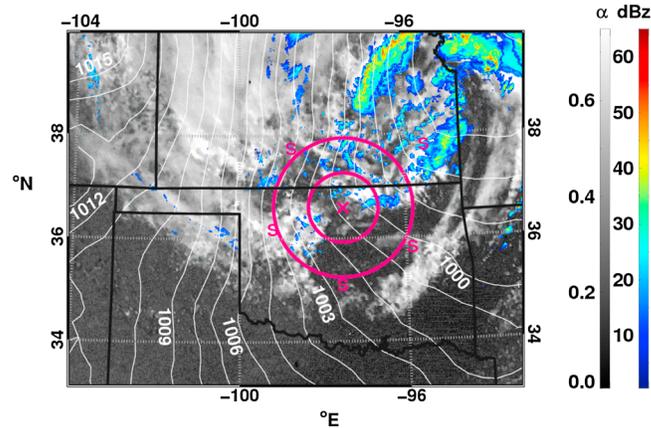


Figure 1. Synoptic configuration at 1800 UTC. Albedo from GOES visible satellite imagery is combined with isobars of the surface pressure field (in hPa) from North American Regional Reanalysis [Mesinger et al., 2006] and a multiradar composite of reflectivity from WSR-88D Next Generation Weather Radar (NEXRAD). The magenta cross represents the location of the ARM Central Facility, and the multiple S's indicate MC3E sounding sites. Concentric circles represent the analysis volumes of the 75 km and 150 km variation analysis domains.

An idealized perspective on the evolution of the 25 May 2011 synoptic conditions can be constructed from MC3E measurements using the variational analysis method of Zhang and Lin [1997], as also in Xie et al. [2014]. These analysis fields serve as the forcing data set for the simulations presented in this study (e.g., Figure 2 and Table 1). In addition to these multiscale forcing data sets, the MC3E campaign launched radiosondes at 3 h intervals from six locations within a 150 km radius of the SGP Lamont central facility [Jensen et al., 2015] (locations highlighted in Figure 1). Figure 2 shows time-height representations of several fields from the variational analysis product. This figure includes potential temperature and moisture advection, as well as vertical motion, at two characteristic

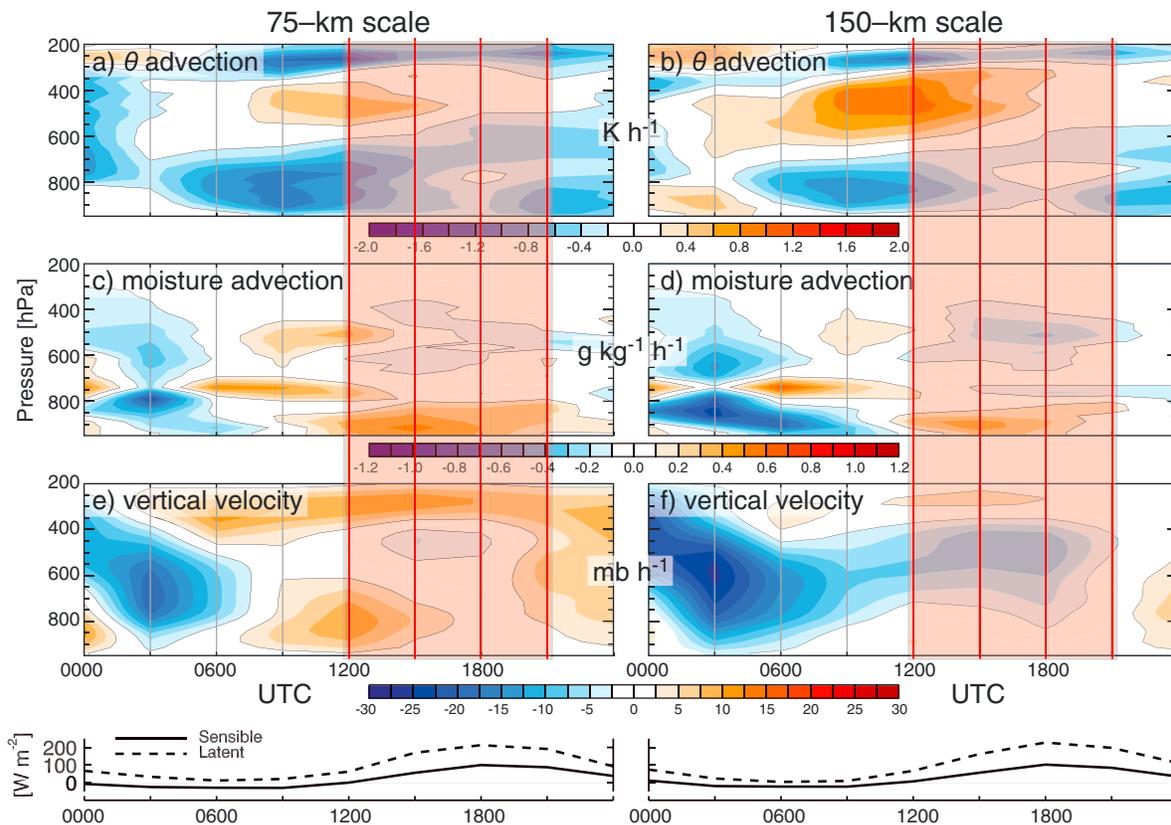


Figure 2. Time-height sections of potential temperature and moisture advection, large-scale vertical velocity, and surface-sensible and latent heat fluxes at two different analysis scales. The SAMEX forcing for the different simulations is constructed from different combinations of vertical profiles, denoted by the red vertical lines inside the rose boxes. The rose shading denotes the time period for which simulations are performed.

Table 1. Description of Simulation Suite^a

Simulation Name	Forcing Scale (km)	Time-(V)arying/ (S)teady State	Time or Interval of Forcing (UTC)	Potential Temperature Advection (K h^{-1})		Moisture Advection ($\text{g kg}^{-1} \text{h}^{-1}$)		Vertical Motion (mb h^{-1})	
1200v_75	75	V	1200–2100	−0.40	−0.30	0.17	0.13	1.33	1.31
1200v_150	150	V	1200–2100	−0.17	−0.25	0.10	0.02	−1.68	−0.01
1500s_75	75	S	1500	−0.47		0.19		2.15	
1500s_150	150	S	1500	−0.24		0.13		−0.96	
1800s_75	75	S	1800	−0.32		0.15		0.52	
1800s_150	150	S	1800	−0.10		0.07		−2.40	
2100s_75	75	S	2100	−0.27		0.10		2.10	

^aThe mean forcing terms for the two time-varying simulations represent density-weighted mean values from the surface to 640 hPa over the early (3–6 h) and late periods (6–9 h) of the simulations.

spatial scales (75 km and 150 km). These characteristic forcing domains are centered on the SGP central facility in Lamont, Oklahoma (Figure 1).

