The Creation of a High Equivalent Potential Temperature Reservoir in Tropical Storm Humberto (2001) and Its Possible Role in Storm Deepening

KLAUS P. DOLLING AND GARY M. BARNES
University of Hawaii at Manoa, Honolulu, Hawaii

(Manuscript received 16 March 2011, in final form 15 August 2011)

ABSTRACT

Thirty global positioning system dropwindsondes (GPS sondes) were used to identify and examine the creation of a reservoir of high equivalent potential temperature ($\theta_e$) in the nascent eye of Tropical Storm Humberto (2001). The $\theta_e$ did not increase in the high surface wind portion of the storm as it does in mature hurricanes; instead air spiraled into the light-wind center of the developing storm where it was trapped by subsidence under a mesoscale convectively generated vortex (MCV). An energy budget revealed that the inflow column took 7 h to reach the storm center during which a combined average surface enthalpy flux of $\sim 230 \text{ W m}^{-2}$ was diagnosed via the bulk aerodynamic equations. This estimate is close to the 250 W m$^{-2}$ required for balance based on the energy acquired by the column. The high $\theta_e$ in the lowest kilometer, overlain by a near dry-adiabatic layer under the anvil base, resulted in convective available potential energy (CAPE) exceeding 2500 m$^2$ s$^{-2}$. This conditionally unstable air later served as fuel for the convection within the nascent eyewall. The authors speculate that CAPE of such a large magnitude could accelerate the updraft and stretch the vorticity field, essentially turning garden-variety cumulonimbi into the vortical hot towers argued by several researchers to play a role in tropical cyclone formation and intensification.

1. Introduction

Tropical storms, in contrast to high Saffir–Simpson category tropical cyclones (TCs), are far more asymmetric based on observations of their cloud and reflectivity patterns (e.g., Stossmeister and Barnes 1992; Sippel et al. 2006; Reasor et al. 2005; Heymsfield et al. 2006). Most tropical storms (TSs) have a poorly defined wind maximum located more than 50–75 km from the circulation center with their core characterized by light winds (Riehl 1954). Recently there has been some evidence to support the conjecture that the spatial arrangement of equivalent potential temperature ($\theta_e$) in the subcloud layer of a TS differs considerably from a mature TC. There is no increase of $\theta_e$ collocated with the strongest surface winds in the region under and radially outward of the dominant arc of convection. Instead, it is found in the light-wind core of the developing circulation. In TS Chantal (2001) Heymsfield et al. (2006) detected high $\theta_e$ entering in a channel from the south of the center. Some of this air served as inflow to an arc of convection located to the north, but based on their schematic reproduced here (Fig. 1) some of this air swirled into the center. In the TS stage of Danny (1997) Molinari et al. (2004) showed that the increases in $\theta_e$ occurred radially inward of the radius of maximum winds (RMW; Fig. 2), also in a light-wind regime.

In both of these cases the creation of a dome of high $\theta_e$ in the lower troposphere seems to differ from the reservoir typically observed in the eye of a mature TC. In the mature TC the low-level inflow undergoes substantial and rapid increases of $\theta_e$ in the ring under and outward of the eyewall (Hawkins and Imbembo 1976; Jorgensen 1984b; Black and Holland 1995;Sitkowski and Barnes 2009). A small portion of the inflow apparently fails to ascend and enters into the eye where it continues to receive surface fluxes and after a number of hours becomes the reservoir for the highest $\theta_e$ found in a TC (e.g., Eastin et al. 2005; Barnes and Fuentes 2010). In the TSs observed by Molinari et al. (2004) and Heymsfield et al. (2006) there is no evidence of an annulus with a strong radial gradient of $\theta_e$ located outward of the dominant convective arc. In this paper we investigate the $\theta_e$ increase in the core of TS Humberto (2001).
Until the last decade the TS has been an undersampled stage of a TC for at least two reasons. First, researchers initially focused their observational resources on the higher-category TCs since these are the greatest threat to society. Second, the TS stage often occurs far from aircraft reconnaissance bases and can be a relatively short lived and therefore elusive stage to sample. Examination of this stage may be particularly fruitful since it occupies the period between formation when the circulation and warm core appears and intensification when the link between the ocean and atmosphere is established. Over the last decade the National Oceanic and Atmospheric Administration (NOAA) sampling strategy has responded to the need for more TS observations and there are now available several good datasets; we shall concentrate on Humberto (2001) that was sampled during the Fourth Convection and Moisture Experiment (CAMEX-IV). On 22 September the National Aeronautics and Space Administration (NASA) DC-8 and ER-2, and a NOAA WP-3D captured Humberto in the TS stage when it was 999 hPa.

We shall use these observations to examine the following questions.

1) What is the disposition of $\theta_e$ in TS Humberto (2001) and how was it created?
2) What is the role of the mesoscale convective systems embedded in the developing vortex in the creation of a reservoir of high $\theta_e$?
3) What is the potential role of this reservoir of high $\theta_e$ in either formation or intensification?

