Observations of Temperature in the Upper Troposphere and Lower Stratosphere of Tropical Weather Disturbances

CHRISTOPHER A. DAVIS, DAVID A. AHIJEVYCH, AND JULIE A. HAGGERTY

National Center for Atmospheric Research,* Boulder, Colorado

MICHAEL J. MAHONEY

Jet Propulsion Laboratory, California Institute of Technology, Pasadena, California

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ABSTRACT

Microwave temperature profiler (MTP) data are analyzed to document temperature signatures in the upper troposphere and lower stratosphere that accompany Atlantic tropical weather disturbances. The MTP was deployed on the National Science Foundation–National Center for Atmospheric Research Gulfstream V (GV) aircraft during the Pre-Depression Investigation of Cloud-Systems in the Tropics (PREDICT) in August and September 2010.

Temporal variations in cold-point temperature compared with infrared cloud-top temperature reveal that organized deep convection penetrated to near or beyond the cold point for each of the four disturbances that developed into a tropical cyclone. Relative to the lower-tropospheric circulation center, MTP and dropsonde data confirmed a stronger negative radial gradient of temperature in the upper troposphere (10–13 km) of developing disturbances prior to genesis compared with nondeveloping disturbances. The MTP data revealed a somewhat higher and shallower area of relative warmth near the center when compared with dropsonde data. MTP profiles through anvil cloud depicted cooling near 15 km and warming in the lower stratosphere near the time of maximum coverage of anvil clouds shortly after sunrise. Warming occurred through a deep layer of the upper troposphere toward local noon, presumably associated with radiative heating in cloud. The temperature signatures of anvil cloud above 10-km altitude contributed to the radial gradient of temperature because of the clustering of deep convection near the center of circulation. However, it is concluded that these signatures may be more a result of properties of convection than a direct distinguishing factor of genesis.

1. Introduction

Observations of temperature, at high temporal and spatial resolution, are important to resolve a variety of phenomena in the tropical upper troposphere and lower stratosphere (UTLS) region, including gravity waves, tides, perturbations from moist convection, and the spatial and temporal variations of the tropopause due to tropical waves. Temperature data in this region have been obtained by a variety of remote sensing platforms in recent years. These observations include global positioning system (GPS) radio occultations (Anthes et al. 2000; Lin et al. 2010) from which temperature soundings are retrieved. Such measurements are important for documenting climatological seasonal and intraseasonal conditions (Kim and Son 2012) and for sampling profiles through clouds (Biondi et al. 2012). Radio occultation data have also been used to improve temperatures retrieved from the Atmospheric Infrared Sounder (AIRS) aboard the Aqua satellite (Borbas et al. 2008). Furthermore, high-vertical-resolution and high-time-resolution data have been collected using radio acoustic sounding techniques (Alexander and Tsuda 2008).

A limitation of UTLS observations of temperature pertains to tropical subsynoptic-scale weather systems featuring deep, moist convection. While the UTLS region in the tropics, or tropical tropopause layer (TTL), has been examined extensively on larger time and spatial scales (Fueglistaler et al. 2009), the mesoscale dynamics of the TTL are relatively unexplored. Measurements

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Corresponding author address: Christopher A. Davis, NCAR, P.O. Box 3000, Boulder, CO 80307.

E-mail: cdavis@ucar.edu

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following tropical weather disturbances are important for understanding how the TTL evolves on time scales from hours to 1–2 days in the presence of deep, moist convection and how processes in this region leave a signature in the spatial variation of temperature. The combination of a high-altitude research aircraft targeting specific atmospheric circulation features and a microwave temperature profiler (MTP) allows us to depict spatial and temporal aspects of temperature in the TTL that have not been studied in detail before. The purpose of the present paper is to provide insight into the thermodynamic processes in tropical disturbances in the TTL region.

An important feature of the TTL is the cold point—the altitude that often marks a pronounced change in lapse rate and defines the absolute upper bound of the troposphere. By virtue of being the coldest temperature in the profile, the cold point strongly constrains the amount of water vapor that can cross into the lower stratosphere [see the review by Randel and Jensen (2013)]. Variations in the cold-point temperature often indicate the presence of vertical motion that affects the exchange of trace constituents between the troposphere and stratosphere (Holton et al. 1995). On annual and interannual scales, the upward branch of the Brewer–Dobson circulation exerts significant control on tropical cold-point temperature. Intraseasonal oscillations of various types can modulate the cold point (Kiladis et al. 2009) and Kelvin waves are perhaps the most significant waves in the tropics from this perspective. Modulations of the cold-point temperature influence the maximum potential intensity of tropical cyclones and tropical cyclone activity (Emanuel et al. 2013). On short time scales, deep convection can modulate the cold-point temperature (Danielsen 1993; Teitelbaum et al. 2000).

The upper-tropospheric thermodynamic signatures that accompany organized deep moist convection are also related to tropical cyclone formation and mature tropical cyclones. It has been argued from numerical calculations that anvils in tropical cyclones are expanded because of radiative heating in clouds, and destabilized because of longwave cooling at cloud top above the radiative heating (Bu et al. 2014). Warming in the upper troposphere has also been shown to reduce the buoyancy of convecting air parcels and induce a lower altitude of detrainment (Bister and Mapes 2004; Raymond and Sessions 2007; Komaromi 2013; Zawislak and Zipser 2013). This, combined with tropospheric moistening, is believed to induce a vertical mass flux that maximizes at lower altitudes and is more conducive for surface cyclone intensification.