Using the forcing fields in Figure 2 as a guide, the gradual deepening of the clouds from shallow cumulus to congestus can be attributed to several factors. The northern Oklahoma region experienced positive low-level moisture advection (from the surface up to ~800 hPa) throughout the event. This moist advection included moisture advected from the Gulf of Mexico wrapping around the mature low-pressure system. Throughout the day, a deep layer of cold advection (from the surface up to a level of 650–500 hPa aloft, depending upon the time and analysis scale) brought the layer closer to saturation and likely promoted cloud development. Upward vertical motion overlaid low-level subsidence (or very weak vertical motion), although the relative strength of these features depended greatly on the scale of the forcing.

3. ARM SGP Site Observations and Cloud Macroscale Properties

The cloud macroscale properties selected for this 25 May 2011 LES sensitivity study rely on the radar observations now routine for the ARM SGP. For simplicity, we limit this discussion to four observational quantities when evaluating the LES ensemble solutions. Our first consideration when selecting these products is to target cloud properties that are simple to estimate from radar, focusing in particular on fields less susceptible to known radar challenges such as miscalibration or attenuation in cloud/rain [e.g., Ryzhkov *et al.*, 2005; Kollias *et al.*, 2014]. Our efforts emphasize macroscale cloud properties that follow naturally from LES model outputs without the need for advanced forward-radar modeling or computations.

Historically, the ARM radar data stream most referenced for the evaluation of LES cloud simulations is the ARSCL product (Active Remote Sensing of CLOUDs) [Clothiaux *et al.*, 2000; Kollias *et al.*, 2005]. This radar product merges observations from the vertically pointing 35 GHz Ka-band radar (KAZR), collocated laser ceilometer, micropulse lidar, and microwave radiometer to identify cloud boundaries in the vertical profile. New scanning radar capabilities complement those single-column cloud insights in several ways. Precipitating cell properties are now available to within 150 km of the SGP Lamont site using the surveillance 5.4 GHz ARM C-band Scanning ARM Polarimetric weather Radar (C-SAPR) and the nearby Vance AFB, Oklahoma (KVNIX) Weather Surveillance Radar, 1988 Doppler (WSR-88D) [e.g., Giangrande *et al.*, 2014]. MC3E also unveiled the first generation of ARM dual-frequency 35/94 GHz (Ka/W-band) Scanning ARM Cloud Radars (SACR) [e.g., Kollias *et al.*, 2014]. During MC3E, the SACR system sampled cloud elements using 2-D slices (range-height; RHI) to quickly track cloud features in time and evaluate their geometric properties [e.g., Borque *et al.*, 2014]. The four estimates of cloud system properties as obtained from these facilities are briefly discussed below.

3.1. Cloud Fraction

Cloud fraction (CF) is widely considered for model performance evaluation [e.g., Bedacht *et al.*, 2007; Wilkinson *et al.*, 2008; Walsh *et al.*, 2009]. Interpreting observational CF estimates that best align with model CF definitions is nontrivial, since current observations are unable to unambiguously reconstruct an instantaneous 3-D cloud field [e.g., Wiscombe, 2005; Wu *et al.*, 2014]. This study considers five observational definitions for CF that range from 1-D, column-based radar approximations to satellite-based horizontal integrations over spatial footprints comparable to LES domain sizes. Our analysis includes the following estimates of CF:

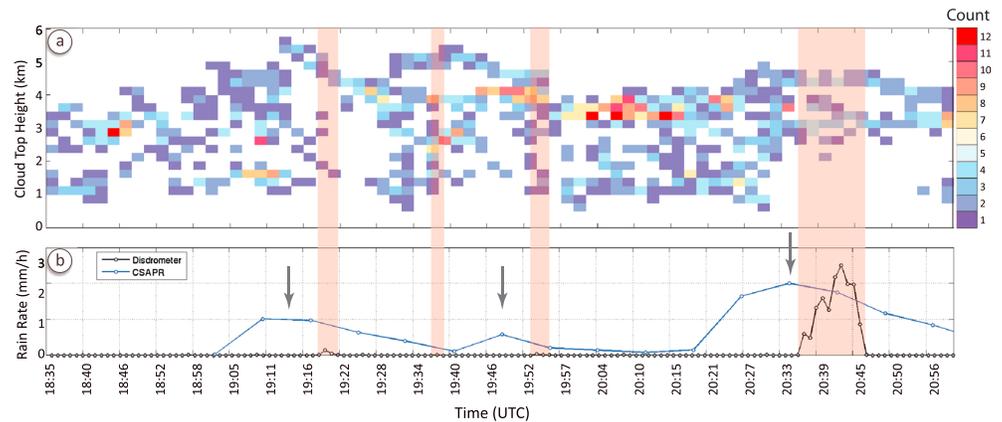


Figure 3. (a) Histogram of cloud top heights from the SACR as a function of time. (b) Rain rate from the disdrometer at the SGP central facility and an averaged radar-derived estimate from the C-SAPR using a standard Oklahoma Z-R relationship along the SACR scanning path. Shaded regions indicate the presence of precipitation as recorded by the disdrometer, and arrows in Figure 3b indicate periods of heavier area-mean precipitation measured by the C-SAPR. Adapted from *Borque et al.* [2014].

1. The 2-D VISST (Visible Infrared Solar-Infrared Split Window Technique) height-integrated GOES geostationary satellite estimate. This technique estimates the “cloudy pixel” area relative to the $0.5 \times 0.5^\circ$ latitude-longitude grid box that overlaps with the ARM SGP [Minnis *et al.*, 2011].
2. The CF estimates from 2-D hemispheric projections obtained using an ARM Total Sky Imager (TSI) camera [Kassianov *et al.*, 2005; Long *et al.*, 2006].
3. An estimate using shortwave radiation measurements from the ARM Solar Infrared Radiation Station (SIRS) data [Xie and Liu, 2013].
4. From the ARM radars, the standard ARM vertically pointing ARSCL definition that defines CF as the fraction of an hour that the radar/lidar column designates cloud echo overhead.
5. A new CF approach, developed from the 25 May 2011 MC3E radar data set [Borque *et al.*, 2014] that uses SACR “along-wind” RHI modes. This CF estimate is defined by projecting the horizontal area covered by cloud elements sampled along the 40 km RHI slice. Since the scanning orientation follows the prevailing wind, these estimates assume that cloud fraction is homogeneous over the domain in the crosswind direction.

3.2. Cloud Top Distributions

Several studies of cloud populations in the tropical environment have documented cloud tops that coincide with the trade wind inversion or the weakly stable layer often present at the melting level [e.g., Johnson *et al.*, 1999; Jensen and Del Genio, 2006; Mechem and Oberthaler, 2013]. Spaceborne systems can help document cloud top distributions and evolution, especially when coupled with temperature profile information [e.g., Mace *et al.*, 2009; Luo *et al.*, 2010]. However, spaceborne and surface weather radar methods are most reliable for documenting the cloud life cycle of cumulus congestus and deep convection. For shallow cumulus and congestus properties, the high sensitivity and rapid scanning ability of the SACR promoted sampling of the time-evolving distribution of cloud tops during the 25 May 2011 event.

