Thermodynamic Environments of Deep Convection in Atlantic Tropical Disturbances

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ABSTRACT

Conditional composites of dropsondes deployed into eight tropical Atlantic weather systems during 2010 are analyzed. The samples are conditioned based on cloud-top temperature within 10 km of the dropsonde, the radius from the cyclonic circulation center of the disturbance, and the stage of system development toward tropical cyclogenesis. Statistical tests are performed to identify significant differences between composite profiles. Cold-cloud-region-composite profiles of virtual temperature deviations from a large-scale instantaneous average indicate enhanced static stability prior to genesis within 200 km of the center of circulation, with negative anomalies below 700 hPa and larger warm anomalies above 600 hPa. Moist static energy is enhanced in the middle troposphere in this composite mainly because of an increase in water vapor content. Prior to genesis the buoyancy of lifted parcels within 200 km of the circulation center is sharply reduced compared to the buoyancy of parcels farther from the center. These thermodynamic characteristics support the conceptual model of an altered mass flux profile prior to genesis that strongly favors convergence in the lower troposphere and rapid increase of circulation near the surface. It is also noted that the air–sea temperature difference is greatest in the inner core of the pregenesis composite, which suggests a means to preferentially initiate new convection in the inner core where the rotation is greatest.

1. Introduction

It is well recognized that tropical cyclone formation requires a rotationally constrained, thermodynamic transformation of the tropical atmosphere that often requires many days to complete. An important component of the recent set of field experiments—the Pre-Depression Investigation of Cloud Systems in the Tropics (PREDICT; Montgomery et al. 2012, hereafter M12), National Aeronautics and Space Administration (NASA) Genesis and Rapid Intensification Processes (GRIP; Braun et al. 2013), and National Oceanic and Atmospheric Administration (NOAA) Intensity Forecasting Experiment (IFEX; Rogers et al. 2006)—was to sample tropical disturbances for several days from their earliest stages of organization through tropical storm formation or throughout the failure of such formation.

A total of eight systems were well observed during the 2010 Atlantic hurricane season, and a sufficient number of total missions were flown in PREDICT alone to allow stratification of the observations—primarily dropsondes—into various categories of system organization (Komaromi 2013).

In situ thermodynamic observations of nascent tropical systems are relatively rare, especially through the depth of the troposphere and with sufficient time and spatial resolution to document mesoscale evolution of tropical systems. The most comprehensive composite analysis of tropical systems in various stages of development is the work of McBride (1981) and McBride and Zehr (1981). This seminal work was nonetheless constrained by a modest number of soundings available for any given tropical system, roughly six to eight soundings within a 7° radius for each disturbance in a given stage of development. Thermodynamic soundings have more recently been collected from dropsonde missions for many instances of developing and nondeveloping tropical disturbances, but they have typically sampled systems only once or twice in the days leading to tropical cyclone formation (Ritchie and Holland 1997; Bister and Emanuel 1997; Raymond et al. 1998; Zawislak and Zipser 2010; Montgomery et al. 2010; Raymond et al.
2011). Bister and Emanuel (1997) and Raymond et al. (1998) analyzed flight-level and Doppler radar data from numerous flights into several developing eastern Pacific systems, but did not analyze data from dropsondes. Raymond et al. (2011), using Doppler radar analyses of western Pacific systems, is one of the few studies to distinguish the thermodynamic character of regions within the broad cyclonic circulation of the pregenesis disturbance that are dominated by deep convection from those regions that are relatively suppressed.

The large number of dropsondes available from the Atlantic field studies of 2010 have been analyzed by Smith and Montgomery (2012), Davis and Ahijevych (2012, hereafter DA12), and Komaromi (2013), who collectively documented the relatively moist conditions that accompany the “pouch,” a cyclonic vorticity region that is characterized by recirculation in the reference frame moving with the system (Dunkerton et al. 2009, hereafter D09). Davis and Ahijevych (2012) emphasized the importance of vertical alignment (or near alignment) of horizontal circulations through a deep layer prior to the genesis of Karl and Matthew. This alignment, also noted by D09, Raymond et al. (2011), and Rappin and Nolan (2012) is important for eliminating relative flow than can prevent the establishment of near saturation through a deep layer (Nolan 2007).

Previous studies based on data from PREDICT have emphasized the broad-scale aspects of three tropical disturbances observed during the project: Gaston, Karl, and Matthew. These studies have not addressed clouds, principally because the National Science Foundation (NSF)-National Center for Atmospheric Research (NCAR) Gulfstream V (GV) lacked remote sensing of cloud and precipitation. However, the present study focuses on the thermodynamics of regions characterized by different cloud types and addresses an important question related to the genesis of tropical cyclones—namely, how the thermodynamics of cloudy regions in developing tropical disturbances differs from the thermodynamics of tropical cloud regions elsewhere, particularly from cloudy regions in nondeveloping disturbances. The thermodynamic characteristics are then used, in conjunction with results from other recent studies (e.g., Raymond et al. 2011; Kilroy and Smith 2012; Komaromi 2013), to infer aspects of the divergence profiles that bear directly on vorticity changes.