The MTP was deployed on the National Science Foundation (NSF)–National Center for Atmospheric Research (NCAR) Gulfstream V (GV) aircraft during the Pre-Depression Investigation of Cloud-Systems in the Tropics (PREDICT) in 2010 (Montgomery et al. 2012). The PREDICT project focused on westward-moving, synoptic-scale, tropical weather systems over the Atlantic during August and September 2010. Sometimes generically referred to as tropical waves, these disturbances lasted many days individually and hence could be sampled many times. This allowed the examination of changes in clouds, wind, and thermodynamic quantities that have helped elucidate the process of tropical cyclone formation. Because the GV typically flew near 14-km altitude, the vertical passive scanning of the MTP allowed us to observe a deep layer of the atmosphere (roughly 6–20 km) with profiles obtained every 17 s over time spans of roughly 6 h during each mission.

The paper first examines the behavior of the cold-point temperature as measured by the MTP and how its variations may be related to mesoscale convection and tropical waves. The paper then examines spatial structures in the temperature field, emphasizing differences between disturbances that develop into tropical cyclones from those that do not. Finally, time dependence of temperature in the diurnal cycle is considered wherein we focus on contrasts between temperature profiles through anvil clouds and those not through anvils to gain insight about diabatic heating and its relation to mesoscale processes.

2. The MTP, data, and analysis method

a. Sensor description

The PREDICT airborne payload included a newly designed MTP, built for the NCAR GV at the Jet Propulsion Laboratory (JPL) (Lim et al. 2013). The MTP is a microwave radiometer that measures thermal emission at spectral lines where oxygen absorption and emission dominate over water vapor and liquid water absorption and emission. The MTP normally operates at three frequencies: 55.51, 56.65, and 58.80 GHz. The MTP scans from near-zenith to near-nadir angles, measuring brightness temperatures that are used to generate vertical temperature profiles along the flight track.

The NCAR MTP is based on previous MTPs built and operated by JPL for many research aircraft (Denning et al. 1989; Gary 2006) but includes a number of design improvements that take advantage of current technology. Major components of the NCAR MTP are housed within a standard 6-in. (15.2 cm) canister that mounts on an aircraft wing strut. A fiberglass fairing attached to the front of the canister contains a microwave-transparent
window. A forward-looking scan mirror within the 
fairing rotates 360°, sampling emission from angles be-
tween ±80° from horizontal. During each rotation, the 
mirror pauses to sample at 10 specified elevation angles 
followed by measurement of a heated reference target to 
calibrate the signal gain. The signal received by the scan 
mirror is directed to a radiometer and sent to a controller 
board that communicates with the MTP cabin-mounted 
computer. The controller board also receives instructions 
from the cabin-mounted computer.

Specifications for the NCAR MTP are listed in Table 1. 
Because of improvements in hardware technology, each 
sampling frequency can now be tuned to the center 
frequency of an oxygen absorption line (previous MTPs 
were tuned to frequencies between spectral lines because 
of hardware performance limitations). This capability 
results in improved performance, with flight-level tem-
perature errors on the order of ±0.2°–0.3°C (Lim et al. 
2013). The scan mirror completes a 360° rotation every 
17 s, yielding derived temperature profiles at approxi-
mately 4-km horizontal spacing for typical GV air speeds 
of 230 m s⁻¹.

b. Temperature profile retrieval and quality control

Vertical profiles of temperature are obtained from 
brightness temperature measurements using a statistical 
retrieval process with Bayesian aspects designed for the 
airborne environment (Nielsen-Gammon et al. 2008). A 
historical archive of radiosonde profiles from stations in 
the vicinity of the flight provides a priori information 
needed in the retrieval scheme. Forward radiative 
transfer calculations use radiosonde profiles to estimate 
expected brightness temperatures. Linear, multivariate 
regression relates calculated brightness temperatures to 
physical temperatures for each radiosonde profile. The 
regression process provides a set of coefficients at 
specified altitudes that can be applied to convert MTP-
measured brightness temperatures to physical tem-
perature profiles. For a given MTP scan, measured 
brightness temperatures are compared with all avail-
able calculated brightness temperature profiles. An 
information-theory-based metric is used to select the 
most similar calculated brightness temperature profile 
(from radiosondes); the retrieval coefficients associated 
with that radiosonde profile are applied to the measured 
brightness temperatures of the given scan to obtain a 
temperature profile.