Figure 3a (adapted from Borque *et al.* [2014]) documents transitions between periods of solely shallow cumulus, periods of commingled shallow and congestus clouds, and periods of primarily congestus clouds. The modes are reflected in the cloud top height distribution, where we see a narrow upper level distribution mode between about 4 and 5 km, representing primarily cumulus congestus periods, and a broader distribution representing shallow cumulus and commingled times. Please note that although the upper portions of the congestus clouds are colder than their tropical counterparts [Mechem and Oberthaler, 2013] and may reach temperatures down to -10°C , for our simulation purposes we assume that they are largely governed by warm-rain microphysical processes (an assumption we address in section 4.3). Precipitation onset and the subsequent generation of cold pools, in this case, are reflected in the cloud top distribution. We confirm the former by consulting the corresponding rainfall rate time series based on nearby ARM systems (Figure 3b). These rainfall rate traces in Figure 3b include measurements from the surface disdrometer in

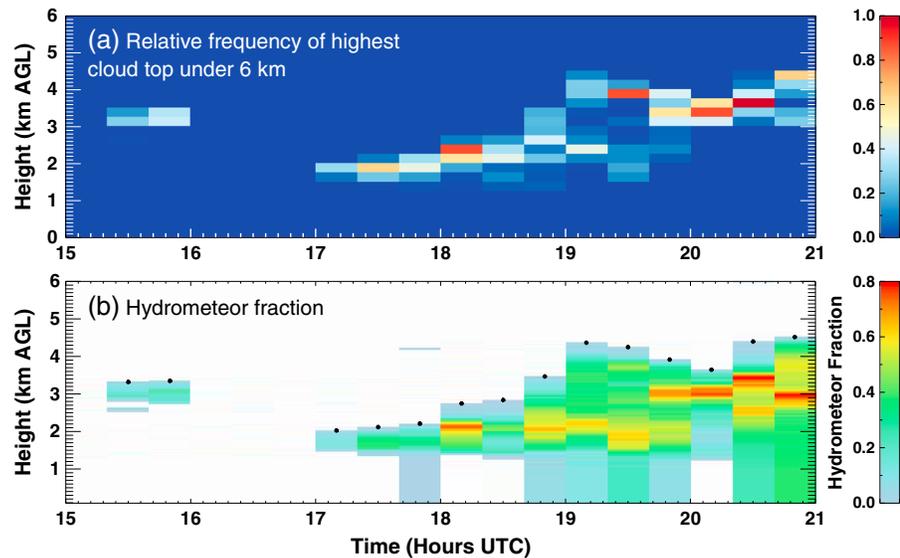


Figure 4. (a) Frequency of occurrence of highest cloud top under 6 km, calculated from KAZR-ARSCL cloud top product and binned in intervals of 20 min in time and 250 m in the vertical. (b) Time-height section derived from KAZR-ARSCL of hydrometeor fraction, averaged over 20 min intervals, overlaid with cloud top (black dots).

close proximity to the SACR and an averaged C-SAPR estimate collected along the SACR scanning path. The presence of precipitating cumulus is confirmed by considering the enhancements in higher cloud top height counts at the expense of lower topped clouds. These time-height distributions complement traditional CF estimates less capable of differentiating multilayer or commingled cloud types. However, these observations from the 25 May 2011 SACR data set are somewhat noisy and reflect some limits of the current ARM single-radar sampling strategies (only approximately 2.5 h of data from the date under investigation). Thus, a combination of multiplatform observations (e.g., ARSCL, TSI, SIRS, GOES), for cross validation, provides additional support for interpretations of the limited SACR observations.

Despite the single-column limitations mentioned above, the KAZR-ARSCL product provides an additional perspective on cloud system behavior over the period (Figure 4). From 1500 to 1700 UTC, the KAZR does not capture the isolated shallow cumulus visible in the TSI (not shown). However, starting at 1700 UTC, the radar does sample a cloud layer that deepens and increases in cloud fraction over the next 2 h. Designated cloud top heights then undulate by ~ 1 km in height from 1900 to 2100 UTC. KAZR-ARSCL cloud tops and hydrometeor fraction (Figures 4a and 4b) hint at the bimodal (multilayer) cloud structure suggested by the SACR statistics in Figure 3. Some differences in cloud statistics in Figures 3 and 4 during overlapping sequences may be partially attributable to fundamental differences in scanning strategies and capture volumes between the Ka-SACR (RHI) and the KAZR (profiling).

3.3. Areal Precipitation Fraction and Totals

A common weather radar application is to document surface rainfall accumulation and precipitating cell properties [e.g., Doviak and Zrnica, 1993]. For climate model applications, this may include properties such as domain-mean precipitation rate, accumulation, and time-varying behaviors. The fractional area of a representative domain covered by precipitation is one such property and is analogous to spatially averaged CF estimates. For this study, precipitation area fraction can be estimated from the C-SAPR data set by calculating the area covered by near-surface radar reflectivity factor Z exceeding a minimum threshold. The selection of a 30 dBZ threshold is based on previous warm-season precipitation studies in the vicinity of the ARM SGP site [e.g., Giangrande et al., 2012, 2014]. To allow for a range of radar measurement uncertainty in light rain conditions, this study considers a spread of area fraction corresponding to a less restrictive 25 dBZ threshold and a more restrictive 30 dBZ threshold.

For domain-mean precipitation accumulations, the simulations are forced with horizontally homogeneous profiles representative of 75 km and 150 km diameter circles centered on SGP Lamont. Radar rainfall products are available at high resolution and may be resampled to those domains. The variational approach that

generated the forcing data sets was constrained using surface precipitation estimates from the Arkansas River Basin Forecast Center (ARBFC) 4 × 4 km hourly WSR-88D radar-surface gauge hybrid rainfall product [e.g., Young *et al.*, 2000]. The forcing from the variational analysis is therefore constrained by these mean rainfall estimate products, which should then be reflected in model behavior. However, the influence of this precipitation constraint on the forcing (mediated through diabatic heating and drying of the column) should be weak based on the small value of domain-mean precipitation relative to cases involving substantial precipitation [e.g., Xie *et al.*, 2014]. We compare LES domain-mean rainfall properties to those from an unattenuated standard gauge-adjusted WSR-88D radar rainfall product, the National Mosaic and Multi-Sensor Quantitative Precipitation Estimation (NMQ) product [e.g., Zhang *et al.*, 2005]. For this event, the accumulations were nearly identical to those of the ingested ARBFC products, providing additional confidence in the quality of these areal precipitation estimates.