Deep convection in the tropics, organized on the mesoscale, includes both buoyant updrafts and attendant stratiform anvils (Houze 1997). Mesoscale convection often leaves a thermodynamic signature that may persist over a relatively long time compared to the mass flux signature, such as a cold pool, and furthermore, the thermodynamic structure is related to the rotational structure itself through dynamical balance, making thermodynamic signatures a useful marker of both dynamics and cloud physics. The present paper examines thermodynamic structures using soundings composited according to the cloud-top temperature surrounding them, distance from the circulation center, and stage of tropical cyclone development. Through this multifaceted conditional sampling, we address whether convection in certain regions and phases of development is somehow distinct. The present paper emphasizes the use of statistical significance tests to determine the salient differences between composite profiles. We will then infer from these results how the mesoscale dynamics of the pouch, the preferential organization of convection within the pouch, and the character of the convection itself are interrelated and how such relationships pertain to tropical cyclone formation. The thermodynamic focus of the present paper complements other recent studies of the evolution of horizontal wind circulations during PREDICT by DA12 and Komaromi (2013). Our focus should not be interpreted to mean that thermodynamic processes operate independently of rotational constraints. Circulation is implicit in our adoption of a comoving reference frame, centered on the intersection of the wave trough axis and critical line (D09; M12; DA12) and circulation will be an integral part of the interpretation of our results.

It will be shown that the innermost 200 km of pregenesis circulations are moister and more stable than other environments, particularly when contrasted with the inner circulation of nondeveloping tropical waves. These differences exist in regions devoid of cold cloud within 200 km of the center, and are accentuated in regions dominated by cold cloud. Moist, stabilized profiles exhibit only modest buoyancy given realistic thermodynamic assumptions, and help confine the depth of positive buoyancy. These findings will support the interpretation of pregenesis convection having a mass flux profile maximizing in the lower-to-middle troposphere and a relative absence of unsaturated downdrafts, both of which favor the spinup of a near-surface circulation. These attributes are especially evident in cold-cloud regions that tend to amalgamate near the circulation center as the tropical cyclone develops.

2. Method

The dropsonde data used in this study are discussed in detail in M12, DA12, and Komaromi (2013). Table 1 lists the aircraft missions with dropsondes used in this study, and the phase of development into which each mission is placed. The four phases of development are nondeveloping (ND), developing but greater than 2 days prior to genesis (G2), developing and within 2 days of
Table 1. Month and date (based on UTC) of missions by each aircraft partitioned by development stage.

<table>
<thead>
<tr>
<th>Aircraft</th>
<th>ND</th>
<th>G2</th>
<th>L2</th>
<th>TS</th>
</tr>
</thead>
<tbody>
<tr>
<td>GV</td>
<td>0817, 0818, 0821, 0902, 0903, 0910 (2), 0911, 0912, 0920, 0921</td>
<td>0830, 0913, 0914, 0922, 0927</td>
<td>0831, 0901, 0924, 0928</td>
<td></td>
</tr>
<tr>
<td>G-IV</td>
<td>0805, 0906, 0907</td>
<td>0912, 0913, 0914</td>
<td>0916</td>
<td></td>
</tr>
<tr>
<td>DC-8</td>
<td>0906, 0907</td>
<td>0921 (one drop)</td>
<td>0912, 0913, 0914, 0921, 0922</td>
<td></td>
</tr>
<tr>
<td>P-3</td>
<td>0913, 0914</td>
<td>0913, 0914</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Total missions</td>
<td>11</td>
<td>6</td>
<td>15</td>
<td>5</td>
</tr>
</tbody>
</table>

genesis (L2), and tropical storm (TS). No data from a storm at hurricane intensity are included. A distinction from Komaromi (2013) is that data from four aircraft were used in the present study. The only nonhurricane mission excluded herein was the GV mission on 23 August into PG130. This was excluded because of some problems with the dropsondes, and also because the disturbance had almost no convection associated with it.

To obtain an indicator of the character of clouds near dropsondes, the cloud-top temperatures retrieved from 4-km-resolution Geostationary Operational Environmental Satellite (GOES) infrared (IR) data are averaged within 10 km of each dropsonde location at 700 hPa, using the time of each point in the profile to determine the GOES data nearest in time. GOES data are available approximately every 30 min, and we use the data nearest in time rather than temporally interpolating. The temporal frequency of geostationary data is necessary to estimate conditions near each dropsonde separately.

The choice of a radius of 10 km as a “dropsonde neighborhood” is somewhat arbitrary. A representative drift of a dropsonde relative to cloud features is roughly 5 km, given a fall time of 15 min and a velocity difference of 10 m s⁻¹ between the release and splash points of the dropsonde. In addition, the average cloud-top temperature in a region of 10-km radius does not change much during the 15 min that represents both the dropsonde fall time and the maximum temporal difference between the dropsonde and satellite data.