The MTP operated on all but one PREDICT flights, 
typically providing temperature profiles from altitudes 
of about 6–20 km. Occasional, short-duration data gaps 
resulted from malfunction of the frequency synthesizer 
because of encounters with high levels of ice water 
content (this behavior has since been corrected). In 
addition to temperature profiles, the dataset includes 
tropopause heights estimated using the World Meteo-
rological Organization (WMO) lapse-rate definition.¹

Prior to performing meteorological analysis, we fil-
tered out a small amount of additional data by requiring 
the aircraft to be above 12 km (thereby removing the 
ascent and decent portions of missions) and using the 
quality-control metric meridional regional index (MRI).² 
MRI is a unitless measure from 0 to 2 that increases 
during rapid ascent or descent and while making turns. It 
also reflects the degree of mismatch between the ob-
served radiance profile and the expected radiance pro-
files derived from a library of temperature soundings. 
Values less than 1 are generally considered good, but we 
chose a unique cutoff value for each mission based on 
histograms of MRI. The frequency distribution of MRI 
usually peaked near an MRI of 0.1 and steadily de-
creased to 0. To eliminate data of questionable validity, 
we binned MRI values in increments 0.01, found the first 
bin beyond the peak with not more than one MRI value, 
and assigned that as the cutoff value (Fig. 1). All data 
with MRI values higher than the cutoff value were dis-
carded. The example shown contains more than 1000 MRI 
values and is representative of all missions. The first bin 
in which there are one or fewer counts is MRI = 0.29. 
There are eight profiles with greater MRI values, and 
these are ignored. Considering all missions, the cutoff 
MRI value varied from 0.18 to 0.56. While this quality-
control technique might seem to remove usable data, 
it removes far less data than filtering based on the

¹The lapse-rate tropopause is defined as the level above which 
the lapse rate remains less than 2°C km⁻¹ over a vertical distance of 
at least 2 km.

²The meridional regional index formerly referred to the quality 
of blended tropical and midlatitude retrievals but has since come to 
signify a more general measure of the quality of the retrieval.
number of channels contributing to the radiance measurement.

The spatial locations of all MTP observations passing the quality-control test appear in Fig. 2. These data are confined to the tropical Atlantic Ocean and the Caribbean Sea, covering a narrow latitude range and a 50° range of longitude. In all, eight distinct tropical weather systems were examined, four of which developed subsequently into Tropical Cyclones Fiona, Karl, Matthew, and Nicole. A notable nondeveloping system was Gaston. It was classified as nondeveloping because it failed to reintensify after weakening below minimum tropical storm intensity. Summaries of all cases are available from the tropical cyclone reports from the National Hurricane Center (http://www.nhc.noaa.gov/data/) and from Beven and Blake (2014). A total of 24 missions flown with the MTP are analyzed herein, beginning with 17 August and ending with 30 September 2010 (Table 2). A mission was defined as a flight into a disturbance. The GV actually flew 26 times during PREDICT. We excluded three flights: an initial test flight, a flight with a dropsonde malfunction (23 August), and a flight where the MTP was not on the GV (7 September). The flight on 30 September was counted as two missions because two distinct circulations were sampled. More details about the GV missions may be found in Montgomery et al. (2012).

With the MTP data that passed quality control, we proceeded with a linear vertical interpolation to constant pressure-altitude surfaces. While the vertical resolution of the MTP varies from roughly 150 m near the altitude of the GV to more than 1 km at ≥6 km from the aircraft, we chose to vertically interpolate the MTP data to a uniform pressure-altitude spacing of 250 m. To remove seasonal and regional-scale variations in the data, a temperature anomaly (also referred to as a deviation) was defined as the departure of temperature $T'$ from the mission average at a given pressure altitude. This allows comparison of mesoscale temperature structures between weather systems widely separated in space and time.

Approximately 500 dropsondes were deployed from the Airborne Vertical Atmospheric Profiling System (AVAPS; more on the AVAPS system can be found at https://www.eol.ucar.edu/isf/facilities/dropsonde/gpsDropsonde.html ) aboard the GV. Temperature data from these GPS dropsondes were collected every 0.5 s, or at a vertical spacing of approximately 10 m. Dropsondes near the MTP profiles could also have been employed to provide retrieval information a priori, albeit only below the aircraft rather than above and below. However, the dropsonde profiles were not used in retrievals so as to preserve them as an independent dataset for validation of derived MTP profiles.

We also constructed a temperature anomaly dataset for the dropsonde data following the same procedure as for the MTP data. For each MTP profile that survived the quality-control procedure, a search was made for the closest dropsonde profile based on the release point of each drop. A temporal displacement limit of 20 min was imposed, but typically the displacement was less than 1 min. This procedure resulted in a given dropsonde being included many times because there were about 30–40 times as many MTP profiles as dropsondes in the raw data. However, the temporal spacing between dropsondes was nearly constant (10–12 min), so each dropsonde was represented nearly equally. The height levels for dropsonde data (0–13 km) were lower than for the MTP (6–20 km). The resulting dropsonde dataset, which was matched one to one with MTP profiles and with 250-m
vertical spacing in pressure altitude, was used for comparison of spatial structures of temperature anomalies with the MTP data (section 3b).

A direct comparison was performed of nearly 500 PREDICT dropsonde profiles with the portion of MTP profiles below the aircraft. For this comparison only, the profile of full temperature from the dropsonde profile nearest each MTP profile was linearly interpolated to the pressure altitudes of the MTP profile and the differences were grouped in 500-m bins (Fig. 3). In the upper troposphere, MTP temperatures are more than 1.0°C greater than dropsonde temperatures above 12 km, and the systematic difference decreases downward (Fig. 3). In contrast, the standard deviation is fairly small near 12 km and increases down to 6 km. The latter effect may be expected, based on coarser vertical resolution of the original MTP data compared with the original dropsonde profiles and the greater spatial displacement between the dropsonde and MTP.