The soundings obtained from the dropwindsondes provide an estimate of the moist entropy in an inflow column, the low-level flow patterns of the TS, and the requisite observations to estimate the sensible and latent fluxes at the air–sea interface when combined with sea surface temperature (SST) estimates. From these measurements we can determine how the $\theta_e$ field develops in the early stages of a TC.

2. Data and methodology

a. GPS sonde quality control

The global positioning system dropwindsondes (GPS sondes) have a 2-Hz sampling rate and realize ~7-m vertical resolution in the lower troposphere. Typical errors for pressure, temperature $T$, and relative humidity (RH) are 1.0 hPa, 0.2$^\circ$C, and <5%, respectively (Hock and Franklin 1999). The Atmospheric Sounding Processing Environment (ASPEN) program developed at the National Center for Atmospheric Research (NCAR) was used as a first step to process the raw sonde data initially in the form of the Airborne Vertical Atmosphere Profiling System (AVAPS) files. Details of these quality control algorithms can be found online in the ASPEN users’ manual (http://www.eol.ucar.edu/rtf/facilities/software/aspen/Aspen%20Manual.pdf).

Barnes and Fuentes (2010) compared ASPEN to Editsonde; the in-house software package developed at the NOAA/Atlantic Oceanographic and Meteorological...
Laboratory/Hurricane Research Division (NOAA/AOML/HRD), and found the two programs to produce similar state variable results for 44 sondes deployed in Hurricane Lili (2002). This is after the postprocessed ASPEN data are corrected further for additional moisture errors identified by Wang (2005) and Barnes (2008).

Wind estimates from the GPS sonde depend on the number of satellites the sonde connects with at a given time. Franklin et al. (2003) found that very strong winds near the surface caused an increase in data loss, but at 50 m and above the sonde maintained its links to the various satellites reasonably well. In Humberto on 22 September the winds are lighter resulting in only a 5% loss of wind data from 100 to 10 m.

b. GPS sonde deployment

On 22 September one NOAA WP-3D flew at approximately 600 hPa and deployed 15 successful GPS sondes, the NASA DC-8 was at 300 hPa and dropped 13 sondes, and the Lockheed ER-2 flew at 50 hPa and delivered 2 sondes. Figure 3 displays the deployment locations of 23 of the GPS sondes relative to the storm center for the NOAA and NASA aircraft. There are another seven that contributed to the analyses, but these lie outside the perimeter of the region of interest. Horizontal spacing is about 20 km near the circulation center and about double that by 100 km from the center.

c. Aircraft instrumentation

The NOAA WP-3D deployed 12 airborne expendable bathythermographs (AXBTs) that are used to discern the sea surface temperature under Humberto. The accuracy of the AXBT is discussed by Boyd and Linzell (1993) and is about 0.2°C.

The WP-3D reflectivity observations from the 3.2-cm tail and 5.5-cm lower-fuselage radars are discussed by Marks (1985). We use these scans to determine the convective and stratiform regions, and to place the GPS sonde data with respect to the persistent mesoscale features found in the TS. In situ aircraft sensors are described by Jorgensen (1984a) with updates mentioned by Aberson et al. (2006). These data are handled in a manner similar to that executed by Sitkowski and Barnes (2009).

d. Analysis scheme

Determining storm track was the first step of the analysis. We employed the technique of Willoughby and Chelmon (1982) to establish the circulation center based on aircraft wind data and combined these fixes with best-track satellite centers for periods that extended on either side of the flight time by a few hours. This allowed the GPS sondes deployed at the start and finish of the flight to be placed as accurately as possible. On 22 September the six fixes fit a regression curve with an $r^2 = 0.99$.

The second step was to determine the position of each sonde relative to the circulation center. As the sonde fell its position was corrected for its motion relative to the moving circulation center. These data were then all composited to a central time. Humberto deepened only a few hecypascals during the flight, but reflectivity features did evolve more resulting in our ability to link our derived vortex-scale wind and state variable fields only to the long-lived mesoscale reflectivity features associated with the TS.

In the third step the storm motion $u$ and $v$ components were subtracted from the earth-relative $u$ and $v$ winds for the GPS sondes to obtain storm relative winds. These were then placed into cylindrical coordinates to obtain relative tangential and radial components.

The fourth step was to fill gaps in the data and standardize the data to common altitudes. Gaps less than 300 m thick were populated with linearly extrapolated values. Most of the gaps were much smaller, between 50 and 100 m. The data were then interpolated in 10-m increments starting at 10 m above the sea and continuing to the level where the sonde first measured atmospheric conditions after exiting the aircraft.
In the final step a piecewise cubic Hermite interpolation (Fritsch and Carlson 1980) was performed to create a field for any chosen variable for each level. This cubic spline method preserved the monotonicity and the shape of the data and reproduced the observed value at its observed location. The assembly of each plan view map results in a three-dimensional matrix. This matrix can then be sliced through at any point and in any direction to produce a vertical cross section. The highest resolution is below 600 hPa, just below the altitude of the WP-3D, but coarser maps above that level can be created based on the high-altitude GPS sonde deployments made from the NASA aircraft.

