Clouds and Water Vapor in the Tropical Tropopause Transition Layer over Mesoscale Convective Systems

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ABSTRACT

Observations from A-Train satellites and other datasets show that mesoscale convective systems (MCSs) affect the water vapor and ice content of the tropical tropopause transition layer (TTL). The largest MCSs with radar reflectivity characteristics consistent with the presence of large stratiform and anvil regions have the greatest impact. Most MCSs are associated with clouds in the TTL. Composites in MCS-relative co-ordinates indicate enhanced cloudiness and ice water content (IWC) extending toward the cold-point tropopause (CPT), particularly in large and connected MCSs. Widespread anvils in the lower TTL are evident in the peak cloudiness diverging outward at those levels. Upper-tropospheric water vapor concentrations are enhanced near MCSs. Close to the centers of MCSs, water vapor is suppressed at TTL base, likely because of the combined effects of reduced moistening or dehydration at the higher TTL relative humidities and subsidence above cloud top. Weak moistening is observed near the CPT, consistent with sublimation of ice crystals at the tops of the deepest MCSs. In the outflow region, moistening is observed in the lower TTL near the largest MCSs. Enhanced water vapor in the upper troposphere and lower TTL extends beyond the area of substantially enhanced cloudiness and IWC, in agreement with the observed radial outflow, indicating that MCSs are injecting water vapor into the environment and consistent with the possibility that MCS development may be favored by a premoistened environment.

1. Introduction

Deep convective clouds in the tropics sometimes aggregate and develop extensive regions of stratiform precipitation and nonprecipitating anvil. Large, organized cloud features of this type are called mesoscale convective systems (MCSs; Houze 2014, chapter 9). Over half of tropical precipitation and latent heating is produced by MCSs (Yuan and Houze 2010, hereafter YH10). The latent heating in MCSs is top heavy as a result of their significant stratiform precipitation areas (Houze 1982, 1989), and this top-heavy heating is important in forcing the large-scale tropical circulation (Hartmann et al. 1984; Schumacher et al. 2004). The net radiative heating in MCSs is a positive maximum in the upper troposphere, with the strongest heating near anvil cloud top (Houze 1982).

Observations from the CloudSat satellite, launched in 2006, have permitted the analysis of the vertical structure of clouds around the world. Cetrone and Houze (2009) and Yuan et al. (2011, hereafter YHH11) have used CloudSat to examine characteristics of tropical MCS anvils. The distribution of CloudSat reflectivities varies with anvil thickness and distance from the MCS center and is consistent with the detrainment of ice particles from the MCS updrafts into the anvils and the sedimentation, or gravitational settling and eventual removal, of the larger particles as the anvil age increases. Both studies noted contrasts between the anvil structures of land and oceanic MCSs. Land MCSs exhibit broader and higher reflectivity distributions consistent with more intense convection and less extensive stratiform rain areas, while oceanic MCSs exhibit narrower distributions with modal intensity strongly decreasing with height, consistent with the development of broad stratiform areas.

MCS vertical structure in the troposphere has implications for the systems’ impacts on the overlying tropical tropopause transition layer (TTL). The TTL extends...
from around 14 km, just above the level of main convective outflow, to the highly stable stratosphere above 18 km (Fueglistaler et al. 2009; Randel and Jensen 2013). Air below the TTL undergoes net radiative cooling as it subsides toward the ground, while air in the TTL tends to ascend slowly while undergoing net radiative heating. The TTL links the rising branch of the tropospheric Hadley circulation, whose ascent is concentrated in areas of tropical moist convection, and the broader rising branch of the stratospheric Brewer–Dobson Circulation. The TTL is characterized by extremely low temperatures (Anthes et al. 2008; Kim and Son 2012), and air ascending through the TTL undergoes freeze drying, producing the low stratospheric water vapor concentrations noted by Brewer (1949).

The mean level of neutral buoyancy (LNB) in convective regions of the tropics lies between 12 and 14 km, just below TTL base (Takahashi and Luo 2014). Di vergent outflow from regions of tropical deep convection occurs in the upper troposphere and at TTL base (Highwood and Hoskins 1998; Folkins and Martin 2005). Based on CloudSat observations of anvil characteristics, Takahashi and Luo (2012) concluded that maximum mass detrainment from convection is around 11 km, just below the LNB determined from atmospheric soundings. The geographic distribution of cirrus in the upper troposphere strongly resembles the distribution of tropical precipitation, consistent with cirrus forming in and spreading out from the upper portions of deep convection (Virts et al. 2010). Convectively generated cirrus also occur in the TTL (Massie et al. 2002; Luo and Rossow 2004), although in situ formation in ascending layers becomes increasingly important at higher altitudes (Riihimaki and McFarlane 2010; Virts et al. 2010). Trajectory analysis suggests that cirrus can persist up to 2 days after convection weakens and be advected up to 1000 km from the region of the parent convection (Luo and Rossow 2004). As this advection occurs, sublimation of the smaller ice crystals adds vapor to the environment (Soden 2004).