4. Large-Eddy Simulation

4.1. Model Configuration and Microphysics

All simulations here use the System for Atmospheric Modeling—Explicit Microphysics (SAMEX). SAMEX employs the anelastic dynamical core from the System for Atmospheric Modeling (SAM) [Khairoutdinov and Randall, 2003] and has been successfully used in a number of numerical modeling efforts [van Zanten *et al.*, 2011; Kogan *et al.*, 2012; Mechem *et al.*, 2012]. In this study, we configure SAMEX in a manner roughly similar to that of the RICO (Rain in Cumulus over the Ocean) [Rauber *et al.*, 2007] trade cumulus intercomparison [van Zanten *et al.*, 2011]. For these simulations, SAMEX employs a horizontal grid spacing of 100 m and a constant vertical spacing of 50 m. Previous studies demonstrate the importance of using fine vertical grid spacing for representing cumulus congestus [Khairoutdinov *et al.*, 2009; Mechem and Oberthaler, 2013]. The number of grid points is 384 × 384 × 160 to yield a domain size of 38.3 × 38.3 × 8 km³, a volume that captures the cloud system mesoscale variability reasonably well. This experimental setup is employed as a starting point for the simulation suite. A simulation quadrupling the horizontal domain area (768 × 768 × 160 points) exhibits similar behavior to our baseline grid configuration for the majority of the simulation period.

SAMEX employs an explicit representation of microphysical processes [Kogan, 1991] that includes size-resolving representations of droplets and cloud condensation nuclei (CCN). The model predicts distribution functions for CCN and liquid droplets and solves the stochastic collection equation for the droplet spectra at all model grid points. As mentioned in section 3.2, our simulations assume that warm-rain microphysical processes dominate. The SAMEX configuration for our simulations uses 19 CCN bins and 34 mass-doubling droplet bins that range in size from 1 μm to 2.05 mm in radius. The number of bins is the same as that used in the RICO trade cumulus [van Zanten *et al.*, 2011] and Variability of the American Monsoon Systems (VAMOS) Ocean–Cloud–Atmosphere–Land Study (VOCALS) southeast Pacific stratocumulus [Mechem *et al.*, 2012] simulations. We assume an initial CCN concentration of 425 cm⁻³ (at a supersaturation $S = 0.47\%$) as directly measured from the ARM CCN counter at SGP, distributed in a lognormal spectrum with geometric mean radius of 0.03 μm and standard deviation of 1.28 μm. This spectral shape is similar to that used in previous RICO and VOCALS simulations. In our experience, simulation results are more sensitive to differences in total aerosol concentration than they are to modest differences in aerosol distribution shape. We more rigorously address this assumption in section 4.3. Our simulations neglect shortwave and longwave heating, which we justify given that the magnitude of the large-scale forcing appears to be the most dominant factor governing cloud system evolution. Short-duration simulations including longwave and shortwave radiative forcing exhibit only minor differences relative to the simulations without radiative forcing.

4.2. Model Initial Conditions and Forcing

All simulations use the semidiagnostic modeling framework in which the model responds to observationally constrained forcing fields of temperature and moisture [e.g., Wu *et al.*, 1998; Xu *et al.*, 2002, 2005; Xie *et al.*, 2005; Zeng *et al.*, 2009; Mechem and Oberthaler, 2013]. Figure 2 shows the large-scale advective tendencies, vertical motion, and surface heat and moisture fluxes imposed as forcing for the simulations. Although we refer to these quantities as “large scale,” they are characterized by 75 and 150 km analysis scales shown in Figure 1 and therefore include some degree of mesoscale variability, in addition to contributions from the synoptic and planetary scales. Table 1 summarizes the forcing configurations for the seven simulations in our ensemble, largely differing in the way the forcing fields are imposed (e.g., time-varying versus steady state).

The forcing terms in Table 1 are an attempt to distill each forcing profile (which vary in the vertical and, depending on the simulation, in time) shown in Figure 2 into a single “bulk” number for each forcing quantity.

The control simulation (1200v_75) assumes horizontally homogeneous initial profiles of temperature, moisture, and momentum from the 75 km scale variational analysis at 1200 UTC. Over the course of the 9 h simulation, profiles of time-varying horizontal advection and vertical velocity are imposed on the domain (see the shaded boxes covering the 1200–2100 UTC period in Figures 2a, 2c, and 2e). We consider the first 3 h of the simulation as a model spin-up period, until the integrated turbulent kinetic energy (not shown) reaches a near-steady state. Thus, the discussion of simulation results in section 5 includes only the final six simulation hours. Over the simulation period from 1200 to 2100 UTC, all vertical levels below 600 hPa exhibit negative potential temperature (cold) advection (Figures 2a and 2b). Since the advection over this layer is largely constant with height, we surmise that the effect on static stability over the cloud layer is minimal; in other words, the cold advection has neither a stabilizing nor a destabilizing influence. However, all else being equal, stronger cold advection will promote saturation (and hence cloud development) more quickly. Moisture advection (Figures 2c and 2d) is significantly positive below 800 hPa, which should enhance instability over the course of the simulation. At 1200 UTC, subsidence dominates the large-scale vertical motion field below 600 hPa (particularly at the 75 km scale; Figure 2e), and the magnitude of subsidence decreases over the course of the simulation. Thus, the stabilizing effect of subsidence on the mean stratification should decrease with time.

The potential temperature and moisture advection at the 150 km scale (1200v_150 simulation) is qualitatively similar to those terms at the 75 km scale, although the moistening is shallower and overlaid by a layer of weak drying. The large-scale vertical velocity fields, however, differ substantially. The layer of subsidence below 650 hPa that is slowly weakening with time at the 75 km scale is replaced at the 150 km scale with an elevated layer of ascent. Note that the values of vertical velocity over the course of the simulations are rather weak relative to the strong ascent associated with the frontal passage that occurred between 0200 and 0300 UTC. Thus, the effect of the large-scale vertical velocity on simulation outcomes is not obvious a priori.

All other simulations assume that the forcing is constant in time. For example, in the 1800s_75 simulation, initial conditions are taken from 1800 UTC, and the same 1800 UTC forcing is imposed over the entire 9 h simulation. These steady state, fixed-in-time simulations serve to evaluate the importance of the evolutionary nature of the forcing. Although we refer to the simulations as “steady state,” the simulations reach a condition resembling steady state behavior only during the final hours of the simulations. Moreover, imposing constant forcings over the course of a simulation may have unintended consequences. One simulation (1200s_75, not shown) develops a spurious, unphysical layer of midlevel cloud that results from constantly imposing moistening and cooling over the entire course of the simulation.

4.3. Additional Model Considerations

Our simulations employ CCN concentration as measured at the surface. We assume that this measured concentration is representative of CCN at cloud base, which in the well-mixed boundary layer present in this case is a reasonable assumption. We argue above in section 4.1 that simulated cloud properties are more sensitive to differences in total aerosol concentration than they are to modest differences in aerosol distribution shape. Thus, for this study we use the spectral shape we have employed previously in simulations of RICO and VOCALS cases. We evaluate this conjecture by conducting a variation of the 1200v_75 simulation where the total CCN concentration is maintained, but the breadth (standard deviation) of the CCN spectrum is increased by 50%. Differences between the two simulations (not shown) are miniscule, particularly for integrated quantities such as boundary layer depth and total precipitation. This finding is unsurprising given the strong congestus updrafts contain large values of supersaturation likely to activate all but the smallest of hygroscopic aerosol. Exploring differences in CCN concentration would be a natural choice for additional sensitivity simulations but is beyond the scope of the present study.