The number of sondes deployed, their spatial distribution, and the assumption of the steady state over the ~3 h to create the composite fields results in our ability to resolve horizontal scales of about 20–25 km near the center and 40–50 km beyond about a degree latitude from the center. Subjective hand analyses were compared for several variables at different levels and matched well with the cubic-spline-derived analysis. More details of the analysis scheme can be found in Dolling (2010).

e. Humberto (2001)

A trough extending southwest from Hurricane Gabrielle was the initial location for the deep convection that was to become Humberto (Beven et al. 2003). Early on 22 September, Humberto was a tropical depression with a minimum central pressure of 1010 hPa and was located in the Atlantic basin near 29°N, 66°W. The movement was to the north-northwest at approximately 4 m s\(^{-1}\). Humberto continued moving to the north and eventually recurved to the northeast. When the first GPS sondes were jettisoned, late on 22 September, Humberto’s minimum central pressure was 1000 hPa and it had attained tropical storm strength. On 22 September, the storm was sampled from 1840 to 2140 UTC. During these 3 h the minimum sea level pressure decreased about 4 hPa. Humberto will deepen to 983 hPa (category 2) by 0000 UTC 24 September. Humberto went through a weakening and another intensifying stage and eventually decayed late on 27 September north of 42° latitude.

3. Results

a. Reflectivity features

Throughout the ~3 h of sampling, Humberto was characterized by an asymmetric reflectivity field, which was essentially comma shaped. When the aircraft arrived just prior to 1900 UTC the dominant reflectivity feature was an arc of convection about 50 km north of the low-level circulation center (LLCC) and thinner bands of cells from 50 to 120 km to the east (Fig. 4a). Stratiform rain existed on the northern periphery of Fig. 4a and over the LLCC save for one small cell about 25 km north of the center. Near the LLCC one can infer a circular flow pattern based on the appearance of the reflectivity field. At about midpoint in the flight (1942:41 UTC; Fig. 4b) the original arc of convection about 50 km north of the LLCC is visible at the top of the picture and a new arc of convection has developed about 25 km north of the LLCC. This was where a single cell was seen in the earlier panel. The returns greater than 32 dBZ near the center again have a pattern that suggests the presence of a circulation. Toward the end of the sampling on 22 September (2006:46 UTC; Fig. 4c) the arc nearest the center of the storm still contained the heaviest rain rates while the original arc farther north had weakened and new cells started to fill the area between the two arcs. Cells to the east of the LLCC in the northern portion of the band remained active (not shown). The full suite of plan views (~70) revealed that Humberto’s reflectivity field remained asymmetric throughout the sampling period. The cells that make up the bands that spiraled in from the southeast through east were shallow (5–6 km) on the upwind portion and became deeper (10–15 km) as the band merged into the northern arc. The most noticeable change was the development of an arc of convection about 25 km north of the LLCC. Tail radar revealed that this arc had tops that reached to 15 km, some of the tallest cells in Humberto during this day.

b. Near-surface structure

The average minimum sea level pressure (MSLP) for Humberto during the sampling was 999 hPa. The LLCC, based on the streamlines at ~10-m altitude (Fig. 5a), was located on the western side of the low pressure area. A confluence zone north of the LLCC was collocated with the convective arc seen in Figs. 4a,c. The wind speed field at 10-m altitude (Fig. 5b) was asymmetric with the winds over 20 m s\(^{-1}\) from the northeast to the north. To the west and southwest, speeds were less than 15 m s\(^{-1}\) and in the low pressure core wind speeds were slightly less than 10 m s\(^{-1}\). The temperature field at 10 m (Fig. 5c) revealed that the high-wind region coincided with air cooler than 24°C while the center of Humberto was warmer than 27°C. The cool air was collocated with the reflectivity field and was at least partially due to downdrafts as equivalent potential temperature decreased as well. Based on the streamlines this cool air did not enter into the core. Specific humidity at 10 m (q, Fig. 5d) was highest in the storm core and lowest to the north in the environmental flow and downstream of the convective arc that was 50 km north of the circulation center. SSTs were 28.5°–28.0°C within a 50-km radius of the center of
the storm. The observations at 10 m will be combined with the SSTs to apply the bulk aerodynamic equations in the lighter wind core of the storm.