Observations across the tropics indicate that some deep convection overshoots its LNB and penetrates into the TTL. Overshooting occurs over both ocean and land, although the most vigorous convection is concentrated over land (Alcala and Dessler 2002; Liu and Zipser 2005; Zipser et al. 2006). Alcala and Dessler (2002) found that approximately 5% of convective elements observed by the Tropical Rainfall Measuring Mission (TRMM) radar penetrated into the TTL. Takahashi and Luo (2012) reported that mean cloud-top height of the deepest convective cores observed by CloudSat is 15 km, within the TTL. However, less than 1% of deep convective clouds reach the cold-point tropopause (CPT; Alcala and Dessler 2002; Gettelman et al. 2002; Dessler et al. 2006). These observations suggest that tropical convection affects the moisture budget of the TTL.

Analysis of the impact of deep convection on water vapor profiles in the TTL has addressed the fundamental question of whether the net contribution is to moisten or dehydrate the environmental air. Relative humidity with respect to ice increases with height in the upper troposphere and TTL because of the decrease in saturation mixing ratio in the ascending cold air, and supersaturation frequently occurs in the TTL (Luo et al. 2007; Fueglistaler et al. 2009). Previous studies have emphasized the importance of the ambient relative humidity in determining the impact of convection: convection in subsaturated regions moistens the environment, while in supersaturated regions net dehydration is observed, because excess water vapor is deposited onto the ice particles, which may be removed via sedimentation (Jensen et al. 2007; Wright et al. 2009; Hassim and Lane 2010). Chae et al. (2011), analyzing water vapor anomalies from the Microwave Limb Sounder (MLS) below the cloud-top heights identified in data from the Cloud–Aerosol Lidar and Infrared Pathfinder Satellite Observations (CALIPSO) satellite, observed strong moistening in the upper troposphere and decreased moistening above TTL base, shifting to dehydration above 16 km if the tops of clouds were above this level, consistent with less effective moistening with increasing background humidity. They furthermore emphasized that the mechanism they posited is not restricted to convective towers and cirrus outflow but could apply to any cirriform cloud; the mechanism could occur in situ or in the upper reaches of stratiform or anvil tops of MCSs.

As mentioned above, a fraction of deep convective clouds penetrate the CPT. The penetrating towers contain extremely cold, dry air and have been hypothesized as a mechanism for lower-stratospheric dehydration (Danielsen 1993; Sherwood and Dessler 2000). However, modeling studies of such clouds indicate net hydration because they leave behind small ice particles that undergo sublimation in the relatively warmer stratosphere (Chaboureau et al. 2007; Jensen et al. 2007). These modeling results have been confirmed by observational evidence (e.g., Corti et al. 2008; Khaykin et al. 2009). However, the horizontal scale of the penetrating cores of active convection is very small, and it seems unlikely that they alone can account for the water vapor properties of the TTL.

Rossow and Pearl (2007) concluded that most convection penetrating near the CPT is associated with organized (i.e., large) MCSs; however, to date little analysis has been attempted of MCS characteristics in the TTL or how MCSs might affect the moisture content of the TTL. Yet MCSs dominate much of the upper-level
cloud coverage of the tropics. The stratiform portions of
MCSs are orders of magnitude larger in scale than the
convective towers embedded in the MCSs, and the strat-
iform regions are buoyant and have very cold tops rising
and diverging over wide areas, suggesting that they might
have a substantial effect on the moisture of the TTL. In
this paper, we investigate the water vapor and ice content
of MCSs in the upper troposphere and TTL to determine
the impact of these larger convective entities on the water
content at upper levels.

To identify MCSs, we use the dataset of YH10, who
introduced a methodology for identifying MCSs in ob-
servations from NASA’s A-Train satellites (L’Ecuyer
and Jiang 2010): high, cold cloud tops observed by the
Moderate Resolution Imaging Spectroradiometer
(MODIS) and precipitating areas observed by the Ad-
vanced Microwave Scanning Radiometer for Earth
Observing System (AMSR-E; see section 2 for details).
Other instruments aboard A-Train satellites observe
cloud vertical structure (CloudSat), including that of
optically thin cirrus layers (CALIPSO), and measure
profiles of ice water content (IWC) and water vapor
concentration in the TTL (MLS). Collectively, these
observations offer a unique opportunity to analyze the
vertical structure of MCSs and their contribution to
clouds and moisture in the TTL. We analyze these ob-
servations of water vapor and ice in the vicinity of each
MCS in the YH10 database.