The difference between MTP and dropsonde temperatures in the upper troposphere is also consistent from mission to mission. The distribution of temperature differences obtained by averaging separately over each of the 24 missions reveals a mean of 0.8°C at 12-km pressure altitude with a standard deviation of only 0.4°C. The reason for the systematic difference between dropsondes and MTP at higher altitudes is not fully understood and is an area of ongoing research. One would expect the highest quality data from MTP near the aircraft altitude because its vertical resolution maximizes there and because the in situ temperature sensors were used to constrain the MTP profiles.

c. Spatial and temporal characteristics of MTP retrievals

As an example of the structures contained in the MTP data, Fig. 4 shows the time–height series of MTP temperature anomalies for the GV mission on 14 September 2010 during the formation of Tropical Cyclone Karl. Data gaps in Fig. 4 are partly the result of high ice water contents that caused several instruments on the GV to malfunction. Other gaps result from data that failed the quality-control algorithm. Horizontal variations of temperature are difficult to discern from the full temperatures (Fig. 4a). The perturbations, however, reveal a rich structure consisting of warm anomalies near the cold point between 16 and 17 km in regions above deep convection and mesoscale negative anomalies (Fig. 4b).

The center of Karl was overflown at about 1740 UTC when pronounced warming was detected in the 15–18-km-altitude range. These warm anomalies also coincide with the availability of only a single channel from the
MTP, although not all periods of single-channel availability in anvil cloud reveal this sort of temperature structure.

Another perspective of the data is afforded by a three-dimensional rendering of the temperature perturbations from a flight over ex–Tropical Storm Gaston on 5 September (Fig. 5). The square–spiral flight pattern is flown counterclockwise. Along the southernmost leg, roughly along 14.5°–15°N, temperatures in the middle and upper troposphere are 2°–3°C cooler than temperatures on the northernmost leg—around 19°N. This meridional temperature contrast is primarily the result of prior subsidence throughout the middle and upper troposphere. Supporting this interpretation is the widespread dry air to the north and east of ex-Gaston in the middle and upper troposphere (Davis and Ahijevych 2012). The positive temperature anomaly is particularly strong on the northern side of the vortex shortly after 1800 UTC (Fig. 6) when the GV made its closest approach to the center of ex-Gaston and to the organized convection near the center. At the time of the deep warm anomaly (indicated by the arrow in Fig. 6), the GV was flying through cirrus cloud. Dropsondes (not shown) also suggest that the anvil is warm relative to its surroundings.

To quantify spatial scales represented in the MTP data, we analyzed the temporal autocorrelation of the MTP temperature anomalies to investigate what equivalent length scales were represented. Using a drop in correlation of temperature anomalies to 1/e, we obtained a correlation length scale of approximately 80 km near altitudes of 12–13 km ranging to 40 km at altitudes of roughly 6–8 and 18–20 km. High-frequency fluctuations were present but mesoscale signals were clearly evident in the data, consistent with visual impressions from Figs. 4b and 6.

To obtain a quantitative spatial analysis of the MTP data, we followed the approach of Wang et al. (2012) by anchoring our horizontal coordinate to the average location of the center of the lower-tropospheric cyclonic circulation during each mission. This center actually represents an ensemble estimate, obtained using operational analyses from four different agencies,3 of the center of a recirculation region defined as the intersection of the critical line (where the zonal wind component and disturbance translation speed were equal) and the axis of the wave trough (Dunkerton et al. 2009; Montgomery et al. 2012). Although this intersection of critical line and wave trough can be identified at multiple levels, the 700-hPa level (3-km altitude) was nearly always the level chosen to define the disturbance center. Importantly, the center of circulation was defined in a frame of reference moving with the system. For each mission, data were time–space corrected to the location of the circulation center. The location of the circulation center and its estimated motion vector are listed in Table 2 for each mission. Also following Davis and Ahijevych (2013, hereafter DA13), we grouped missions according to the stage of development on the path to becoming a tropical cyclone (Table 2). There were 9 missions into disturbances that never developed and 10 missions into pregenesis disturbances. Five of those 10 missions were flown into a disturbance that developed less than 2 days later. Nondeveloping disturbances are denoted “ND” and disturbances within 2 days of development are denoted “L2.”

3 The agencies were the National Centers for Environmental Prediction (NCEP), the European Centre for Medium Range Weather Forecasts (ECMWF), the Met Office, and the Fleet Numerical Meteorology and Oceanography Center (FNMOC).