c. Conditions above the surface circulation center

The streamlines at 5 km, derived from the GPS sondes, seven of which lie just beyond the region depicted, showed the presence of a mesoscale vortex (Fig. 6) with a center displaced toward the northeast, downshear of the LLCC. This upper-level circulation was in stratiform rain (Fig. 7); the flow pattern supports the conjecture that the anvil, which occupied the 4–10-km layer, emanated from the arc of convective cells 50 km to the north of the surface center. This is verified by the wind patterns up to 7.5 km from the GPS sondes, the winds observed by the DC-8, and the WP-3D winds measured at 600 hPa. A deep sounding in the nascent eye from the DC-8 (Fig. 8) shows that conditions were similar to the classic stratiform rain sounding described by Zipser (1977). Below 575 hPa the sounding was near dry adiabatic and dew-point depressions reached as much as 13°C. A strong inversion with a base near 920 hPa inhibited convection from initiating and subsequently transporting the moist air below that level upward within 25 km of the LLCC.

d. Equivalent potential temperature and streamlines

The flow and $\theta_e$ patterns at 200 m are representative, with only very minor differences, of any level from 20- to 2000-m altitude (Fig. 9a). There was a pool of high $\theta_e$ in the inner core of the storm and low values were located to the north-northwest of the center. This low $\theta_e$ was under and downstream of the convective arc located about 50 km north of the LLCC. The streamlines reveal three general pathways for air parcels initially located around the periphery of the storm. Far to the north there was environmental air, slightly less than 345 K, that moved cyclonically around the west side of the storm. As it reached the latitude of the center it began to warm and exceeded 350 K before it exited to the south. It did not enter into the core of the storm. The second set of streamlines originated from the east and east-southeast, with a $\theta_e$ between 345 and 350 K. This air underwent

---

Fig. 4. The reflectivity field as viewed by the lower-fuselage radar of the WP-3D at (a) 1858:34, (b) 1942:41, and (c) 2006:46 UTC. All three pictures are 120 km x 120 km, tick marks are every 12 km. The black dot marks the center and the white line in (b) is where the tail radar scan is in Fig. 7. The lighter returns on the periphery of each frame that are concentric to the aircraft (small white cross) are due to sea clutter. Color table at right shows dBZ.
confluence that was collocated with the eastern side of the arc of convection located 50 km to the north of the center. As the streamline exited from the convective area it had $u_e$ lower than 340 K. Here the initially warmer air was obviously replaced by downdraft air. This downdraft air then wrapped around to the west of the center and at that point two possibilities emerge because of the uncertainty in the streamlines. The air could continue to the south and not enter the core or there is the possibility that the air could spiral into the core. The third path is air that entered from the southeast then swirled around the center of the storm. This path remained radially inward of the arc of convection to the north. This air underwent substantial increases in $\theta_e$, from 350 to more than 360 K. Note that $\theta_e$ was not increasing in the high-wind region (see Fig. 5b) that was collocated with the convection. In the high-wind region any increase in the low-level $\theta_e$ from surface fluxes apparently was overwhelmed by the convective transports occurring there.

**e. Energy budget for the inflow that reaches the circulation center**

An energy budget for the air that spirals into the LLCC requires an estimate of 1) the SST, 2) conditions at $\sim 10$ m to apply the bulk aerodynamic equations, 3) the divergence the column undergoes during its journey to the LLCC, 4) the trajectory of the air as it flows into the LLCC, and 5) the energy content for the inflow column at its initial and final locations. Airborne expendable...
bathythermographs dropped within 100 km of the LLCC supply an estimate of the SST. The GPS sondes were the source of data for the three-dimensional wind field used to determine the 10-m conditions, the most likely trajectories to the LLCC, divergence, and the column energy at the initial and final locations. Items 1–4 provide an estimate of the total energy gained via the fluxes while item 5 leads to another estimate of the energy gained that is independent of the fluxes and their duration. If the two estimates are close then the budget, despite the

![Streamlines at 5-km altitude determined from the matrix assembled from GPS sondes as a function of latitude and longitude. The red dot depicts the circulation center at the surface. The black line shows the orientation and extent of the RHI scan in Fig. 7. Sondes are listed on the plan view as four-digit integers.](image1)

![Tail radar view in the west–east plane nearly through the center of the storm at 1944:06 UTC. The picture is 20 km high and 120 km across with the aircraft depicted by a white cross in the lower-middle part of the picture at about 4-km altitude. Tick marks in the vertical are every 2 km and in the horizontal are every 12 km. Reflectivities are shown in the color table at the bottom.](image2)
numerous uncertainties common to all energy budgets, is deemed to have captured the essential processes.