Because the cloud tops of some cumulus congestus within the domain may lie above the 0°C line, ice may be present. Congestus clouds are typically thought to be dominated by warm-rain, collision-coalescence processes [Lasher-Trapp *et al.*, 2001; Mechem and Oberthaler, 2013], though ice-phase processes may play a role in deeper congestus [Johnson *et al.*, 1999]. The bin microphysics parameterization represents the evolution of only the liquid part of the condensate. However, in an attempt to estimate the importance of

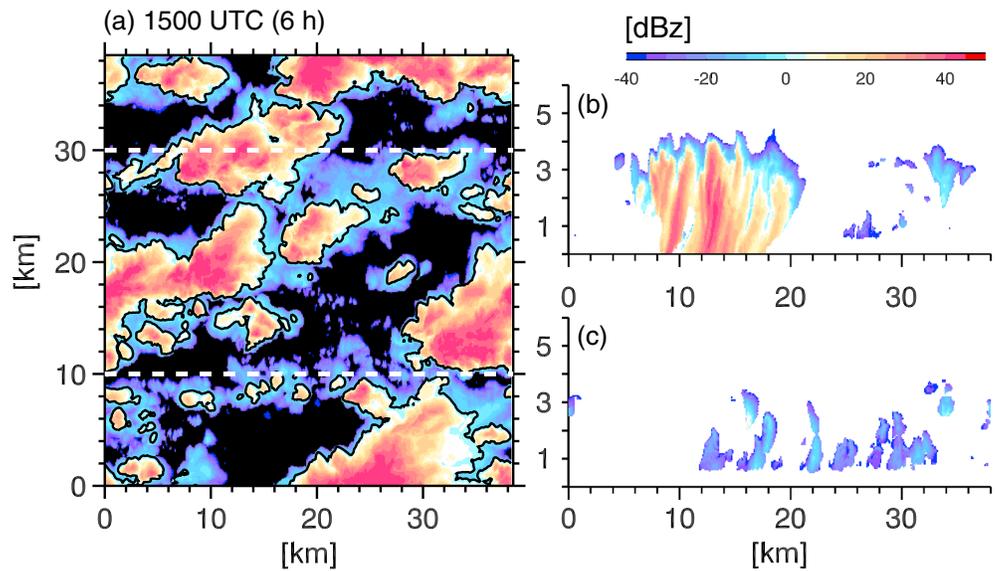


Figure 5. Snapshot of the simulated reflectivity field from the 1200v_75 control simulation, calculated directly from the bin microphysics drop-size distributions. (a) Plan view of column maximum reflectivity. Solid black contours are 0 dBZ, representative of typical C-band radar sensitivity. (b and c) Vertical cross sections taken through the northern and southern regions of the domain in Figure 5a.

ice-phase processes for this event, we conducted additional bulk microphysical simulations of the 1200v_75 and 1500s_75 cases. These simulations used the two-moment Morrison et al. parameterization [Morrison et al., 2005], with and without ice-phase processes. For these tests, ice-phase processes had only a modest effect on the simulation results, increasing accumulated precipitation by ~8% and ~6% in the 1200v_75 and 1500s_75 simulations, respectively. Furthermore, radar observations for this event also suggest the reasonableness of the warm-rain assumption. In particular, the observed cloud behaviors in Borque et al. [2014] demonstrate the lack of a prominent radar bright band/ice-media melting signature. When a bright band is visible on radar, very little cloud depth extends above it.

5. Ensemble Performance for the 25 May 2011 Event

A snapshot of the model-generated radar reflectivity field for the 1200v_75 control simulation at 1500 UTC shows an array of shallow cumulus, deeper cumulus clouds, and strongly precipitating cumulus congestus (Figure 5). Mean LES ensemble member performance metrics and time series of LES ensemble and observational results for the 25 May 2011 event are presented in Tables 2 and 3 and Figures 6–8, respectively. The tables document LES results according to two periods of 3 h averages of the cloud field properties that are representative of “early event” periods (3 h to 6 h simulation time; 1500–1800 UTC; Table 2) and “late-event” periods (6 h to 9 h simulation time; 1800–2100 UTC; Table 3). These periods were determined according to the observed precipitation onset time (~6 h simulation time; 1800 UTC). For the time-varying forcing configurations (labeled

Table 2. Summary of Simulation Results From the Early Period of the Simulation (3–6 h)^a

Simulation Name	CF	CT Height (km)	Precipitation Onset Time (h)	Precipitation Area Fraction	Precipitation (mm)
1200v_75 (Control)	0.68 (0.71)	1.80 (2.57)	5.0	0.014–0.021 (<0.01)	0.16
1200v_150	0.74 (0.79)	2.40 (3.04)	4.8	0.014–0.024 (<0.01)	0.28
1500s_75	0.65 (0.66)	2.85 (3.92)	3.3	0.044–0.060 (0.043–0.060)	0.59
1500s_150	0.54 (0.56)	2.16 (3.64)	5.1	0.019–0.026 (<0.01)	0.10
1800s_75	0.73 (0.71)	3.57 (4.56)	3.9	0.047–0.064 (0.053–0.077)	0.52
1800s_150	0.87 (0.93)	3.89 (4.92)	4.0	0.050–0.066 (0.063–0.080)	0.57
2100s_75	0.91 (0.93)	2.47 (2.79)	4.1	0.010–0.020 (0.008–0.016)	0.33

^aData columns represent mean cloud fraction (median in parentheses); median cloud top (CT) height (90th percentile height in parentheses); precipitation onset time (defined as the time when the surface precipitation rate first reaches the threshold value of 1 mm day⁻¹); range of mean and median precipitation area fraction (area fraction of 30 and 25 dBZ echo); and total precipitation accumulated over the 3 h period.

Table 3. Summary of Simulation Results From the Late Period of the Simulation (6–9 h)^a

Simulation Name	CF	CT Height (km)	Precipitation		
			Onset Time (h)	Precipitation Area Fraction	Precipitation (mm)
1200v_75 (Control)	0.63 (0.61)	3.56 (4.49)	5.0	0.053–0.074 (0.059–0.078)	1.01
1200v_150	0.72 (0.71)	4.10 (4.88)	4.8	0.048–0.067 (0.047–0.059)	0.97
1500s_75	0.51 (0.55)	3.64 (4.73)	3.3	0.038–0.049 (0.032–0.043)	1.07
1500s_150	0.41 (0.38)	3.07 (6.10)	5.1	0.045–0.056 (0.045–0.051)	0.51
1800s_75	0.60 (0.60)	4.39 (5.48)	3.9	0.050–0.066 (0.053–0.069)	0.76
1800s_150	0.51 (0.48)	3.98 (5.91)	4.0	0.057–0.071 (0.058–0.074)	0.89
2100s_75	0.86 (0.87)	2.88 (3.29)	4.1	0.024–0.041 (0.024–0.039)	0.90

^aColumn quantities are defined as in Table 2.

“V” in Table 1), early and late-event breakdowns should roughly correspond to the intervals of shallow cumulus (early) and subsequent precipitating congestus (late) conditions. For the steady state runs (labeled “S” in Table 1), early period averages reflect LES solutions from the 3 h period (3 h to 6 h) that immediately followed the 3 h model spin-up. Late-event results highlight the cloud properties from the final 3 h period (6 h to 9 h), at a time when the observations and all featured simulations indicate precipitating clouds.