Subsamples of dropsondes were identified based on the average cloud-top temperature in the neighborhood of each dropsonde. We emphasize that the cloud-top temperature was used only to assign dropsondes to a given composite. A cloud-top temperature of −20°C divided the total dropsonde sample into roughly equal samples of “cold-cloud soundings” and “warm-cloud soundings.” This temperature threshold has a physical justification in separating most of the cumulus congestus from deep convection (Johnson et al. 1999). Their study determined the separation point to be near 7–8-km altitude, and this altitude is close to −20°C in the tropics (Dunion 2011).

Because highly skewed or even bimodal distributions of cloud-top temperatures are possible in a dropsonde neighborhood, the average cloud-top temperature may not indicate the dominant cloud top. Counting the IR temperature of each pixel in the 10-km-radius circles, it turns out that fewer than 7% of the pixels have temperatures that are not consistent with the category represented by the average. In other words, fewer than 7% of the pixels in the cold-cloud sample of dropsonde neighborhoods (≤−20°C) indicate cloud tops warmer than −20°C, and vice versa. Thus, from the point of view of the threshold of −20°C, the cloud-top temperatures surrounding each dropsonde are spatially consistent with each other.

We also checked the sensitivity of the assignment of cloud-top temperature category to the time displacement of the satellite data relative to the time of the dropsonde data. Given the two scans that bracket the time of each dropsonde, choosing the cloud-top temperature derived from the IR data further removed in time results in a change of the assigned cloud category in only 4% of the dropsondes.

The classification according to cloud-top temperature is not indicative of whether a dropsonde actually falls through a convective updraft or anvil cloud. Rather, the cloud-top classification is meant to indicate whether the region around the dropsonde is directly influenced by deep convection. A few dropsondes will indeed penetrate deep convection cells, though not many because the cells are small. Many will fall through anvils with varying rates of precipitation beneath. The collection of such dropsondes represents cold-cloud soundings; that is, soundings influenced directly by convection either through detrainment, diabatic heating, or downdrafts. The sampling uncertainty associated with each dropsonde is mitigated by considering composites of several tens of dropsondes. A similar rationale was adopted by Sherwood and Wahrlich (1999). Herein, cold-cloud soundings are contrasted with warm-cloud soundings that descend through areas with cloud-top temperatures greater than −20°C. The influence of deep convection is less direct on warm-cloud soundings and we expect to see a greater influence of mesoscale circulations on the measurements.

In addition to cloud-top temperature, the dropsondes are further subsampled according to the distance from
the center of the pouch (defined as in M12 and DA12). We examined both a 200-km- and 300-km-radius threshold to separate the “inner core” from the outer portion of the pouch. While the greater-radius threshold divides the subsamples fairly evenly, the more distinct structures are found within the smaller radius, consistent with Wang (2012). We will show only results stratified according to the distance of the dropsonde from the circulation center relative to a 200-km radius.

Rather than composite the full soundings, we composite anomalies from a reference state obtained from the 6-hourly National Centers for Environmental Prediction (NCEP) Final Analysis (FNL) averaged at each mandatory pressure level over an area of 2000-km diameter centered on the pouch, but not including the innermost 200 km. The reference state nearest the time of each dropsonde is used. We use this reference state to reduce biases that can result from missions being conducted in different regions or times and the highly variable number of dropsondes in the different cloud-top temperature or radius partitions for any given mission. For instance, we might have 15 dropsondes in clear air and 5 drops through cold clouds in a system over the central Atlantic, but the reverse for a system several weeks later over the Caribbean. Rather than distinguishing soundings with different cloud neighborhoods, straight averaging of the full profiles would partly reflect regional differences, large-scale pattern changes, or even the seasonal cycle. The use of an instantaneous, large-scale average reference state allows us to composite relatively small anomalies without such potential biases. The FNL may have biases of its own when compared to the dropsondes, but overall biases in the tropics do not affect the differences in composite profiles. The use of this space- and time-dependent reference state will produce results that differ from recent studies by Smith and Montgomery (2012), who used the average of all dropsondes collected over the lifetime of a given disturbance as a reference profile, and Komaromi (2013) who used the mean of all dropsondes from PREDICT as a reference profile. Nonetheless, our basic conclusions appear consistent with Komaromi (2013) despite substantial differences in methodology. The reference state is used only to compute virtual temperature anomalies. Quantities such as relative humidity or buoyancy do not require a reference state.

Statistical significance is assessed using 95% confidence intervals based on a one-sample Student’s $t$ distribution [Wilks 2006, his Eq. (5.3)] given $n$ soundings and $n - 1$ degrees of freedom. In other words, the true population mean is likely to fall within the confidence interval 95% of the time. We also tested explicitly the significance of differences between profiles within a 200-km radius and profiles between 200 and 700 km from the center. This test was also performed at the 95% level, assuming unequal variance for the inner- and outer-radius soundings (Ruxton 2006).