The remainder of the paper is organized as follows: the
YH10 MCS identification scheme is described in section
2, along with the other satellite datasets used in this study.
The climatological distributions of MCSs and of clouds
and moisture in the TTL are shown in section 3. MCS
vertical structure and microphysical characteristics in
the troposphere are discussed in section 4, while the
distribution of clouds, ice, and water vapor in MCSs in
the TTL is presented in section 5. Conclusions are in
section 6.

2. Data

a. Observations from A-Train satellites

NASA’s A-Train constellation operates in a sun-
synchronous orbit, crossing the equator around 1:30
and 13:30 LT. The A-Train instruments used in this
study, including the YH10 MCS database derived from
observations from two of those instruments, are de-
scribed in this section.

1) MCS DATABASE

Clouds and precipitating areas are two prominent
MCS components readily observed by satellites. The
YH10 methodology begins with the identification of
these components in data from two instruments on the
Aqua satellite (L’Ecuyer and Jiang 2010): high, cold cloud tops observed by the
Moderate Resolution Imaging Spectroradiometer (MODIS) and precipitating areas observed by the Advanced Microwave Scanning Radiometer for Earth Observing System (AMSR-E; see section 2 for details). Other instruments aboard A-Train satellites observe cloud vertical structure (CloudSat), including that of optically thin cirrus layers (CALIPSO), and measure profiles of ice water content (IWC) and water vapor concentration in the TTL (MLS). Collectively, these observations offer a unique opportunity to analyze the vertical structure of MCSs and their contribution to clouds and moisture in the TTL. We analyze these observations of water vapor and ice in the vicinity of each MCS in the YH10 database.

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1) MCS DATABASE
exhibit a greater range of sizes than CMCSs, so they are further subdivided into small SMCSs (the smallest 25% by area, with HCSs less than approximately 11 000 km$^2$) and large SMCSs (the largest 25%, with HCSs greater than approximately 41 000 km$^2$). In this study, we analyze all MCSs identified in the tropics (30$^\circ$N–30$^\circ$S) during 2007–10.

2) **CLOUDSAT REFLECTIVITY**

CloudSat carries a 94-GHz cloud profiling radar (CPR) that measures backscatter from clouds (Stephens et al. 2002; Marchand et al. 2008). The backscatter profiles are available every 2.5 km along track and have a vertical resolution of 240 m at nadir. A cloud mask is provided in the CloudSat geometric profile product (2B-GEOPROF). In this study, observations within clouds are identified by screening for cloud mask values greater than or equal to 20.

The CloudSat profiles analyzed in this study are those identified by YHH11 as sampling some portion of an MCS. In that study and in YH10, it was noted that clouds with tops above 10 km identified by CloudSat fall into two categories: precipitating clouds with bases near the surface and elevated anvils with bases primarily above the freezing level. Accordingly, they defined anvil clouds as having base above 3 km and top above 10 km. The same definition is used in this study. We will also present statistics on CloudSat profiles sampling nonanvil regions of the MCS. This category is dominated by profiles sampling the deep convective or stratiform precipitating areas of the system and is referred to herein as the precipitating category.

3) **CALIPSO CLOUD OCCURRENCE**

The Cloud–Aerosol Lidar with Orthogonal Polarization (CALIOP) is a two-wavelength polarization lidar carried aboard CALIPSO. Cloud layer base and top heights are identified from the lidar backscatter and are provided at a resolution of 60 m in the vertical in the TTL and 5 km along track. Following Fu et al. (2007), opaque features (those that completely attenuate the lidar signal) are assumed to be deep convective clouds and assigned a cloud base at Earth’s surface. While CloudSat’s estimated operational sensitivity (approximately −32 to −30 dBZ) is insufficient to see clouds with small IWC, CALIPSO can detect cloud layers with optical depths as low as 0.01 or less (Winker et al. 2007). In this study, we take advantage of the complementary capabilities of these instruments, using CloudSat to examine the vertical structure of the deep convective clouds and CALIPSO to analyze the extent of cirrus shields in the TTL.

4) **MLS ICE WATER CONTENT AND WATER VAPOR**

The MLS aboard the Aura satellite observes thermal microwave emissions in five spectral regions. We analyze MLS version 3 vertical profiles of IWC and water vapor mixing ratio, whose validations are described by Read et al. (2007) and Wu et al. (2008), respectively. Profiles are provided at 1.5° intervals along track and have vertical resolutions of 4 and 3 km, respectively, and horizontal resolutions of approximately 7 km across track and approximately 300 km along track. The useful altitude ranges are from 215 to 83 hPa for IWC and above 316 hPa for water vapor. The uncertainties for water vapor are 20% below 100 hPa and 10% at 100 hPa and above, while the uncertainty for IWC is a factor of 2, largely owing to the uncertainty from particle size assumption (Jiang et al. 2012). The data are screened following the recommendations in Livesey et al. (2011).