5.1. Cloud Fraction

Cloud fraction observations and simulation results are summarized in Figure 6. The different observational CF estimates are consistent (~70–90% CF) over the later period, when congestus were present. A significant spread in estimates was observed during the shallow cumulus period prior to precipitation onset (~10–90% CF). The VISST satellite-based product (“GOES low+med” in Figure 6) estimated the largest CF, with a sizeable component of this CF value attributed to middle to upper level clouds (~8 km altitude) wrapping into the northern sections of the domain. The CF values estimated by narrower field-of-view ARM sensors (TSL, SIRS, and ARSCL) were consistent with campaign visual reports of scattered low cumulus transitioning to a more overcast cloud field populated by congestus (e.g., Figures 4 and 6). The SACR CF estimate adopted by this study

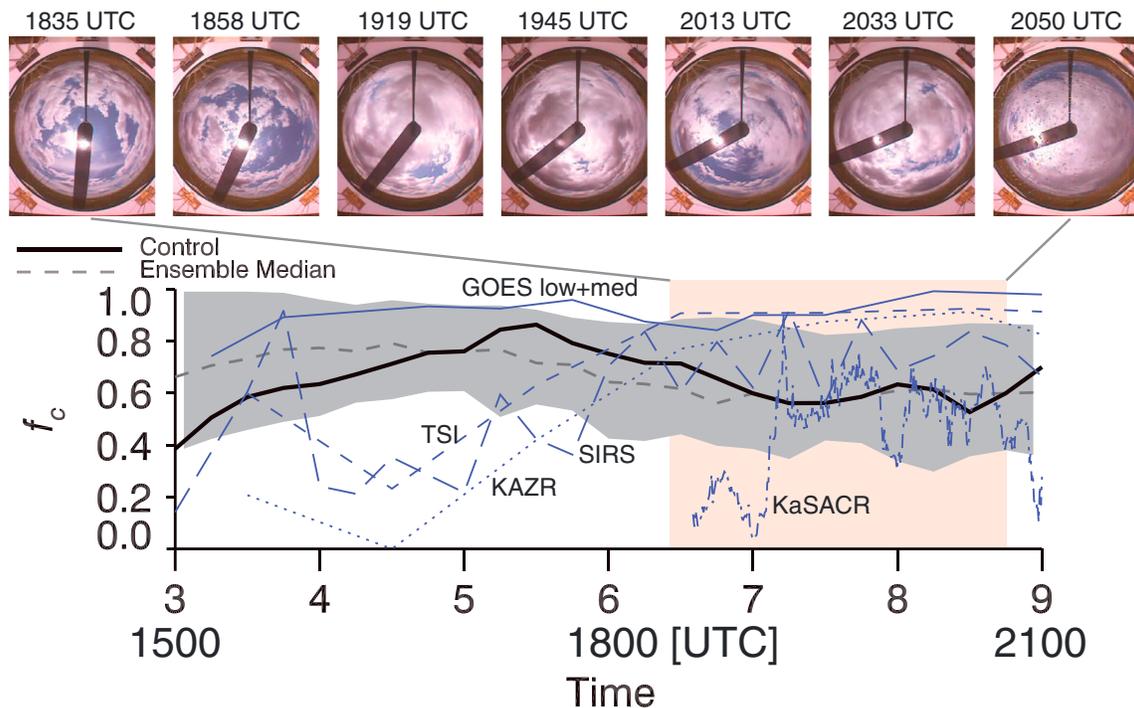


Figure 6. (top) A series of imagery from the TSI. Five observational estimates of cloud fraction, all denoted by (bottom) blue lines overlaid on cloud fraction calculated from the simulation suite. The black line represents the control (1200v_75) simulation. The dashed gray line is the ensemble median, and the gray bands represent the minimum and maximum cloud fraction values of the simulation ensemble. The rose-shaded area indicates the period covered by the TSI images.

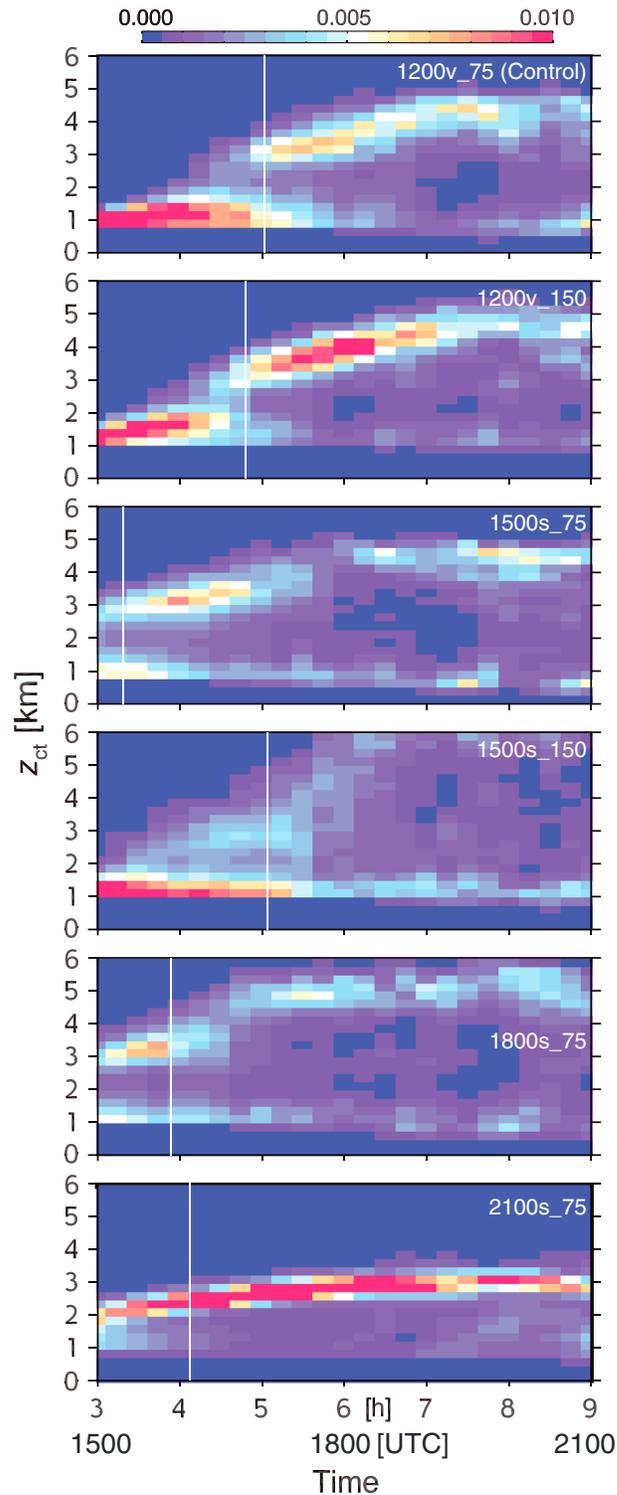


Figure 7. Time-varying probability density functions (PDFs) of cloud top height for six of the simulations. The white line represents the time of precipitation onset, taken to be a precipitation rate of 1 mm day^{-1} at the surface.

was unavailable prior to 1830 UTC and when available reported the lowest CF values when congestus were present (~60%). TSI and satellite images indicate that the discrepancy is partially the consequence of directional sampling limitations. Specifically, the SACR scanning prioritized a direction along the prevailing northwesterly wind such that half of the radar RHI was performed toward the southeast, the region least favorable for cloud development.