3. Results

a. IR temperature

To determine if the clouds in dropsonde neighborhoods are representative of the overall distribution of clouds, the statistics of cloud-top temperature are composited both in dropsonde neighborhoods and over the entire disturbance out to a 700-km radius. The only constraint on the latter sample is that it represents the distribution of cloud-top temperatures during the times of aircraft missions (GV, NOAA G-IV, NASA DC-8, and NOAA P-3). A direct comparison of the relative frequencies of cloud-top temperatures between the two datasets shows that the distributions are nearly the same within 200-km radius of the center of circulation (Fig. 1a). For radii beyond 200 km, there appears to be a greater sampling of cold clouds by the aircraft than would be expected from the overall distribution. The reason for this “fortuitous” sampling of cold clouds is not clear, but it is helpful for obtaining an acceptable number of cold-cloud soundings farther from the pouch center.

Considering now distributions of cloud-top temperature for different degrees of system organization, there is a clear contrast of sparse cold-cloud neighborhoods in nondeveloping systems from a concentration of cloud tops colder than $-40^\circ C$ in developing cases, especially within 2 days of genesis and beyond a radius of 200 km (Fig. 2). The sample sizes for tropical storms and developing systems more than 2 days prior to genesis are relatively small, corresponding with the smaller number of missions (Table 1).

b. Thermodynamic profiles

1) VIRTUAL TEMPERATURE

Profiles of virtual temperature anomalies (Fig. 3) indicate several common features but also some notable trends as a function of system organization. One common aspect is the appearance of a sharp negative anomaly at 600 hPa in cold-cloud soundings. This would appear to be the signature of the melting level. The signature is weaker for the warm cloud-top sample, and if we restrict a sample to still warmer cloud tops (warmer than $-10^\circ C$) the signal disappears altogether. We should be cautious with this interpretation because it could also indicate the melting of snow or ice on the dropsonde package itself. However, Johnson et al. (1996) have observed a cool anomaly near the melting level in tropical rawinsondes as well. We also
note that there is generally a signature of cooling near and below 900 hPa, approximately cloud base. While there could be effects of evaporation of water from the instrument influencing the data, Sherwood and Wahrlich (1999) and Davis and Lee (2012) found a similar feature using rawinsondes.

A more subtle similarity involves the warming signature in the upper troposphere. Working down the left or right column, the warm anomaly in the 300–400-hPa layer grows with increasing degree of development. Warm anomalies of 1°–2°C are consistent with those found in the composite study of organized convection by Sherwood and Wahrlich (1999).

Virtual temperature anomalies in the lower troposphere vary considerably with both radius and stage of development. Nondeveloping cases feature a warm anomaly that peaks at 925 hPa (Figs. 3a,e). This warm anomaly is stronger in the inner-core sample of cold-cloud soundings than at larger radii. In the early stages of developing cases (G2), the same signal is present. The inner-core features positive virtual temperature anomalies between the nominal boundary layer top (roughly 950 hPa) and the 0°C level. But in the L2 stage, the air beneath the 0°C level in the cold-cloud composite becomes significantly cooler than air outside the inner core (Fig. 3c). The most likely source of this cool air is evaporation beneath anvils. Note also that the inner-core boundary layer of the warm-cloud composite is also cooler than at greater radii, suggesting the transport of cool air laterally from cold-cloud to warm-cloud areas in the lowest 2 km. The combined effect of cool air below 700 hPa and warm air above 600 hPa yields a pronounced stabilization in the L2 composite, especially in cold-cloud regions. This stabilization is most pronounced if one compares the cold-cloud inner-core composite with the warm-cloud composite in the outer region, the latter representing air that is relatively undisturbed by convection.

The enhanced stable stratification is consistent with a positive anomaly of potential vorticity maximizing near 600 hPa and a cyclonic vortex whose tangential wind maximizes in the lower-to-middle troposphere. These attributes are consistent with the analyses of Komaromi (2013) and Raymond (2012).

In the tropical storm stage, the virtual temperature profile is dominated by the warm anomaly in the middle and upper troposphere (Figs. 3d,h). The relatively small sample size for most of these composites results in few significant differences. The warm anomaly in the warm-cloud, inner-core soundings notably exceeds that in the warm-cloud soundings at greater radii. This warming away from areas of cold cloud may be the result of subsidence warming induced through the tipping of the vortex in vertical shear and associated downshear displacement of the precipitation. Tropical Storm Fiona, during which roughly half the data in the tropical storm composite were obtained, developed a highly asymmetric precipitation pattern as it became influenced by vertical shear.