Equivalent potential temperature is used for the budget since it is conserved for the evaporation of rain or spray. The change in $u_e$ can be described following the scheme developed by Yanai et al. (1973), and applied by numerous investigators (e.g., Anthes 1982; Wroe and Barnes 2003):

$$\frac{\partial \theta_e}{\partial t} + \mathbf{u}_j \frac{\partial \theta_e}{\partial x_j} - Q_r + \frac{\partial (w \theta_e')}{\partial z} - \epsilon = 0,$$

where $t =$ time, $u_j$ represents the mean or grid-scale components of the three-dimensional velocity vector, $x_j$ represents the spatial components of a three-dimensional vector, $Q_r$ is the radiative divergence, $w$ is vertical velocity, $z$ is height, and $\epsilon$ is dissipative heating. Equation (1) states that the change in storage of $u_e$ plus grid-scale advection $[u_j(\partial \theta_e/\partial t)]$, minus radiative divergence $Q_r$, plus the subgrid fluxes in the vertical, minus dissipative heating are equal to zero. We have made the standard assumption that the subgrid-scale fluxes in the horizontal are much smaller than the vertical fluxes and may be scaled out of the equation. We also assume that the radiative cooling in an environment dominated by a thick anvil cloud is negligible. Dissipative heating has been argued to put sensible heat into the atmospheric boundary layer, especially at high wind speeds (Bister and Emanuel 1998). However, the wind speeds along the trajectories of interest are less than 12 m s$^{-1}$. For a 10 m s$^{-1}$ wind speed the maximum dissipative heating (equivalent to density $\times$ the cube of the wind speed $\times$ the drag coefficient) would be about 1 W m$^{-2}$. Recent direct measurements in the hurricane boundary layer support the notion that only a small portion of this heating may be realized as most of the dissipative heating is apparently lost to the oceanic mixed layer (Zhang 2010), so we have neglected this term as well.

We assume that the horizontal wind and the horizontal gradient of $\theta_e$ are unchanging during the several hours that the inflow would take to reach the LLCC; this steady-state assumption eliminates the storage term, but it is also more suspect given that Humberto is deepening over the period. If this assumption is far from reality, then we would expect poor agreement between the energy difference between the initial and final states and the estimated total energy gained from the fluxes. We are left with the following:

$$V \frac{\partial \theta_e}{\partial n} = -W \frac{\partial \theta_e}{\partial z} - \frac{\partial (w \theta_e')}{\partial z}.$$

The horizontal advection of $\theta_e$ along a trajectory ($V \theta_e'/\partial n$) is maintained by the mean vertical transport ($-W \theta_e'/\partial z$) and the subgrid-scale vertical flux $\partial (w \theta_e')/\partial z$. Mean vertical transport is of course related to the divergence that the column undergoes. This term (first term on rhs)
is important only if there is a vertical gradient of $\theta_e$; this issue will be amplified shortly when we discuss the initial and final column conditions.

A schematic of the situation (Fig. 9b) simplifies the conditions that existed on 22 September. The fates of the low-level air parcels may be classified into three groups. Path “C” is the environmental air that did not interact with the storm, paths labeled “B” fed the convection to the north, were replaced by downdraft air that wrapped around the core then either exited to the south or perhaps entered the core (lack of observations makes either path plausible), and path “A” was the air that enters the core from the southeast. We shall examine paths “A” and “B” to determine if either scenario can be used to explain the observed conditions in the LLCC.

Can the air spiraling into the center along pathway A acquire the requisite energy to account for the observed changes in the column? The size of the core with the higher $\theta_e$ was approximately 25 km in radius (Fig. 9b). This region was devoid of any convective elements and was just radially inward of the new cells that were developing north of the circulation center so we do not have to contend with any convective processes. A vertical cross section, running south-southwest–north-northeast through the LLCC of Humberto (Fig. 10) reveals that the high $\theta_e$ occupied a depth of about 2200 m. Differences between the column of air in the core and the surrounding environment were a maximum near the surface and diminished with height. Note that the vertical gradient of $\theta_e$ is quite strong above $\sim$700 m (Fig. 10), which corresponds to the base of the inversion (Fig. 8). In Fig. 9b “I” denotes the initial location of the column, and the location marked “F” is the final location of the column after it has swirled into the center. These two soundings are shown in Fig. 11. In the lowest 500 m the difference is $\sim$10 K, but this difference becomes quite small by 2200 m.

At first glance one would imagine that the difference in the two vertical profiles represents the energy acquired during the spiral inward. This is not correct, because the column is undergoing convergence during its inward journey. This results in a lifting of the lower layers that have higher $\theta_e$ and a significant alteration of the column values. A schematic of how convergence would alter the sounding is shown in Fig. 12. The initial column in Fig. 12 has 360-K air just below 500 m. Have this initial sounding undergo convergence during the period as it is advected to the final location and the height of the 360-K contour will rise and the result is the second column in Fig. 12. The difference between this second column and the final column is due to the subgrid-scale fluxes, the second term on the rhs of Eq. (2).