For the early cloud period, all LES solutions ostensibly overestimated CF (Figure 6, gray and black shading and control lines; Table 2) compared to the median observational estimate of 0.49 calculated from 1500 to 1800 UTC. Comparisons with CF observations suggest that the most credible shallow cumulus representation was obtained when applying time-varying forcing conditions initiated at 1200 UTC (1200v_75, 1200v_150). These two forcing configurations include the constraints that we expected would best capture the time evolution of cloud system properties for this challenging transition event. A reversal in model-observation CF characteristics was observed following precipitation onset, with the ensemble members underpredicting the CF associated with the congestus mode as compared to the observational consensus (e.g., Figure 6 and Table 3). One noteworthy exception was the 2100s_75 simulation that predicted extensive cloud coverage associated with low-topped cumulus (cloud top heights of ~3 km, Table 3). The time-evolving forcing configurations, and the steady state configurations corresponding to the times when congestus were present, typically resulted in higher CF values and better agreement with CF observations during the late-event period. Beyond 9 h, several of the steady state runs trend toward a domain-limited behavior of a few prominent cloudy cells.

The usefulness of CF comparisons beyond this basic qualitative agreement is not clear. Cloud fraction is an important quantity via its influence on the radiation budget, and therefore it constitutes a bulk measure of the cloud system that models should strive to simulate. However, as a tool for diagnosing model pathologies

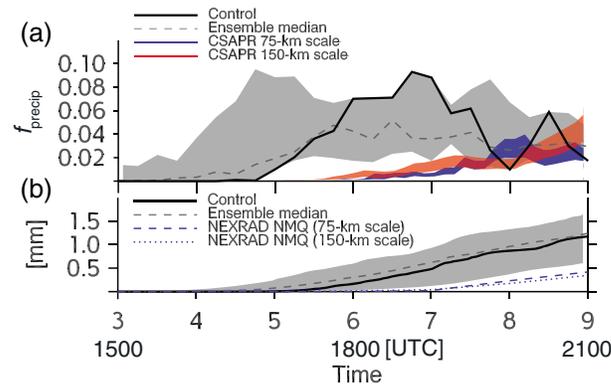


Figure 8. (a) Time series of echo area fraction from the C-SAPR and the simulation ensemble. The width of the bands two C-SAPR time series represents precipitation area defined by area bound by 30 and 25 dBZ values. The dashed gray line is the ensemble median, and the gray bands represent the ensemble minimum and maximum, corresponding to a 30 dBZ reflectivity threshold. The black line represents the echo area fraction of the control (1200v_75) simulation. (b) Accumulated precipitation from the model ensemble (black, dashed gray, and gray envelope), along with observational estimates at two scales from the NEXRAD NMQ.

(i.e., the root cause of model-observation mismatch and how to fix it), CF is inadequate. Furthermore, the SAMEX modeling framework is highly idealized and assumes horizontally homogeneous initial conditions and forcing. For this study, the prevailing forcing of positive moisture advection and negative potential temperature advection (e.g., Table 1) is conducive to simulating the observed transition from shallow clouds to a field that includes both shallow cumulus and congestus. The homogeneity of forcing, however, dictates that clouds initiate with an equal likelihood everywhere within our LES domain. In other words, cloud behavior is statistically homogeneous across the model domain, in contrast to the highly variable spatial structure visible in Figure 1. The SACR observations, on the other hand, indicate a cloud population that is not statistically homogeneous. For this reason,

CF metrics between the statistically homogeneous simulations and the horizontally heterogeneous observational cloud field may differ substantially.

During the later period when congestus are present, LES CF values are closer to the observations (median value of 0.85 from 1800 to 2100 UTC). Based on similar arguments as above, this agreement may be fortuitous. We note that LES solutions progress toward stronger isolated congestus reminiscent of the deep precipitating congestus clouds sampled by the C-SAPR. However, the LES behavior of a simple evolution from shallow cumulus to isolated deeper congestus conflicts with the complexity found in the ARM CF observations that favor the coexistence of a spectrum of cloud depths; these include the deep, strongly precipitating congestus and also weaker congestus elements and commingled shallow cumulus clouds. Although we cannot address this discrepancy with certainty, we speculate that the differences may arise from the highly idealized nature of the forcing, particularly in two respects: First, since the piecewise continuous forcing is applied at 3 h intervals, we should not necessarily expect the model to reproduce fine-scale cloud behavior in time. Second, the forcing profiles and doubly periodic boundary conditions impose an assumption of statistical homogeneity across the simulation domain, whereas the observed transient cloud system exhibits substantial horizontal variability over the same scale.

5.2. Cloud Top Distributions

LES cloud field median and 90th percentile tops for all simulations (Tables 2 and 3) were consistent with SACR and ARSCL observations of the tail end of distributions of shallow cumulus transitioning to later congestus. The 90th percentile values typically range from top heights less than 4 km early to heights of 5.5–6 km at later stages. In this simple manner, the two-dimensional SACR observations are immediately useful for providing a more realistic range for cloud top and bottom heights (e.g., Figure 3) as compared to traditional column-based radar observations (e.g., Figure 4). Further, SACR time series depictions of the cloud top distributions (Figure 3) provide additional support for scrutinizing the evolution of LES cloud fields. For this event, the results emphasize the benefits of forcing that closely mimics observations (either time varying or well matched to the observational period of interest). For example, the most accurate depiction of the cloud top distribution is found when using the time-varying forcing (1200v_75, 1200v_150). These simulations properly capture the bimodal cloud top features in the observations (Figures 3 and 4) and suitably delay the initiation of significant congestus to better match with precipitation onset timing (see the precipitation onset lines in Figure 7 and the evolution of precipitating echo area in Figure 8). The 1500 UTC and 1800 UTC steady state forcings also developed bimodal cloud top behaviors (Figure 7)

characteristic of those observation times. The reason for the bimodal cloud behavior is not clear, but preliminary analysis suggests that variations in low-level stability may lead to differences in cloud depth, precipitation initiation, and subsequent cloud system behavior. The early steady state forcing conditions (1200 UTC) preclude rapid development of precipitating congestus, whereas using later steady state forcing (2100 UTC) results in a cloud population with characteristic tops inconsistent with observations (e.g., Figure 7). Although SACR insights are beneficial, the limits of the SACR scanning strategy employed for the 25 May 2011 case suggests that additional radars and scanning modes are required to fully characterize three-dimensional distributions of cloud properties.