2) MOIST STATIC ENERGY

The radial contrast in moist static energy (MSE) prior to genesis is also quite striking (Figs. 4c,g). While MSE within 200 km of the circulation center is higher than
MSE outside 200 km for each stage of organization, whether near cold clouds or not, an initial enhancement of MSE near 500 hPa more than 2 days prior to genesis (Figs. 4b,f) becomes large through the layer from 800 to 400 hPa prior to genesis in the L2 composite. This is primarily because of the enhancement of water vapor mixing ratio within 200 km as genesis approaches. This enhancement occurs even in regions away from the coldest cloud tops.

In contrast to the L2 composite, the MSE perturbation in the nondeveloping composite is largest in the lower troposphere and decreases upward, most notably within the innermost 200 km. The fact that MSE anomalies are generally positive in all profiles suggests the presence of a dry bias in the FNL relative to the dropsondes. This bias does not alter the basic result that the vertical gradient of the perturbation MSE is reversed between nondeveloping and developing cases.

3) RELATIVE HUMIDITY

Composite profiles of relative humidity, obtained without subtracting a reference state, reveal higher values in the innermost 200 km of circulations compared to greater radii primarily within 2 days of genesis (Figs. 5c,g). One might expect that cold-cloud soundings would reveal greater relative humidity owing to spreading anvils and evaporation of precipitation beneath them. This effect is clearest in the nondeveloping composite, and subtly indicated in the L2 composite (see below). In developing cases, the warm-cloud, inner-core region is particularly moist in the middle and upper troposphere compared with nondeveloping cases. In the G2 composite, humidity near 80% extends up to 500 hPa in the warm-cloud composite, but only to 850 hPa in the corresponding nondeveloping composite. This deeper moisture in the G2 composite is roughly coincident in time with a warm

![Fig. 2. Histograms of counts of cloud-top temperatures averaged within 10 km of dropsondes in 10° bins; gray denotes histograms within 200 km of the circulation center; black outline denotes histograms composited around dropsondes within 700 km of the center for (a) ND composite, (b) G2 composite, (c) L2 composite, and (d) tropical storms.](image-url)
FIG. 3. Profiles of virtual temperature anomalies as a function of development stage, cloud-top temperature [(a)–(d) cold and (e)–(h) warm] and radius (<200 km: dark gray and solid lines and >200 km: light gray and dashed lines). The 95% confidence interval is indicated by gray ribbons centered on the mean at each level. Statistically significant differences between inner and outer radii are denoted by black dots on right side of each panel.
FIG. 4. As in Fig. 3, but for moist static energy anomalies (kJ kg$^{-1}$).
FIG. 5. As in Fig. 3, but for relative humidity (not deviations from a reference state).
anomaly in the upper troposphere in the inner-core cold-cloud composite. These temperature and moisture enhancements may reflect greater spatial organization of deep convection in the early stages of the developing composite compared with the nondeveloping composite (Komaromi 2013).

In addition, cold-cloud soundings in the innermost 200 km of the pregenesis (L2) composite show greater relative humidity than cold-cloud soundings in the inner core of both the nondeveloping (not shown) and early-stage developing (G2) composite (Fig. 6a). The difference is largest at 700 hPa. Combined with the relatively warmer air slightly below this level (Fig. 3a), this implies a contribution from unsaturated mesoscale downdrafts and an “onion”-type sounding (Zipser 1977) near the pouch center for nondeveloping and early-stage developing cases. A difference of similar magnitude exists between the cold-cloud and warm-cloud composites within the inner core of the pregenesis composite where the cold-cloud soundings are moister throughout much of the troposphere (Fig. 6b). We suggest this is evidence for deep, moist convection and its attendant stratiform anvils extending the depth of the moisture in the inner core and, perhaps in concert with shallow convection, bringing the lower troposphere closer to saturation.

Given the profound increase of precipitation efficiency that occurs near saturation (Holloway and Neelin 2009), the seemingly modest increase in relative humidity in the pregenesis inner-core composite is nonetheless vital for development. But what determines the timing of this final push to near saturation and then genesis? Raymond and López Carillo (2010) and Davis and Ahijevych (2012) argued that vortex alignment (or at least partial alignment) is a crucial process for genesis because it eliminates system-relative flow across the region of strongest circulation, and even with a relatively moist pouch, such ventilation can inhibit genesis. Without ventilation, the moistening resulting from stratiform regions can be circulated through the inner-core regions devoid of cold cloud (D09), thus raising the relative humidity throughout the inner core. The time scale for air to circulate around the inner core is roughly one day, given pregenesis winds of 10 m s$^{-1}$ and a radius of 200 km. Therefore we would expect the consequences of alignment on the humidification of the inner core to operate within the temporal granularity of the development stages used herein. In other words, humidification occurs quickly after the vortex aligns.