MLS also observes atmospheric temperature, with recorded biases of about 1.5 K in the TTL (Schwartz et al. 2008). For this paper, we apply an offset at each pressure level based on comparison with global positioning system (GPS) temperature observations, as given by Chae et al. (2011). The water vapor and corrected temperature observations are then used to calculate relative humidity with respect to ice (referred to herein as relative humidity). Because it is dependent upon the resolution of the temperature observations, the vertical resolution of the relative humidity is coarser than that of water vapor: approximately 5 km in the TTL (Schwartz et al. 2008; Livesey et al. 2011).

b. **ERA-Interim**

The horizontal winds ($u$, $v$) and vertical velocities in pressure coordinates ($\omega$) in this study are from the European Centre for Medium-Range Weather Forecasts (ECMWF) interim reanalysis (ERA-Interim; Dee et al. 2011). ERA-Interim fields are available four times daily at 1.5° resolution for five pressure levels in the upper troposphere and TTL (200, 175, 150, 125, and 100 hPa). The $u$ and $v$ fields are used to calculate the radial wind—that is, the component of the horizontal wind directed toward or away from the center of each MCS.

c. **Compositing technique**

For the results presented in section 5, each MCS in the YH10 database is assigned a separate coordinate system, where the origin corresponds to the center of the largest raining core. Observations from CALIPSO, MLS, and ERA-Interim are extracted and stored in system-relative coordinates—that is, as a function of
vertical height or pressure level and distance from the MCS. For MLS and ERA-Interim, the dataset pressure levels and horizontal sampling frequency (every 1.5°) are retained. As described earlier, CALIPSO observations are available at higher spatial resolution; however, in order to reduce noise, we calculate cloud occurrence for 0.25 km vertical and 1° horizontal bins. No further spatial filtering is applied to the data. When indicated, anomalies are calculated by subtracting the climatological monthly mean at each location prior to compositing.

3. Climatology of MCSs and the TTL

The annual distribution in Fig. 1 indicates that all three types of MCSs are observed throughout the deep convective regions of the tropics—the tropical continents, the intertropical convergence zones (ITCZs), and ubiquitously over the western Pacific warm pool. However, as shown by YH10, both large and small SMCSs are more frequently observed over land or near-coastal regions, while the merging of systems to produce CMCSs occurs mostly over the oceans, especially the warm pool.

Annual-mean distributions of clouds and moisture at 147 hPa, near TTL base, are shown in Fig. 2. Maxima in each field are located over the regions of intense convection over the tropical continents and the warm pool. Clouds and IWC are more strongly focused in areas with high annual-mean precipitation and frequent MCS occurrence (Fig. 1). Water vapor and relative humidity exhibit more diffuse distributions, indicating the advection of water vapor beyond the areas of frequent deep convection as well as the effect of the large-scale temperature field.

The climatological, zonal-mean vertical structure of tropical clouds and moisture is shown in Fig. 3. In the upper troposphere, clouds and ice are most frequent around 7°N—the latitude of the rising branch of the annual-mean Hadley circulation. A secondary maximum in IWC is observed south of the equator, associated with convection in the austral summer Hadley cell. IWC concentrations decrease with height above ~170 hPa, while cloud occurrence broadens and becomes more equatorially symmetric in the TTL, where cirrus can form either through the shearing off of convective anvils or in situ in rising, adiabatically cooling layers.

Water vapor concentrations decrease by about three orders of magnitude from the upper troposphere to the lower stratosphere (Fig. 3c). This strong decrease with height reflects the temperature dependence of the saturation mixing ratio, particularly at the low temperatures observed in the TTL. The decrease in the
infusion of water vapor into the environment by convection also contributes to the water vapor distribution at higher altitudes. Despite the low water vapor concentrations in the TTL, relative humidities are high (Fig. 3d), with zonal-mean values in excess of 80% because of the extremely low temperatures. The high relative humidities in the TTL promote cirrus occurrence.

Before considering the contribution of MCSs to clouds and water vapor in the TTL, it is useful to analyze their characteristics in the troposphere, specifically their vertical structure and microphysical attributes. In the following section, these characteristics are investigated using CloudSat observations.