We estimate the divergence with the radial wind component along the circumference of the circle with a 25-km radius. The magnitude of the radial flow is determined about every 20 km around this circle, which is essentially octant scale. These radial flow components are determined for altitudes of 50, 100, 200, 500, and 1000 m. For a given layer we estimate the net mass flux into the 25-km radius ring, then estimate the divergence based on the total mass within the chosen layer. This divergence is used to estimate a vertical velocity for the sounding at the aforementioned levels. The horizontal resolution does not allow us to calculate divergence on much smaller scales than this circle. The divergence for the lowest 200 m is $-2.2 \times 10^{-4}$ m s$^{-1}$, which yields a vertical velocity of $4.4 \times 10^{-2}$ m s$^{-1}$. For the 500-m
layer, the mean divergence is $-1.2 \times 10^{-4}$ s$^{-1}$ resulting in a $w = 6.0 \times 10^{-2}$ m s$^{-1}$. These rise rates of 4.4 to 6.0 cm s$^{-1}$ are applied over the time it takes for the air to reach the circulation center, which is approximately 7.0 h. The 500-m layer rose about 1500 m during this time, while the 200-m layer rose over 1100 m. We apply the ascent rates to the initial $\theta_e$ profile to produce the sounding adjusted for the convergence (Fig. 11). The shaded region between this sounding and the final sounding is a graphical depiction of the amount of moist entropy needed from the sea to account for the change. The amount of energy that the column must acquire is

$$\int_{\text{sfc}}^{Z_T} \rho(C_P \partial \theta_e) \partial z = 6.27 \times 10^6 \text{ J m}^{-2}. \quad (3)$$

The integration is performed from the sea to the height $Z_T$ where the two soundings have essentially the same $\theta_e$, $\rho$ is density, and $C_P$ is the specific heat at constant pressure.

The sea surface flux estimates [the subgrid-scale fluxes at the bottom of the column, the second term on the rhs of Eq. (2)] follow the scheme discussed by Wroe and Barnes (2003), but with the updated transfer coefficients recommended by Fairall et al. (2003). These transfer coefficients are based on the Tropical Ocean and Global Atmosphere Coupled Ocean–Atmosphere Response Experiment (TOGA COARE) measurements combined with other observations in higher-wind regimes. Direct observations of the sensible heat flux in hurricanes supports the choice of at least the heat transfer coefficient (Zhang et al. 2008). TOGA COARE flux observations did reach to the speeds we observed, about 12 m s$^{-1}$. We are not in a situation where we have to extend the results into a high-speed regime where there are no supporting measurements. Spray is unlikely to be a significant factor in these conditions (Zhang et al. 2008). The transfer coefficients have a slight dependence on wind speed increasing from $1.08 \times 10^{-3}$ at 5 m s$^{-1}$ to $1.2 \times 10^{-3}$ at 18 m s$^{-1}$. The uncertainty of the fluxes with the COARE 3.0 algorithm is 10% for wind speeds between 10 and 18 m s$^{-1}$ (Fairall et al. 2003). GPS sonde errors of wind speed, temperature, and specific humidity could increase that uncertainty by another 25% if all the observations were in error in such a way to maximize or minimize the flux estimate. If we assume that the sensor errors are not biased then uncertainty overall is about 20%–25%.

To estimate the surface fluxes we have broken the inflow path into five ~45-km segments and determined the variables for each segment based on the matrix. Combined sensible and latent heat fluxes yield a mean value of $230 \pm 23$ W m$^{-2}$. To supply the required $6.27 \times 10^6$ J m$^{-2}$ it would take just over 7.5 h. This is within 10% of the time we estimated for the air to spiral into the center. If we were certain of the 7 h for the air to reach the center then a flux of 250 W m$^{-2}$ would be required. This is a reasonable agreement given the range of uncertainty in our estimates of the total inflow path, the fluxes, and the divergence.

Without any measurements at the top of the column we have been forced to assume that any entrainment at that location is negligible. Such entrainment would bring low $\theta_e$ into the column given the observed decrease of $\theta_e$ with height and worsen the agreement. However, this
may not be a serious omission because of the strong capping inversion that exists over the LLCC, which would inhibit mixing across the interface. The wind shear in the 1000–2000-m layer is also weak, which would also limit any mixing across the top.

The energy balance along path B was also investigated. These streamlines initially reside in the stronger wind region (20–22 m s\(^{-1}\)) of the TS. We break this pathway into two parts. In the inflow to the convective arc, from the eastern edge of the analysis (Figs. 9a,b) to where the major confluence occurs, \(\theta_e\) decreases 5–7 K demonstrating that the surface fluxes, which exceed 500 W m\(^{-2}\), are overwhelmed by the convective-scale transports. The second portion of the pathway is the outflow emanating from the arc of convection; we examined this portion of the trajectory to see if it could have wrapped into the core of the storm. The observed changes in the column energy and the fluxes balanced closely for the early portion of the trajectory just downstream of the \(\theta_e\) minimum on the northwest to the southwest side of the low-level center. At this point the streamlines could be drawn two ways starting at what we call the decision point in Fig. 9b. The air could either spiral into the core or it could continue toward the southeast and east. When we assumed that this air spiraled in from the south and entered the core starting at the decision point in Fig. 9b, the estimated fluxes from the bulk aerodynamic equations and the required fluxes based on the changes in the column over the time it would take to move from the initial to the final location revealed a major disparity. Given the speed, distance, and energy gain the surface fluxes would have to exceed 3500 W m\(^{-2}\), but the estimated fluxes were nearly an order of magnitude less (400 W m\(^{-2}\)) based on the near-surface measurements. Dolling (2010) discusses this scenario in detail demonstrating that this inflow path is unrealistic.