3. Results

a. Cold-point temperature

The MTP provides an estimate of the cold-point temperature in each profile—that is, approximately every 17 s. The altitude of the cold point is consistently near 16 km. In agreement with Zhou and Holton (2002), the altitude fluctuations of the cold point have no obvious correlation with the cold-point temperature itself. The
cold-point altitude is only about 2 km above the aircraft; hence, the vertical resolution of the MTP is roughly 300 m. In the present analysis, all cold-point values within a single mission are averaged together to produce a mission-average cold-point temperature (Fig. 7). The apparent decrease of cold-point temperature with time (Fig. 7a) is attributed both to the seasonal cycle and to the greater number of flights into developing tropical cyclones (or their precursors) during the last 3 weeks of September. Karl developed on 14 September, Matthew on 23 September, and Nicole on 28 September. Karl, which became a major hurricane, formed beneath the lowest cold-point temperature observed during PREDICT.

Cold-point temperature also decreases westward across the tropical Atlantic (Fig. 7b). This is a well-known feature in this region (Kim and Son 2012). Of note, the early season disturbance that failed to develop as it passed through the Caribbean (cf. Figs. 7a and 7b) had cold-point temperatures roughly 5°C warmer than those for Karl and Matthew. This is perhaps a reflection of the climatological basin-scale circulation changes that favor the shift of tropical cyclone formation from the eastern Atlantic to the Caribbean in late September (National Hurricane Center; displayed at http://www.nhc.noaa.gov/climo/). However, we note that it is not possible to cleanly separate spatial and temporal variations of cold-point temperature across the Atlantic basin given the MTP observations alone.

There is a suggested relationship between elevations of the cold point and the coldest cloud tops associated with deep, widespread convection that, in turn, often accompany developing cyclones (Houze 2010). Figure 7c shows the 1% value of the distribution of infrared cloud-top temperature occurring within 500 km of the circulation center during the times that the GV was flying (generally from sunrise to early afternoon local time).
There are roughly $10^5$ pixels during each mission that represent the coldest 1% of clouds, equivalent to an area of roughly $10^5 \text{ km}^2$ in each half-hourly satellite analysis. The figure shows that for temperatures below approximately $-70^\circ\text{C}$, the cold-point temperature is modestly correlated with the coldest cloud-top temperatures of the distribution. One might expect a correlation only for IR temperatures sufficiently cold that the clouds were close to the cold-point altitude. In support of this result, Hamada and Nishi (2010) found that an IR temperature...
from −70° to −73°C corresponded to a cloud-top height of 16 km, which is approximately the average cold-point height in our data. Thus, we infer that the deepest clouds penetrate to the cold point, so there is the likelihood of a direct influence of convection on cold-point temperatures by a transient elevation of the tropopause through mesoscale ascent and some influence of radiative cooling. This result is consistent with the direct influence of convection on the cold point that was clearly shown by Danielsen (1993).

However, there is also a well-known influence of tropical waves on cold-point temperature. Wheeler et al. (2000), Randel et al. (2003), Zhou and Holton (2002), and Kiladis et al. (2009) emphasize how equatorial Kelvin waves modulate the TTL. In particular, it is known that the cold-point temperature is reduced in advance of the lowest outgoing longwave radiation (OLR) values within a Kelvin wave. This arises because of tilted structures in the temperature field in the lower stratosphere. Griffin (2012) documented the presence of a Kelvin wave packet that entered the Caribbean from the eastern Pacific and created synoptic-scale conditions favorable for the formation of the precursor vortex to hurricane Karl. The lowest cold-point temperature, however, occurred in advance of a second OLR minimum in the Kelvin wave frequency band (see Griffin 2012, his Fig. 4.1). In the composite vertical structure of Kelvin waves (Ventrice et al. 2012), the minimum cold-point temperature leads the OLR minimum by 1–2 days, which, for the observed rate of propagation (15 m s⁻¹), is consistent with the minimum cold-point temperature being collocated with the formation of Karl on 14 September. Thus, the passage of a Kelvin wave also appeared to reduce the cold-point temperature around Karl near the time of its development.

It appears that both the ascent within Kelvin waves and deep convection within developing tropical cyclones affect the cold-point temperature. The MTP data alone are not sufficient to discern the relative contributions of these phenomena. However, it is clear that the result of Danielsen (1993) was not an isolated incident. Deep convection in tropical cyclones clearly approaches, and therefore influences, the cold point.

b. Spatial variation of temperature anomalies

Composite profiles of temperature anomalies were computed to highlight systematic radial temperature variations in developing and nondeveloping disturbances. Profiles were constructed by averaging temperature anomalies within two radial rings: 0–200 and 400–500 km. A similar approach was used in DA13, although the precise definitions of temperature anomaly and radial averaging were somewhat different. By following identical averaging procedures for MTP and dropsonde temperature anomalies, we may also compare the two datasets in the range of pressure altitude where they overlap. In addition to subdividing composites according to radius and data source, we consider separately ND and L2 subsets as defined in section 2 and Table 2.

To assess the robustness of the anomaly profiles, confidence intervals were computed as in DA13. The 95% confidence level was chosen and the underlying distribution of temperature anomalies was assumed to be normal. Of particular concern in this calculation is the number of degrees of freedom in the data. Based on an e-folding length scale of 80 km, we assume independence over two
$e$-folding lengths or 160 km. This corresponds to approximately 36 MTP profiles. A typical number of MTP profiles in both the 0–200- and 400–500-km radial bins ranges from about 700 in the L2 sample to 1400 in the ND sample. Hence, the number of degrees of freedom varies from 20 to 40. Dropsondes, spaced at roughly 130 km, are assumed independent as in DA13. The number of dropsondes varies from 25 to 45 in each sample.