4. Vertical structure of MCSs in the troposphere

The distribution of CloudSat reflectivities in MCSs as a function of height can be represented using contoured frequency by altitude diagrams (CFADs; Yuter and Houze 1995; Masunaga et al. 2008). Cloud microphysical processes can then be inferred from the reflectivity distributions. While YHH11 analyzed CFADs of anvil characteristics as functions of geographic region, anvil thickness, and distance from the MCS center, we examine reflectivity characteristics for both the precipitating portions and the nonprecipitating anvils of each category of MCS (Fig. 4). The primary features of the CFADs are shared by each type of MCS: in the precipitating areas, peak reflectivities of 10–15 dBZ are observed in the melting layer (around 4–5 km). Reflectivities decrease below the melting layer as a result of signal attenuation by the precipitation. High reflectivities dominate the distribution through most of the troposphere, indicating the presence of large droplets in the convective updraft below the melting layer and the formation of graupel and large ice particles above. In the upper troposphere, above 9–10 km, reflectivities generally decrease with height because of...
lower concentrations of particles of generally smaller dimension.

As air from the MCS raining cores is incorporated into the anvils, microphysical processes modify the characteristics of the ice particles. Reflectivities in MCS anvils are lower, generally below 10 dBZ (Fig. 4). Peak reflectivities are observed in the lower portions of MCS anvils, around 6–10 km, and the median reflectivity decreases with height to approximately ~20 dBZ near 13 km. Vertical velocities in anvils are generally weak, such that ice particles tend to drift downward (Houze 2014, chapter 6). The presence of larger particles near anvil cloud base reflects this sedimentation, as well as ice particle growth by aggregation (McFarquhar and Heymsfield 1996). YHH11 further noted weaker reflectivities and a narrowing of the reflectivity distribution in anvils as distance from the MCS core increased, consistent with aging.

While the CFADs for the three MCS categories share these basic similarities, key differences also exist. In the raining cores of small SMCSs, the reflectivity distribution exhibits a more vertical orientation, with median reflectivities remaining high (greater than 0 dBZ) up to 13 km. In contrast, in large and connected MCSs, reflectivities decrease more rapidly with height. To more clearly contrast the MCS types, difference plots are shown in Fig. 5. These plots were constructed by subtracting the CFADs for small SMCSs from those for large SMCSs, such that blue shading indicates the reflectivities proportionately more likely to be observed in small systems. In Fig. 5, above the melting layer, high reflectivities are more commonly observed in small SMCSs, indicative of strong convection and less developed stratiform regions. The reverse is observed above the melting layer in large SMCSs, where the narrower reflectivity distribution is consistent with the presence of stratiform rain areas. A narrower distribution is also observed in the anvils of large SMCSs compared to those in small SMCSs, as shown by the pattern of blue shading in Fig. 5b. The greatest contrast is observed in the occurrence of high reflectivities above 9 km, which are much more common in the small SMCSs, as the larger particles formed in the intense convective updrafts find their way into the anvils of the small MCSs.

5. MCS impacts on the TTL

Having examined MCS vertical structure in the troposphere, we now investigate the impact of MCSs on the overlying TTL. In this section, we present analyses of clouds, wind, IWC, and water vapor in the vicinity of the MCSs, with the variables composited in system-relative coordinates as described in section 2c.

a. Clouds and wind

Cross sections of composited cloud fraction in the upper troposphere and TTL near MCSs are shown in
the left column of Fig. 6. The basic cloud distribution is similar for all MCS types: while the largest cloud fractions are observed lower in the troposphere, cloud occurrence is frequent in the upper troposphere near the MCS center. About 74% of CALIPSO profiles within 100 km of the center of an MCS observed a cloud top above 150 hPa, at TTL base, and about half of such layers were opaque to the lidar (not shown). A cloud top above 120 hPa was observed in about 47% of the profiles. Above that level, cloud occurrence decreases more rapidly with height. Beginning within about 200 km of the MCS center and extending farther away, an outward bulge in cloud fraction indicates the spreading of cirrus anvils. This bulge is centered near 170 hPa near the MCS core and shifts upward to 150 hPa at greater distances.