4) BUOYANCY

Buoyancy is assessed separately for each sounding in a given composite by comparing the virtual temperature of a parcel lifted from 100 m above the surface to the virtual temperature of the “environment” given by the sounding. A parcel initial altitude of 100 m was argued by Romps and Kuang (2011) as being most representative of the origins of lifted parcels in tropical moist convection. As demonstrated in several studies (Xu and Emanuel 1989; Williams and Renno 1993; Molinari et al. 2012), specific assumptions about cloud processes make a large difference in buoyancy and convective available potential energy (CAPE). There are three major considerations: (i) whether ascent is pseudoadiabatic or reversible, (ii) whether the latent heat of fusion is
included, and (iii) how entrainment is handled. The entrainment rate generally exerts a large influence on buoyancy computations and the rate of entrainment is difficult to represent unless one considers large-eddy simulations (LES).

We follow Emanuel (1994) in defining moist entropy that is conserved under reversible processes, but we include the effects of ice. Entrainment is included by prescribing a fractional rate at which a parcel mixes with its environment $e$ expressed per kilometer, and mixing a fraction of environmental air $e\Delta z$, where $\Delta z$ (also in kilometers) is the increment by which a parcel is lifted. Romps and Kuang (2010) showed that the entrainment rate in LES is approximately constant with height. In the present treatment, temperature, water vapor, and condensate (if any) are mixed separately. Through mixing of temperature, dilution of water vapor, and evaporation and sublimation of condensate, the parcel entropy decreases. Entropy is recalculated and the parcel is lifted incrementally once again. Freezing of condensate is applied after each incremental lift, but before entrainment. The incremental amount of condensate frozen at each step is chosen so as to force the fraction of total frozen condensate to vary linearly from 0% at 0°C to 100% at −40°C.

Because buoyancy is already a difference of virtual temperatures (parcel minus environment), there is no need to use a reference state in the compositing. This allows us to use many more levels of data (every 10 hPa) that can reveal shallow layers of convective inhibition.

Herein, we use entrainment rates of 4% and 10% km$^{-1}$; the latter was used by Molinari et al. (2012). We consider both pseudoadiabatic and reversible lifting. However, since the amount of condensate in deep clouds typically is much less than reversible lifting would imply, we also include a calculation in which half of the water vapor that condenses in each incremental lift is allowed to fall out. While there is no firm justification for this amount of fallout, the assumption produces buoyancy profiles in between pseudoadiabatic and reversible profiles, neither of which is realized in practice. Here, we focus on the warm-cloud composites (i.e., relatively free of deep convection) with the assumption that these regions represent an environment appropriate for parcel-based diagnostics.

The basic result is summarized from buoyancy profiles that include an entrainment rate of 10% km$^{-1}$, and allow half of the condensate produced at each incremental lift to fall out of the parcel (Figs. 7a,b). The least buoyant parcels are found in the innermost 200 km of the L2 composite. The outer regions of both the ND and L2 composites are fairly similar. The primary difference is the thermodynamic character of the inner pouch prior to genesis.

A similar comparison of buoyancy profiles from nondeveloping and near-genesis composites is valid across a wide range of thermodynamic assumptions (Figs. 7c,d). Using smaller entrainment values preferentially increases the buoyancy in the middle and upper troposphere. Condensate loading exerts the greatest negative effect in the lower-to-middle troposphere because above this level the heat capacity of condensate compensates for water loading in the case of reversible lifting. In the case of entrainment and fully reversible lifting, buoyancy is essentially eliminated in the inner core of the pouch in the L2 composite. Under assumptions generally considered most thermodynamically realistic, the convective available potential energy one would infer from the buoyancy profiles is at most a few hundred joules per kilogram in the L2 composite.

A more subtle aspect of the profiles is that the L2 composite tends to be fairly close to neutrality over a deep layer for a wide range of entrainment rates from about 4% to 8% km$^{-1}$ for reversible lifting (not shown). The implication of that is, given a moderately wide confidence interval (such as shown for the innermost 200 km in Fig. 7b), one should expect both deep and shallow convection in such an environment. The prevalence of both convection types is hypothesized to contribute to a bottom-heavy mass flux profile (Schumacher et al. 2007) that is favorable for low-level spinup of vorticity, and is consistent with the results of ensemble simulations in moist, more stable environments (Raymond and Sessions 2007).

Also of note is the presence of negative buoyancy immediately above the initial altitude of the parcel. The negative buoyancy peaks at about −0.5°C for the L2 composite and is roughly half that for the nondeveloping composite. The inhibition in the L2 composite is clearly overcome because further deep convection and genesis soon follow. Surface fluxes of heat and moisture are well known to initiate the recovery of the boundary layer after convection (Johnson and Nicholls 1983; Jorgensen et al. 1997). To evaluate the air–sea disequilibrium, we compute the difference between 10-m air temperature and sea surface temperature (SST) using daily 0.25° SST analyses (Reynolds et al. 2007), interpolated spatially to the location of each dropsonde. Of note in the present case is that the air–sea temperature difference in the L2 composite is significantly greater than it is in the ND composite between 50- and 150-km radius, and moreover shows a strong radial dependence (Fig. 8). There were too few dropsondes to make a corresponding statement about the G2 composite. Given that we observed similar earth-relative surface wind speeds of 6–8 m s$^{-1}$ in all composites (not shown), we would...
expect enhanced surface fluxes of heat and water vapor and more rapid destabilization in the inner core of the L2 composite compared to the nondeveloping composite. The radial localization of the more rapid destabilization would help focus new convection near the center of the pouch where the environment is already nearly saturated over a deep layer.