The profiles of temperature anomalies appear in Fig. 8. We emphasize the statistically meaningful radial variations,
which are evident where the confidence intervals (yellow and green ribbons) do not overlap. All profiles indicate a negative radial gradient of temperature in the upper troposphere, but at a greater pressure altitude in the MTP data. The radial variation in the L2 dropsonde composite is deeper and stronger than in the corresponding MTP data (Figs. 8b,d). In the ND composite, the MTP also indicates a negative radial temperature gradient as high as 15 km (Fig. 8a). No statistically meaningful radial gradients are evident in the lower stratosphere (above 16 km).

We also examined whether the largest radial gradients of temperature were consistent from case to case. A negative radial gradient was present at 12 km in each of the five L2 cases in the MTP data. Among ND cases, two cases lacked MTP data in either the 0–200- or 400–500-km rings. Of the seven remaining ND cases, five exhibited negative radial gradients. A similarly consistent radial variation of temperature from case to case was also seen in the dropsonde data at 10-km pressure altitude. For the L2 composite, the negative radial gradient of temperature was seen in each case between 8 and 12 km as well. We conclude that the gradients seen in Fig. 8 are representative of most cases and seem particularly consistent in the 2 days leading up to genesis.

Negative radial temperature gradients in the upper troposphere near 200 hPa (12 km) preceding tropical cyclone formation have been noted in observational studies by McBride (1981) and Bessho et al. (2010). These previous studies, using radiosonde and satellite retrievals [from the Advanced Microwave Sounding Unit (AMSU)], respectively, also indicate that gradient does not extend below about 8 km. Although more consistent with the MTP results, AMSU retrievals are not accurate in heavy rainfall in the lower and middle troposphere (Bessho et al. 2010). The results from McBride (1981) probably represent profiles more than 2 days before genesis in many cases. At that stage, dropsonde results presented by DA13 and Komaromi (2013) agree that temperature anomalies were confined to the upper troposphere. Finally, given that we do not expect subtle temperature variations to be well captured by the MTP far beneath the aircraft, it appears likely that, within 2 days of genesis, the negative radial temperature gradient extends to a lower altitude than captured by the MTP.

c. Profiles through clouds

Similar to the approach in DA13, we now consider composites through deep anvil clouds versus composites through areas relatively free of such clouds. Studies such as Sherwood and Wahrlich (1999), Sherwood et al. (2003), and Biondi et al. (2012) have shown vertical profiles through deep convection anvils that reveal significant temperature anomalies. Sherwood and Wahrlich used widely spaced radiosondes and hence were forced to composite over many cases, thus retaining only the broader time scales of variations. Biondi et al. (2012) used the concurrence of CloudSat cloud profiling radar (CPR) data and GPS radio occultations to derive soundings through convective systems. The use of CPR data allowed an accurate specification of the cloud-top height and, when combined with the high vertical resolution of the occultation soundings, detailed cloud-top-relative profiles could be derived. Results from Biondi et al. (2012) suggested a cold anomaly near cloud top, stronger for cloud tops below 14 km than for cloud tops above 14 km. The results of Sherwood and Wahrlich (1999) resemble those of McBride (1981) for developing tropical depressions over the Pacific Ocean. Both studies found a well-defined warm anomaly centered at about 200–250 hPa in deep convective systems where the cloud tops were probably above 14 km.

Unlike previous studies, our data contain a relatively large number of temperature profiles within individual weather systems—approximately 1000 per mission. From a cloud perspective, the sampling means that we have good statistics of “cloudy” and noncloudy profiles in the same synoptic-scale environment. A cloudy profile is defined as the presence of dense cloud, at the altitude of the aircraft, determined from the condensed water content measured by the counterflow virtual impactor (CVI; Gerber et al. 1998). When the CVI condensed water exceeded 0.07 g m$^{-3}$, the GV was deemed to be “in cloud.” This threshold represents a relatively high ice water content (IWC) for cirrus clouds (Heymsfield et al. 2005) and is consistent with our emphasis on optically thick anvils near active convection. To arrive at this particular value of ice water content, we maximized the correlation between IWC and the time series of an independent proxy for the aircraft being in cloud: the occurrence of IR cloud-top temperature, interpolated to the aircraft position, colder than the aircraft in situ temperature plus 4°C.

Statistics of temperature profiles within and outside of anvil clouds are summarized using contoured frequency by altitude diagrams (CFADs; Yuter and Houze 1995). These diagrams (Fig. 9) represent contours of the relative frequency of a particular variable (temperature anomaly derived from MTP) in a two-dimensional space, where one of the dimensions is the range of values of the variable and the other dimension is pressure altitude in the present context. The composite profiles are divided into “in cloud” and “out of cloud”
samples using data within 500 km of the circulation center.