Large and connected MCSs are associated with substantially greater cloud fractions in the upper troposphere and TTL than small SMCSs, particularly in the area within a few hundred kilometers of the MCS center, where the probability of CALIPSO observing a cloud in the upper troposphere is almost double near a large SMCS compared to a small SMCS. These differences lessen with distance, and at 1500 km from the MCS, composite cloud fractions are ~0.2 in each case. At such large distances, CALIPSO is likely observing neighboring clouds. To reduce the impact of clouds that may not be part of the MCSs, we subtract the climatological monthly mean cloud fraction at each grid point prior to compositing. Anomalous cloud fractions for each MCS type are shown in the right column of Fig. 6. Here, the contrast between small SMCSs and large and connected MCSs is even more striking. Enhanced cloudiness near small SMCSs is observed through the depth of the TTL but is confined to within about 200 km of the MCS center, and the magnitude of the anomalies is small: less than 0.1. In large and connected MCSs, however, cloudiness anomalies are larger and extend higher in the TTL. This result is in agreement with Rossow and Pearl (2007), who reported an increasing likelihood of convective intrusions of the TTL with increasing cloud system size and that such intrusion was observed in almost all systems with radii > 500 km. We also observe enhanced cloud occurrence to distances of 600 km and

**Fig. 4.** CFADs of CloudSat reflectivities in (a),(d) small SMCSs, (b),(e) large SMCSs, and (c),(f) CMCSs, using profiles sampling (top) the precipitating portions of the MCSs and (bottom) the anvil portions. Each CFAD is normalized such that its maximum value is 1.
b. Ice

Ice associated with MCSs can be derived from MLS observations, and IWC anomalies calculated as described in section 2c are shown in Fig. 7. IWC anomalies for the subset of levels above TTL base are shown in the right column. The largest ice concentrations and anomalies in these cross sections are in the upper troposphere near the MCS convective cores and extend upward into the TTL. Large and connected MCSs inject larger quantities of ice into the upper troposphere and TTL than small SMCSs. IWC decreases with distance from the MCS core, indicating the removal of larger ice particles via sedimentation and/or sublimation in the divergent airflow associated with the stratiform and anvil regions. While peak enhanced cloudiness in the MCS outflow lies at or above TTL base, as illustrated by the overlaid cloud fraction anomalies, the largest IWC values are at the bottom of the cross section. This is partly a result of the relatively coarse vertical resolution of the IWC data. In addition, CALIPSO’s sensitivity enables it to detect cirrus layers with IWC too small to be detected by MLS.

c. Water vapor

As shown in Fig. 2, water vapor near TTL base is more broadly distributed than ice. The strong decrease of water vapor concentrations with altitude in the TTL dominates cross-sectional composites of water vapor relative to MCS location (not shown). It is more informative to present the composites as anomalies from the climatological monthly mean (Fig. 8). Each MCS type is associated with enhanced water vapor in the upper troposphere. The anomalies decrease with distance from the MCS but extend radially outward beyond the area of substantially enhanced cloudiness or IWC (cf. Fig. 7). As noted above, the anomalous upper-tropospheric divergence also extends beyond the cloudy region, particularly in large and connected MCSs. The true strength and extent of the MCS outflow can only be estimated from the reanalysis winds. Because A-Train orbits repeat only every 16 days, we also cannot determine whether premoistening of the upper troposphere occurred prior to MCS formation. However, the results in Fig. 8 indicate that water vapor spreads laterally from the MCSs and into the surrounding cloud-free region prior to MCS dissipation. Previous studies have shown that diverging cirrus can persist for up to 2 days after convective dissipation and...
be advected up to 1000 km and that elevated upper-tropospheric water vapor concentrations persist longer than the cirrus (Luo and Rossow 2004).

Water vapor values in the TTL are too small to be seen in the left column of Fig. 8, which is dominated by the moister upper troposphere. Anomalies for the subset of altitudes in the TTL (above ~150 hPa) are shown in the right column of Fig. 8. Contrasting characteristics are observed near the MCS centers (within a radius of about 200 km) and at distances greater than about 400 km. Near the convective core, water vapor concentrations are below the climatological mean in the lowermost TTL, particularly in large and connected MCSs. This altitude can be above, at, or below the cloud top, as indicated by the overlaid cloud fraction anomalies and the discussion in section 5a. Chae et al. (2011) noted rapidly decreasing moistening within clouds above about 14 km and weak dehydration just below cloud-top level when the cloud extended above 16 km, as the environmental relative humidity became supersaturated. They also noted strong dry anomalies in the lower TTL when those altitudes lay just above the convective cloud tops, where subsidence occurs in response to strong divergence in the cloud outflow. They point out that these water vapor patterns are observed with any type of cloud in the TTL: deep convective, convectively generated cirriform, or in situ cirriform. It is therefore reasonable that the processes they describe should be associated with the MCSs discussed herein, thus producing the negative water vapor anomalies in the lower TTL in Fig. 8. The cirriform upper-level clouds of MCSs and their anvils are produced by the combination of convective and mesoscale dynamics that characterize MCSs (Houze 2004, 2014, chapter 9). We further observe weak moistening in the upper TTL above the MCS centers, consistent with the observation of Corti et al. (2008) that the deep clouds that extend upward beyond the CPT leave behind ice particles which sublime and moisten the environment.