c. Spatial patterns

Composite spatial patterns of thermodynamic parameters are constructed by representing the center-relative location of all dropsondes within a given development category with a circle colored by the value of the thermodynamic perturbation from the reference state (Fig. 9). We can infer the composite spatial pattern of deep convection relative to the circulation center simply by noting where the dropsondes are found in the cold- and warm-cloud composites (Fig. 9). Recall that the dropsondes represent a relatively unbiased sampling of cold-cloud and warm-cloud regions (Fig. 1).

The relative humidity in the middle troposphere (Fig. 9) is generally a good discriminator of moist versus dry atmospheric states. While cold-cloud soundings generally have greater relative humidity in the middle troposphere...
compared with warm-cloud soundings regardless of the
stage of development (Figs. 9a,b; see also Fig. 5), the
high humidity is far more pervasive in the warm-cloud
soundings in the L2 composite. Furthermore, even the
cold-cloud soundings in the nondeveloping composite
sometimes reveal dry air at 500 hPa, which implies that
these soundings descended near or through “dead” an-
vils. That is not the case in the L2 composite.

While relatively cool air in the subcloud layer is a
property of all composites in cold-cloud regions (com-
pared with warm-cloud regions), as is cooling at the
melting level (Fig. 3), it is only the composite within
2 days of genesis that exhibits negative virtual temper-
ature anomalies above cloud base. Together with the
warming at 400 hPa, this produces the most pronounced
stabilization of any profile. The spatial pattern of this
stabilization, represented by the difference in virtual
temperature anomalies between 400 and 925 hPa (Fig. 10)
indicates a strong clustering of stable profiles within \( r = 200 \) km only in the pregenesis composite (Fig. 10b). In
this composite, the stabilization between 400 and 925 hPa is
more than \( 3^\circ C \) for several of the soundings. The cold-cloud
profiles near the center of the pouch collectively define
a mesoscale structural modification prior to genesis: cool
at low levels, warm in the upper troposphere, and nearly
saturated.

The simplest explanation for the stabilization is the
vertical mixing of moist entropy in nearly saturated re-
gions. In this view, downdrafts still occur, and indeed
are crucial for reducing the moist entropy in the lower
troposphere. But the constraint of near saturation
requires a decrease in virtual temperature (relative to
the surroundings), just as it requires an increase in vir-
tual temperature near the moist-entropy minimum in
the middle troposphere. It is also important to re-
member that “cold” and “warm” have little absolute
meaning in the present context, because it is only the
vertical and horizontal gradients that are most impor-
tant for the dynamics. Thus, the total temperature may
or may not experience a decrease at a particular level.

One may also note that the spatially localized stabi-
larization is consistent with the presence of enhanced
potential vorticity (PV) in the inner core of the cold-
cloud, pregenesis composite. While we cannot compute
the full PV for each sounding, the coarse-grain property
of PV asserts that there will be a midtropospheric me-
oscale PV anomaly in the composite mean if the flow is
nearly balanced (Raymond 2012). This PV anomaly will
be enhanced within the cold-cloud region compared to
elsewhere in the innermost 200 km. In this sense, the
amalgamation of deep convection near the center of the
pouch is tantamount to the amalgamation of PV.

4. Summary

The present study has analyzed a comprehensive set
of dropsondes for eight tropical disturbances observed
during the 2010 Atlantic hurricane season—including
dropsondes deployed by NSF, NASA, and NOAA
aircraft—by conditionally sampling the data according
to the average temperature of nearby cloud tops, radius
from the circulation center, and stage of development.
We have found results that are broadly in agreement
with other recent studies, especially Komaromi (2013),
but with different stratifications of the data, more ther-
modynamic variables examined, and the addition of
confidence intervals and statistical significance tests to
quantify the robustness of differences between samples.
Statistically significant results include the stabilization
of the virtual temperature profile, the increase of water
vapor over a deep layer, the decrease of parcel buoy-
ancy, and the increase of air–sea temperature difference
within 2 days of genesis and within roughly 200 km of the
center of circulation (in the comoving frame of refer-
cence). The stabilization and moistening are greatest
in soundings deployed in regions of cold cloud tops. In
these regions, dominated by mesoscale anvils, warm
anomalies in the upper troposphere appear more than
2 days prior to genesis. In these same regions, warm
anomalies in the lower troposphere are replaced by cool
anomalies prior to genesis.