f. Importance of the high \(\theta_e\)

The combination of the warm and moist air collecting in the low levels below a thick dry adiabatic layer produces a remarkable and unexpected situation with very high conditional instability. The convective available potential energy (CAPE), using a parcel with the mean traits in the lowest 500 m, reached magnitudes of 2500 m\(^2\) s\(^{-2}\) in the center of the storm (Fig. 13). This calculation is based on the three-dimensional matrix, not on any one sounding. The temperature is used instead of virtual temperature, which leads to a conservative estimate of the conditional instability (Doswell and Rasmussen 1994). This is underneath an anvil and light stratiform rain, where CAPE values would be expected to be quite low. Convective inhibition energy (CIN) magnitudes exceeding 200 m\(^2\) s\(^{-2}\) prevented any deep convection from developing in the core. At 20–30 km north of the center, CIN was reduced to less than 50 m\(^2\) s\(^{-2}\). If the high \(\theta_e\) escapes from under the strong lid (see Figs. 8 and 13) then vigorous convection could break out; convective cells reaching 15-km altitude did appear about 25 km north of the low-level center (Figs. 4b,c). In situ sensors on the WP-3D recorded \(\theta_e\) exceeding 356 K in the main updraft during a pass at 3.9-km altitude through these cells, which could only

---

**FIG. 12.** Schematic of a boundary layer column along a trajectory in the presence of a vertical gradient of equivalent potential temperature and convergence. The first column displays a boundary layer that has a vertical gradient of \(\theta_e\) with no convergence. As the first column is advected to the final position (second column) on the trajectory, convergence occurs causing upward motion and the higher \(\theta_e\) near the surface fills a larger portion of the column. The difference between the middle column and the final column, which is observed at the final location on the far right, shows the energy needed from other sources for balance.
have been from the reservoir of warm moist air found in
the nascent eye below the inversion (not shown).

The large amount of conditional instability found in
the core of Humberto exceeds the typical values found
radially outward of the eyewall or on the periphery of
a mature hurricane (Bogner et al. 2000) or for any of
the typical soundings that describe the Atlantic hurricane
environment (Dunion 2011). Very high values of CAPE
theoretically yield extreme updrafts speeds, but such
magnitudes are almost never realized because of en-
trainment and water loading. Black et al. (1996) used
vertically pointing Doppler radar and Jorgensen et al.
(1985) used in situ measurements to show that the me-
dian magnitude for the maximum updrafts or cores sam-
ped in an eyewall of a mature TC are only 3–5 m s\(^{-1}\).

These modest vertical velocities are evidence of a near-
moist-neutral stability championed by Emanuel (1986)
and Rotunno and Emanuel (1987) and are consistent
with the aforementioned low CAPE values usually seen
near eyewalls and the infrequent lightning observed in
the inner core of a mature TC (Molinari et al. 1999).

Contrasting these modest vertical velocities are
Doppler and in situ measurements that do exceed
20 m s\(^{-1}\) (Black et al. 1994; Black et al. 1996; Heymsfield
et al. 2006), but these are quite rare, in the top 1% of the
log-normally distributed updrafts.

Virtually all of the aforementioned measurements are
for mature TCs, not tropical storms, save for the Chantal
(2001) analysis by Heymsfield et al. (2006). If tropical
storms have large CAPE present near their center then
one can imagine that the accelerating updraft can con-
tribute to spinup via the stretching term \([\zeta + f] \omega / \partial z\),
also known as the divergence term] in the absolute
vorticity equation. Here \(\zeta\) is relative vorticity, \(f\) is the
earth’s vorticity or Coriolis, \(\omega\) is vertical velocity, and \(z\)
height. This is probably the dominant term in the spinup
process given that the environment of a developing TC
has low vertical shear of the horizontal wind (McBride
and Zehr 1981) and density variations (solenoidal term)
are also negligible.

4. Discussion: Top down or bottom up?

There have been numerous observations of a midlevel
mesoscale convectively driven vortex (MCV) being
present at genesis either alone (Chen and Frank 1993;
Fritsch et al. 1994) or in concert with other MCVs (Harr
et al. 1996; Ritchie and Holland 1997; Simpson et al.
1997; Kieu and Zhang 2008). In the top-down scheme of
TC genesis the vorticity that develops in the MCV is
hypothesized to be transported downward in a saturated
stratiform region with heavy rain (Bister and Emanuel
1997). Any MCV ultimately owes its existence to the
deep convective cells organized in a mesoscale convec-
tive system, so top down refers to the propagation of the
circulation down to the surface. Issues with this scheme
include little evidence for a cool, saturated layer with
heavy stratiform rain under the MCV. Instead a warm,
dry layer with very light rain is typically seen under the
stratiform area and MCV (e.g., Gamache and Houze