The clearest signature in Fig. 9 is the existence of a negative temperature anomaly near 15-km altitude in the in-cloud composite (Fig. 9a). Recall that this altitude is about 1–1.5 km above the aircraft. A limitation of the present analysis is that we do not know precisely the cloud-top height above the aircraft. Flight operations dictated that the GV avoid cloud tops colder than about $-70^\circ$C because of the likely occurrence of strong turbulence in the deepest convection. The cloud-top altitude corresponding to an IR temperature of $-70^\circ$C is typically between 15.5 and 16 km (Hamada and Nishi 2010). Given that the GV was in cloud at 13.5–14 km and avoided the coldest cloud tops, it is reasonable to assume that cloud tops were typically near an altitude of 15 km. This places the cool anomaly (Fig. 9a) near the top of the cirrus canopy, consistent with previous studies.

Of note is the similarity of the upper-tropospheric warm anomaly to that shown in the vertical profile of anomalies within 200 km of tropical disturbances (Fig. 8). The negative temperature anomaly in Fig. 9a is manifested in Fig. 8c as a rapid return of the temperature anomaly to near zero around 15 km. The upper-tropospheric warm anomaly and near-tropopause cool anomaly reflect the imprint of the mesoscale signature of organized convection itself. Cold cloud associated with deep convection covers a greater fraction of the area within 200 km of the center in developing cases and hence leaves a correspondingly greater signature. Roughly 25% of the area is covered by cloud tops colder than $-60^\circ$C in developing cases versus 17% for nondeveloping cases. For IR temperatures of $-70^\circ$C or colder, the percentage of area covered was 10% for developing cases and only 4% for nondeveloping cases.

Also evident in Fig. 9a is a warm anomaly in the lower stratosphere around 17–18 km. This feature has no counterpart in Fig. 8. The warm anomaly in the lower stratosphere is apparent in several flights, not just the flight on 14 September (Fig. 4b). Removal of the 14 September mission from the local-time calculation does not significantly alter the composite warm anomaly.

To assess how profiles differ in cloud from out of cloud as a function of time in the diurnal cycle, we bin profiles according to the local time of day. When placed in hourly temporal bins beginning at 0600 LT (essentially sunrise), we obtain mean profiles of temperature anomaly that span the time range from sunrise to early afternoon. Subtracting the out-of-cloud profiles from the corresponding in-cloud profiles yields the sequence.
of profiles shown in Fig. 10. Following sunrise, the temperature anomalies generally amplify until about 0900 LT. Warming centered at 10–12 km and separately at 17–18 km and cooling maximizing at 15 km are consistent with the signatures in the CFADs. Magnitudes of temperature change range from approximately 5 K day$^{-1}$ below 12 km to as much as 10 K day$^{-1}$ in the layer from 13 to 15 km and in the lower stratosphere.

In a climatological context, the diurnal cycle of convection over the tropical Atlantic is relatively weak when measured from an Eulerian perspective compared to other regions of the tropics (Wylie and Woolf 2002; Taylor 2012). However, Tsakraklides and Evans (2003) were able to discern a well-defined diurnal cycle following individual organized convective systems, some of which were tropical cyclones or their precursors. They found that the most vigorous convective elements occurred between midnight and sunrise—a result also reported by Liu and Zipser (2005). The greatest vertical growth of convection occurred near or slightly before sunrise, whereas the greatest horizontal growth occurred closer to midafternoon.

In the present context, the diurnal cycle of clouds is represented by the fractional area covered by clouds with tops colder than a particular temperature. For the days on which MTP data were available, the diurnal time series derived from geostationary satellite data appear in Fig. 11. These time series, representing data from radii less than 500 km, show that cloud tops colder than $-70^\circ$C are most frequently observed between 0200 and 0300 LT. The coverage of warmer cloud tops occurs successively later through midafternoon. The morning hours (0600–1200 LT) feature a broad maximum in clouds with tops between $-60^\circ$ and $-70^\circ$C, but the spatial extent of the coldest cloud tops reaches a minimum during this time. The late night vigor of convection and the later spreading of lower cloud tops are generally consistent with previous studies of deep tropical maritime convection.

To understand the temperature changes observed by the MTP through anvils it is helpful to consider both the local radiative effects and the mesoscale vertical motions simultaneously. However, without detailed modeling, the following interpretation should be treated with caution. What is apparent is that the greatest temperature anomalies for in-cloud profiles develop when the area coverage of the coldest cloud tops is declining but the area of slightly warmer cloud tops is increasing. While this latter feature could result from detrainment from convective towers, it has long been recognized that mesoscale ascent accompanies expanding stratiform precipitation (Houze and Cheng 1981). The cool anomaly at 15 km may result partly from radiative cooling, but Sherwood et al. (2003) argue that radiative cooling is not particularly large at 15 km, and that adiabatic lifting is probably also playing a role in creating the cool anomaly. Holloway and Neelin (2007) also emphasize that adiabatic cooling from mesoscale upward motion produces what they refer to as the “convective cold top.” We have
also implicated upper-tropospheric mesoscale lifting in the behavior of cold-point temperatures (section 3a). As a final piece of evidence, Davis and Ahijevych (2012) demonstrated a clear diurnal cycle of mesoscale mid-tropospheric vertical motion prior to the developments of Karl and Matthew, with strong ascent occurring only in the morning. Thus, while radiative cooling is likely occurring, it is equally likely that mesoscale ascent strongly affects the negative temperature anomaly in the upper troposphere.