5.3. Precipitation Behavior

In following the progression toward deeper congestus, LES ensemble members overestimate the areal coverage and amount of precipitation (Figure 8 and Table 3). The temporal variability of precipitation fraction (area fraction ~ 0.01 – 0.1) and the mean precipitation over the domain from the LES lacks the slower, gradual buildup of these precipitation quantities as viewed by the C-SAPR. Nevertheless, new radar insights still suggest some benefit to well-matched forcing conditions. For example, early (3 h–6 h) sequences from the 1800 UTC runs (Table 2) predicted reasonable 3 h average or accumulated behaviors for radar cloud tops (~ 4.5 – 5.0 km), CF (~ 0.7 – 0.9), precipitation totals (~ 0.5 – 0.6 mm), and areal fraction (~ 0.05 – 0.08), as compared to the observations between 1800 and 2100 UTC. Of all the simulations, the 1200 UTC time-varying runs performed best at capturing the extended event and cloud transition, as gauged by the time-evolving areal coverage and rainfall totals. In particular, the later precipitation onset in these simulations better matched the observed precipitation onset. Precipitation totals for the two time-varying simulations were in the middle of the ensemble (~ 1.2 mm of total precipitation), and the modest areal fraction (< 0.01 early, 0.05 late) compared favorably to observations.

The authors include forcings at multiple scales (75 km, 150 km) in Table 1 to highlight the challenges when attempting to interpret the impact of forcing scale on precipitation behavior. Total precipitation rates for the 1800s_75 and 1800s_150 simulations are similar (1.28 versus 1.46 mm day $^{-1}$), yet the temperature advection, moisture advection, and vertical motion differ substantially (Table 1). These results suggest a balance whereby the weaker temperature and moisture advection in the 150 km simulation are compensated for by enhanced upward vertical motion. We note that strong large-scale ascent was not typical of the model configurations that represented precipitation onset timing favorably relative to the observations. For example, the run with the highest precipitation (1500s_75) featured the least favorable large-scale vertical motion but the highest moisture advection. Ultimately, the ability to predict the evolution of cloud and precipitation properties from these simple bulk measures of forcing in Table 1 is limited. This simplistic interpretation of the forcing also fails to diagnose unexpected behavior in select runs, such as the 2100s_75 simulation that initiates densely packed precipitating cumulus that bear no resemblance to the cloud radar observations. As we mention above in section 5.2, results suggest that differences in low-level stability may be important in controlling the initial precipitation onset, a notion we will explore in further study of our simulation ensemble.

For radar-based observations from the C-SAPR and WSR-88Ds, the choice of spatial scale had little impact on the precipitation properties considered. Although the choice of scale was influential to the performance of several of the LES runs, there was no indication that a particular LES forcing configuration (75 or 150 km) was consistently preferable. The finding may be loosely attributed to the offsetting nature of several bulk forcing measures discussed above. Specifically, stronger large-scale upward motion was associated with reduced moisture and potential temperature advection contributions. Differences in forcing contributions do not always combine to produce such a serendipitous balance, however. For example, the 1500 UTC simulations exhibited a dramatic spread between the 75 km and 150 km domain variants (e.g., Tables 2 and 3), with the 1500s_150 run yielding the smallest 6 h total precipitation (0.61 mm), in best agreement with the observations. In comparison, the 1500s_75 run rapidly initiated precipitation and predicted the highest 6 h total (1.66 mm) but better reflected the areal and cloud field properties of the observations.

6. Conclusions and Key Findings

SAMEX solutions with forcing more closely matched to observations captured several key aspects of the 25 May 2011 cloud event. Our results were encouraging, given the substantial synoptic variability and the highly idealized modeling framework. Here we present our initial investigations into the sensitivity of LES solutions to

changes in spatial and temporal large-scale domain forcing. Our emphasis was to evaluate whether newer scanning radar technologies for cloud and precipitation properties would benefit our effort to assess the influence of the transient nature of the forcing (i.e., time-varying versus steady state) on simulation outcome. A summary of the key findings is as follows.

Our initial findings indicate that measurements obtained from scanning radar such as cloud top height distributions better highlighted the differences across the SAMEX simulations. SAMEX simulations initiated using the time-varying forcing configurations better captured cloud regime transitions for this event. While this result may be expected, a preference toward these time-varying runs may not have been as apparent if evaluated according to simpler bulk metrics of cloud properties (cloud fraction, base, or top). There were promising indications that an improved characterization of cloud field spatial inhomogeneities from multiple scanning radar observations or additional measures of horizontal cell distributions (from additional scanning modes) may help differentiate preferable model configurations. Here simplified attempts to consider more detailed, multidimensional measures of cloud and precipitation geometry from scanning radar systems (e.g., precipitation onset, precipitation area, cloud top probability density functions (PDFs)) demonstrated several key advantages for the time-varying forcing. For example, the cloud top PDF formulated from the SACR (Figure 3) and KAZR-ARSCL (Figure 4) were instructive for highlighting a transition to a bimodal cloud system behavior, helping to identify which LES runs were capable of reproducing these features (Figure 7). Area-mean model precipitation fields exhibited greater short-term temporal variability relative to the observations but were reasonable when averaged over an appropriate timescale. Some of these discrepancies may be attributed to the demanding nature of attempting to model isolated congestus using idealized LES simulations over a relatively small domain forced with horizontally homogeneous conditions.

For this event, the characteristic scale of the analysis domain (75 km, 150 km) did not yield significant differences in the observations, although select simulation results were found sensitive to scale of the forcing (e.g., differences in precipitation for the two 1500 UTC simulations). Linking persistent biases in simulation results to differences in the scale of the forcing or bulk measures of forcing terms (e.g., Table 1) was difficult, however, suggesting that bulk representations of forcing quantities are insufficient in understanding cloud system evolution. Understanding possible secondary linkages was also challenging. For example, total precipitation was unrelated to the LES solutions having the deepest congestus tops or greatest areal precipitation coverage. While we were encouraged by our initial explorations using scanning radar, more complete observations of cloud properties are required (multiple radar views, improved scan modes) to better constrain three-dimensional cloud behavior (e.g., cloud thickness, cloud field topology).

Most CRMs or GCMs, including models used in previous studies to represent the transition from shallow cumulus to deeper precipitating clouds, have employed bulk rather than bin microphysical parameterizations [e.g., Grabowski *et al.*, 2006; Kuang and Bretherton, 2006; Khairoutdinov and Randall, 2006; Wu *et al.*, 2009; Hohenegger and Stevens, 2013; Mechem and Oberthaler, 2013]. Our choice of bin microphysics was made in order to eventually foster the most accurate comparisons possible with the ARM spectral radar observations. Our simulation ensemble framework and comparison with a wide range of observational metrics derived from ARM observations will enable future evaluations of different bulk microphysical parameterizations. Performing both bin and bulk LES runs will promote not only comparisons of cloud macroscale properties but also evaluations of the validity of assumptions inherent in the bulk parameterizations (e.g., assumed drop-size distributions or calculations of dispersion in two-moment parameterizations).

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