In the bottom-up scenario a particular type of cumu-
lonimbus, the vortical hot tower (VHT), concentrates
vorticity, then the vorticity maxima from numerous
VHTs axisymmetrize (Hendricks et al. 2004; Reasor et al.
2005; Montgomery et al. 2006). Here the role of the MCV
seems to serve as a source of vorticity for the VHTs to
stretch (Montgomery et al. 2006). Braun et al. (2010) used
a numerical simulation of the genesis of Gert (2005) to
argue that the midlevel circulation in the stratiform pre-
cipitation region did contribute to spinup, but without
cumulonimbus clouds (Cbs) first concentrating vorticity
in the low levels genesis would not have occurred. The
issues with the bottom-up scenario include simulations
that contain updrafts within VHTs that are far larger than
the typical observed updrafts, and the need for the vor-
ticity produced by many such towers to merge. Models
produce high vorticity in Cbs and do merge these max-
ima, but observations of such remain elusive.

The Humberto observations lead us to suggest an-
other role for the midlevel MCV besides contributing
their vorticity. Under the MCV and stratiform rain there
is sinking air that blocks deep convection, which allows
the moist entropy to increase via the surface fluxes. This
creates a situation where the conditional instability
builds, akin to the tornado proximity sounding discussed
by Fawbush and Miller (1952). Eventually, as the subsidence weakens and or the moist entropy reaches extreme values, new convection breaks out along the edge of the subsidence region dominated by large CIN. The development of Cbs around the edge of the circulation rather than the center eliminates the problematic issue of having deep convection form in the very center of the nascent TC only later to be replaced by an unsaturated, warm, and dry layer that is typical of the upper portion of an eye (Willoughby 1998). Uncommonly large values of CAPE would favor stronger acceleration of the updraft that stretches the column and concentrates the vorticity—such a trait may be the distinguishing factor between hot towers and vortical hot towers. Here we see the mesoscale midlevel circulation aiding in the development of high CAPE, not just being a source for vorticity.

5. Conclusions

The reflectivity and wind fields of Humberto are asymmetric, similar to most tropical storms, with the rain and stronger winds to the north and east of the circulation center. In contrast to this asymmetry are the highest equivalent potential temperatures found in the storm’s center. This reservoir of high $\theta_e$ is nearly circular, well mixed to about 1-km altitude and underneath a strong inversion. The inversion appears to be the result of subsidence beneath an anvil cloud and light stratiform rain, similar to what has been witnessed frequently in the wake of a mesoscale convective system. The reservoir is created as air southeast of the center spirals inward. An energy budget demonstrates that convergence and the surface fluxes are of the magnitude to largely account for the increasing moist entropy found in the core. The high convective inhibition energy in the core, due to the sinking and resultant inversion, is crucial in that it keeps the high $\theta_e$ air from prematurely erupting into cumulonimbis. This creates a situation where conditional instability builds markedly, more than an order of magnitude higher than what has been recorded radially outward of the eyewall in previous mature hurricanes. When this air either breaks through the lid or escapes laterally an arc of deep convective clouds, 15 km high, develops about 25 km from the low-level circulation center. This arc becomes the eyewall of Humberto based on analyses of the observations acquired on the subsequent day. These convective cells that blossom in a high CAPE environment concentrate the vorticity via the stretching (divergence) term more efficiently than regular cumulonimbis and thus may be examples of vortical hot towers discussed by Hendricks et al. (2004), Reasor et al. (2005), and Montgomery et al. (2006). This may be a possible avenue for spinup given that the environment for most developing tropical storms has small vertical shear of the horizontal wind (McBride and Zehr 1981).

Both tropical storms and hurricanes have their maximum $\theta_e$ found in the eye in most circumstances, but how this occurs is subtly different. A mature TC receives the majority of its energy in the high-wind annulus under and radially outward of the eyewall first, then a small portion of this air fails to ascend and becomes trapped in the eye where it receives additional energy from the sea to later become the highest $\theta_e$ found anywhere in a hurricane. In Tropical Storm Humberto the $\theta_e$ builds in a light-wind regime. There is no annulus radically outward of the center where the strong radial gradients of surface winds and $\theta_e$ are collocated. The $\theta_e$ field associated with Humberto appears similar to those observed in other tropical storms (Molinari et al. 2004; Heymsfield et al. 2006). We speculate that the development of a high $\theta_e$ reservoir in the nascent core may be a critical step for TC formation.

Acknowledgments. The National Science Foundation Grant AGS-1042680 supported this research. The data collection efforts of NOAA/AOML/HRD, NOAA/Aircraft Operations Center, and NASA during CAMEX-IV were essential to our work. The reviews by John Molinari and another anonymous reviewer are much appreciated.

REFERENCES