At lower altitudes (8–11 km), heating from longwave radiation (Liou 1986) or freezing of precipitation could be important depending on the height of the cloud base, and these could more than offset the adiabatic cooling in mesoscale updrafts to create the warming observed in the early morning (Fig. 10). The subsequent warming of profiles between 12 and 15 km—between 1000 and 1400 LT—probably represents shortwave absorption in cloud as the sun angle increases and mesoscale upward motion weakens as the area of cloud coverage reaches its peak.

The cause of the transient warming during midmorning in the lower stratosphere is not obvious (Fig. 10). It coincides in time with a decline in the area covered by the coldest cloud tops, which is representative of the climatological timing of the decrease of overshooting tops and convective vigor. Sherwood (2000) proposed a reverse vertical circulation in response to convective overshooting in the lower stratosphere. With a lag of several hours, the implied warming from this subsidence could explain the transient warming observed around midmorning. However, it is also possible that the warm anomaly is related to gravity waves excited by convection, although the exact mechanism remains unclear.

4. Conclusions

Observations of spatial and temporal aspects of temperature in the UTLS region from a microwave temperature profiler (MTP) have highlighted features that accompany synoptic-scale tropical weather systems. Revealing mesoscale structures is particularly challenging in the tropics where horizontal gradients of temperature do not systematically exceed 1°C over a few hundred kilometers. The intrinsic accuracy of the instrument is roughly 0.25°C and high-frequency fluctuations in the data have amplitudes closer to 0.5° and 1°C. Nonetheless, meaningful mesoscale information is obtained.

The organization of convection in tropical disturbances, with a preference for convection to congeal around the disturbance center (in a comoving frame of reference; Dunkerton et al. 2009), leaves a signature in mesoscale temperature anomalies. The most pronounced of these is a negative radial gradient of temperature that maximizes in the MTP data at an altitude of 12–13 km and in the dropsonde data near an altitude of 8–10 km. The magnitude of this gradient reaches 0.8–1.2 K (500 km)$^{-1}$. Dropsondes show that the radial gradient is stronger and deeper prior to genesis than for non-developing cases. The MTP data indicate a marginally more robust gradient in developing cases only near 12–13 km. It appears that more than about 3 km below the aircraft, the MTP may have difficulty identifying subtle mesoscale variations of temperature.

An important aspect of tropical systems and attendant deep convection is their role in modulating the temperature of the cold-point tropopause. It appears that for sufficiently vigorous convection, defined as having IR cloud-top temperatures below −70°C, mesoscale ascent augments the ascent in tropical Kelvin waves to lower the cold-point temperature. Penetration of organized deep convection near the cold point was found only for systems that had attained tropical depression status or greater. These systems maintained a greater coverage of deep convection near the circulation center. This implies more organized mesoscale ascent through the troposphere. Radiative cooling may also contribute to the cold-point temperature reduction, but consistent with previous studies, we suggest it is of secondary importance near 15-km altitude.

Aspects of the system-scale response in the tropopause region are also seen when the samples are stratified by the occurrence of relatively dense anvil cloud at the altitude of the aircraft. Time series of such “cloudy” profiles reveal transient temperature features that develop around midmorning (0900 LT): a cold anomaly near 15 km, a weaker warm anomaly from 10 to 12 km, and a rather strong but short-lived warm anomaly near 17–18 km. The heating rate implied by the data is roughly 5–10 K day$^{-1}$. We hypothesize that subsidence is important for the lower-stratospheric anomaly; a combination of ascent and, secondarily, radiative cooling are responsible for the negative anomaly at 15 km; and radiative heating or heating from ice formation control the transient warming in the layer below.

The fact that cloud-related temperature signatures are apparent in composite temperature profiles near the center of circulation of the disturbances investigated herein is consistent with the greater abundance of organized convection near the center. The greater coverage of cold cloud and the colder minimum cloud-top temperature in developing disturbances is consistent with the greater organization and intensity of convection in these disturbances. The fact that convection congeals around the circulation center in the comoving frame of reference was emphasized by Dunkerton et al. (2009). However, the present analysis does not provide
evidence that temperature anomalies in the TTL play a causal role in genesis. Rather, the temperature signal appears to be an indicator of the intensity and coverage of convection and the physical processes operating within and above the anvil clouds. The deep, relatively warm innermost 200 km of developing disturbances, extending well below the TTL, may directly influence genesis by confining the vertical mass flux of convection to lower altitudes. More analysis is needed to assess how the TTL temperature anomalies project onto balanced motions, such as the upper-tropospheric anticyclone.

An important potential future deployment of the MTP is in conjunction with a W band (radar that will soon be deployed on the NSF–NCAR GV). To the extent that this Doppler radar can scan vertically through 180°, it will be able to observe both cloud-top height and the base of the anvil (in areas of light precipitation). Concurrent vertical velocity and temperature profiles will also be important for documenting the relationship between clouds, thermodynamics, and vertical motion. Such measurements will aid in untangling the relationship between radiative processes and adiabatic temperature changes in the UTLS region.

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