Outside the convective core, at distances of about 400 km from the MCS center and extending outward, enhanced water vapor concentrations are observed in the lower TTL as well as the upper troposphere (Fig. 8), indicating that outflow from the MCSs and evaporative processes are moistening the environment. The water
vapor anomalies in this region decrease more rapidly with height than the environmental water vapor (cf. with Fig. 3), consistent with decreased convective influence and less effective moistening in the TTL. The moistening appears to extend upward to ~120 hPa; however, observations with finer vertical resolution would be needed to examine the precise vertical extent of the moistening. As in the upper troposphere, enhanced water vapor in the TTL extends laterally into cloud-free regions. Weak dry anomalies are observed near the CPT, near the tops of the highest clouds (cf. with Fig. 6).

The water vapor anomaly pattern near small SMCSs exhibits different characteristics than that for large and connected MCSs. Primarily, it is less robust, indicating that it is the largest MCSs that are important to the moistening of these upper levels. Positive anomalies associated with the small SMCSs are observed through most of the TTL, with peak values at 121 hPa. Laterally, the anomalies extend beyond the narrow core of enhanced cloudiness and several hundred kilometers into the region of suppressed cloudiness (i.e., beyond the red line in Fig. 8b). As previously noted, the radial outflow associated with small SMCSs is weak; hence, it is unrealistic to interpret the anomaly pattern in Fig. 8b as moistening of a broad region by the smallest MCSs. Rather, it appears that small SMCSs tend to form in large-scale environments with a slightly moistened TTL.

6. Conclusions

In this study, the vertical structure and microphysical composition of MCSs, as well as their role in detraining ice particles and water vapor in the TTL, have been investigated using a suite of satellite and reanalysis observations. We have examined three types of MCSs, as identified in observations from MODIS and AMSR-E (YH10): small and large separated MCSs and connected systems mostly likely formed via the merging of multiple MCSs. MCSs are ubiquitous in the convective regions of the tropics (Fig. 1), including East Asia, the western Pacific warm pool, and Central and South America, which have been identified as major source regions for air parcels eventually entering the stratosphere (Schoeberl et al. 2013). SMCSs occur more frequently over the tropical landmasses, while the merging of longer-lived systems to form CMCSs is more common in the oceanic ITCZs and over the western Pacific warm pool (Fig. 1).
CloudSat reflectivity distributions in the precipitating regions of active MCSs indicate high reflectivities in the midtroposphere, indicative of large ice and graupel particles forming in the convective updrafts and evolving into stratiform precipitation as the updrafts weaken or being detrained into neighboring stratiform regions of the MCSs (Fig. 4). Reflectivities in the MCS anvils are generally lower and decrease rapidly with altitude in the upper troposphere, consistent with sedimentation and aggregation within the anvils and the removal of larger graupel prior to air entering the anvil, as previously shown by Cetrone and Houze (2009) and YHH11. Small SMCSs exhibit characteristics of intense, young convection and are consequently more likely than large and connected MCSs to contain large ice and graupel particles in the upper troposphere (Fig. 5). Larger MCSs exhibit narrower, more strongly tilted reflectivity distributions in the upper troposphere, consistent with the “size sorting” processes characteristic of mature anvils (McFarquhar and Heymsfield 1996).

Composites of cloud occurrence observed by CALIPSO near MCSs (Fig. 6) reveal that cloudiness anomalies in small SMCSs are primarily vertically oriented and confined near the MCS cores, while large and connected MCSs exhibit not only broader core regions but also a prominent outward bulge in cloud occurrence near TTL base associated with anvil cirrus. Most MCSs are associated with clouds in the TTL: 74% of CALIPSO profiles within 100 km of the center of an MCS exhibited a cloud top above TTL base (about 150 hPa; Fig. 6). Large and connected MCSs are associated with more frequent cloud occurrence in the TTL than small SMCSs, in agreement with the inference of Rossow and Pearl (2007) that the likelihood of cloud penetrating upward into high levels increases with system size. Rossow and Pearl (2007) and Takahashi and Luo (2014) both report that penetration of the TTL tends to occur in the growing stage of the cloud system, although the latter noted significant penetration during the mature stage. Our results associate the most extensive cloudiness in the TTL specifically with large MCSs, in which anvils and overlying cirrus occur in greater quantities. “Overshooting” convective elements are actually less likely in these more widespread MCSs; for example, Virts and Houze (2015) showed that the larger MCSs have less vigorous convective elements, as indicated by lightning occurrence. It is the combination of convective and stratiform dynamics of MCSs that make them robust in affecting cloudiness in the TTL.
MCSs inject ice into the upper troposphere and TTL, as shown by composites of anomalous MLS IWC with respect to MCS location (Fig. 7). The largest ice quantities are near the centers of the MCSs, but elevated ice concentrations also extend laterally into the stratiform and anvil regions. Large and connected MCSs contribute the greatest quantities of ice to the TTL. These systems contain predominately smaller ice crystals and less graupel than the small SMCSs (Fig. 5). The less rapid sedimentation rates of the smaller particles allow them to remain aloft in the stratiform and anvil regions. We note that the IWC anomalies decrease nearly monotonically with height in the upper troposphere and TTL, while the cloud occurrence exhibits an outward bulge associated with the spreading anvils. This difference is largely due to the coarser vertical resolution of the IWC data (approximately 4 km in the TTL; Wu et al. 2008) but is also consistent with lower ice concentrations in the higher cirrus layers observed by CALIPSO.