Turning to the question asked in the introduction,
whether convection near the center of circulation is
distinct from convection elsewhere (and under what

![Fig. 8. Air temperature at 10-m altitude minus sea surface
temperature (°C) for L2 (black) and ND (gray) composites. Thin
lines denote 90% confidence intervals; black for L2 composite and
dashed gray for ND composite.](image-url)
circumstances), the answer appears to be “yes.” The separation of soundings into four categories for a given stage of development proves insightful. For instance, within 2 days of genesis, cold-cloud soundings within 200 km of the pouch center exhibit greater stability than profiles farther from the center but still near deep convection. The stability increase, measured in terms of virtual temperature difference between 400 and 925 hPa, can exceed 3.8°C. In fact, cold-cloud soundings in non-developing disturbances actually feature lower stability and greater conditional instability, consistent with the results of Smith and Montgomery (2012). Within 200 km of the center, soundings in warm-cloud or clear regions also show enhanced stability, but less compared with cold-cloud soundings. The change of the warm-cloud regions near the center is believed to be an indirect effect of organized convection—namely, horizontal and vertical transport of moisture within the inner recirculation region and warming in the mid-to-upper troposphere associated with the vertical structure of the mesoscale vorticity. The region near the center is characterized by a vortex that maximizes its circulation in the lower-to-middle troposphere (Komaromi 2013) and thus would be expected to feature warming at higher levels from gradient-thermal-wind balance (Raymond 2012).

At this point, it is important to reconcile the present results with other studies, especially about the issue of whether there is in fact a cool anomaly in the lower troposphere prior to genesis. The question is “cool relative to what?” In general, some degree of subcloud cooling characterizes all cold-cloud regions, but in the L2 composite within 200 km of the pouch center, the cooling extends well above the subcloud layer. The cooling is also strongly confined in radius. If our inner disk is taken to be 300 km rather than 200 km, the significance of the negative virtual temperature anomaly within the disk vanishes. Komaromi (2013) found essentially the same result between 1 and 2 days prior to

FIG. 9. Relative humidity at 500 hPa represented by circles colored according to humidity value, plotted in a system-following coordinate system for the (a) ND cold-cloud, (b) L2 cold-cloud, (c) ND warm-cloud, and (d) L2 warm-cloud dropsondes. The origin is the pouch center and the circle is drawn at 200-km radius.
genesis, despite using the PREDICT mean sounding as a reference state instead of a space- and time-dependent state as used herein. He also found that the strongest stabilization, with negative anomalies below 600 hPa and positive above, was confined within about 200 km of the center. However, we do not find cool anomalies in the lower troposphere more than 2 days prior to genesis; this is likely due to the use of a different reference state. Nevertheless, collectively these findings echo the results of McBride and Zehr (1981, their Fig. 1), Lee (1989), Bister and Emanuel (1997), and Raymond et al. (2011) from observations, Nolan (2007) from idealized simulations, and Wang et al. (2010, their Fig. 1) from numerical simulations of the genesis of Hurricane Felix (2007). All show enhanced stabilization of the temperature profile preferentially inside a radius of 200–300 km prior to genesis.

The more stable inner-core changes the buoyancy profile of parcels lifted from the boundary layer so that there is much less CAPE than in nondeveloping systems or at greater radii in developing systems. Because of the enhanced water vapor content, buoyancy through the midtroposphere is relatively insensitive to entrainment assumptions. Quasi-equilibrium convection in such an environment has been shown to maximize the mass flux in the middle troposphere, as opposed to the upper troposphere as in other tropical environments. Upward motion is maximized at lower altitudes with convergence dominating divergence in the lower troposphere (Raymond and Sessions 2007). In concert with the greater inner-core relative humidity and static stability prior to genesis is an increase in the air–sea temperature difference. This suggests a preferential recovery of the boundary layer, and the likelihood of new convection, at a radius of about 100 km or so from the center of circulation, similar to the scenario described by Bister and Emanuel (1997). Such convection, even with small buoyancy, would efficiently produce precipitation, maximize its mass flux at fairly low levels, produce few downdrafts with large negative buoyancy, and be

![Figure 10](image-url)  
**Fig. 10.** As in Fig. 9, but for the difference in virtual temperature anomalies between 400 and 925 hPa.
contained within the region of maximum circulation. All of these attributes would lead to rapid increase of near-surface circulation.

It is worth remembering that the conditions documented in the present paper, and other similar studies before it, represent the mesoscale structures that accompany genesis. By emphasizing soundings deployed near different cloud types, we have a partial glimpse into the role of convection, but the details are far from complete. We hypothesize, consistent with concepts in D09, that the congealing of convection around the center of the pouch brings with it a characteristic potential vorticity signature that, in turn, reinforces the favorable vertical mass flux profile for lower tropospheric spinup. But the details of this process, in particular the role of the convective scale, require more detailed observations to elucidate.

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**REFERENCES**