Composites based on MLS water vapor mixing ratios (Fig. 8) demonstrate that MCSs moisten the upper troposphere. Moreover, stronger moistening is associated with the large and connected MCSs. In the TTL, a complex structure is observed that is consistent with previous studies of dehydration and hydration at upper levels. Near the MCS cores, water vapor concentrations are suppressed near TTL base. Higher relative humidity in the TTL (Fig. 3) reduces the possible amount of moistening, and subsidence above cloud top has been observed to produce strong drying in the lower TTL, as shown by Chae et al. (2011). Weak moistening is observed in the upper TTL, near the CPT, as the deepest penetrating convection leaves behind small ice crystals that sublimate and moisten the environment. A reversed pattern is observed from approximately 400 km outward, where the spreading anvils of large and connected MCSs moisten the lower TTL, and weak drying is observed at the CPT. The water vapor and radial wind anomalies in the upper troposphere and lower TTL extend into the cloud-free areas, indicating that the MCS outflow is moistening the environment. We also emphasize the possibility that a premoistened environment favors the development of mature MCS stratiform and anvil regions, but this cannot be investigated here owing to the limitations of satellite data.

Our conclusion that large and connected MCSs more strongly moisten the upper troposphere and lower TTL than small SMCSs contrasts with the notion that moistening of the TTL is primarily caused by isolated overshooting convective cells. The area covered by overshooting convective cells is many orders of magnitude smaller than the area occupied by MCSs. Our results indicate that the largest, mature MCSs are of greater importance than isolated overshooting cells because they are large enough in horizontal scale to be highly impactful on the water vapor field. Quantification of the comparative contributions of convection of various sizes and organization to the total water vapor content of the TTL and lower stratosphere is needed but is not straightforward given the water vapor data currently available.

MCS evolution cannot be examined using A-Train satellite observations, which provide only snapshots in time. However, the results presented in this paper suggest that small SMCSs are primarily young, developing systems, while large and connected MCSs represent the mature stage of the MCS life cycle (Houze 2004, 2014, chapter 9). Small SMCSs contain vigorous convection that produces large ice particles, especially graupel. These systems have limited stratiform and anvil regions, but the anvils that do exist contain large particles detrained from the convective core. Water vapor also detrains from the convective region, moistening the upper troposphere, while reduced moistening in convection penetrating the TTL combined with subsidence above cloud top produces negative water vapor anomalies near TTL base. As MCSs mature (and in some cases merge), their convection becomes less vigorous, resulting in reduced ice and graupel size. This ice is incorporated into the extensive stratiform and anvil regions (either by detrainment or by cells weakening and becoming stratiform), where it undergoes sedimentation and aggregation, such that smaller ice particles dominate near cloud top. At this stage, MCSs are characterized by extensive upper-level divergence. Anvils extend radially outward via this divergence, with peak cloud occurrence near TTL base shifting to slightly higher altitudes with distance from the MCS core. The MCSs thus moisten the upper troposphere via sublimation, and water vapor is detrained laterally into the cloud-free air. At larger distances from the convective core, where the above-cloud subsidence is weaker, this moistening extends into the TTL.

Our discussion has focused on contrasts between small SMCSs and large and connected MCSs. More subtle differences exist between large SMCSs and CMCSs. As shown in Fig. 1, large SMCSs occur more often over or near land, while most CMCSs are observed over the oceans. Thus, in a broad sense, differences between these two MCS categories represent the differences between the largest MCSs over land and oceans. Examining Figs. 4 and 6–8, it can be seen that compared to large SMCSs, CMCSs are slightly larger (Yuan and Houze 2013), exhibit somewhat stronger outflow in the upper troposphere, and are associated with larger ice and water vapor anomalies. However,
these differences are slight, and our results indicate that mature MCSs, whether they form over land or ocean, are major contributors to water vapor in the upper troposphere and TTL.

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