**Observations of the structure and evolution of Hurricane Edouard (2014) during intensity change. Part II: Kinematic structure and the distribution of deep convection.**

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**Abstract**

This paper is the second of two parts, both examining airborne observations of the inner-core and near environmental structure and evolution of Hurricane Edouard during several days of its lifecycle and spanning a broad range of intensities, from tropical storm to major hurricane. The focus of this paper is on the distribution of deep convection during two periods: one when there was a significant intensification as Edouard approached hurricane status on 14 Sept., and the second two days later when Edouard had reached peak intensity and was beginning to weaken. Edouard showed distinct differences in the temporal and spatial distribution of deep convection for these two periods. During the intensifying period, deep convection (echo tops above 16 km and upper-level updrafts > 5 m s-1) was noted in both the downshear left and upshear left quadrants. This convection is traced to strong updrafts that were present in the downshear right quadrant well inside the radius of maximum winds, collocated with strong convergence in the lowest 2 km. Strong updrafts persisted upshear left, where they were collocated with high inertial stability from the inner core. During the weakening period, no deep convection was present, and the precipitation that was observed was associated with weaker convergence downshear right that was located at and just inside the radius of maximum winds. Weak updrafts were seen upshear left, with little coincidence with the high inertial stability region of the inner core on this day. Some potential reasons for the differences in these spatial and temporal distributions of deep convection, including kinematic structures shown here and thermodynamic structures shown in part I, are investigated here.

**1. Introduction**

This study continues an examination of the intensity change of Hurricane Edouard (2014), a storm in which the National Oceanic and Atmospheric Administration (NOAA) WP-3D and G-IV, as well as National Aeronautics and Space Administration (NASA) unmanned Global Hawk, provide detailed measurements of the environment and inner core over several days of its lifecycle. This unique sampling offers an opportunity to investigate the environmental-, vortex-, and convective-scale processes that govern tropical cyclone (TC) intensity change. Part I focuses on Edouard’s vortex-scale thermodynamic changes, as revealed in high-altitude dropsonde observations from the Global Hawk, in relation to the precipitation evolution using data from infrared and passive microwave sensors. Part II describes the kinematic and precipitation structure in more detail during Edouard’s rapid intensification on 14 Sept. and near peak intensity after weakening begins on 16 Sept., exploring possible mechanisms that can explain the radial, azimuthal, and temporal distribution of deep convection and its influence on intensity change.

Previous studies have shown that the structure and distribution of inner-core precipitation is an important determinant of TC intensification. More specifically, satellite-based studies (e.g., Jiang 2012, Kieper and Jiang 2012; Zagrodnik and Jiang 2014; Tao and Jiang 2015) have noted that it is primarily the azimuthal distribution of shallow and moderate convection (defined by echo tops in the lower and middle troposphere) that distinguishes TCs about to undergo significant intensification. In contrast to these studies, airborne and other satellite-based studies have focused on deep convection (defined here as echo tops above 14 km collocated with strong updrafts in the upper troposphere) as a key indicator of TCs undergoing intensification (e.g., Kelley et al. 2004; Reasor et al. 2009; Guimond et al. 2010; Rogers et al. 2013b, 2015; Stevenson et al. 2014; Susca-Lopata et al. 2015). The apparent contradiction regarding the importance of shallow/moderate vs. deep convection has been attributed to the fact that the airborne datasets consist primarily of TCs already undergoing intensification, after the intensifying secondary circulation has had time to develop deep convection, while the satellite studies have a large number of cases prior to the onset of intensification, when low-level forcing mechanisms are important in the development of shallow/moderate convection (Tao and Jiang 2015).

In terms of deep convection, a preferred radial and azimuthal distribution in the inner core has been identified as more favorable for intensification. For the radial distribution, a preponderance of deep convection inside the radius of maximum winds (RMW) is favorable for intensification, since diabatic heating in the region of high inertial stability and upper-level subsidence around the periphery of the convective outflow preferentially warms the region inside the RMW (Shapiro and Willoughby 1982; Schubert and Hack 1982; Nolan et al. 2007; Vigh and Schubert 2009; Pendergrass and Willoughby 2009; Rogers 2010; Zhang and Chen 2012; Chen and Zhang 2013; Rogers et al. 2013b, 2015; Susca-Lopata et al. 2015). The azimuthal distribution of deep convection and its relationship with intensification focuses mostly on its distribution relative to the environmental vertical shear vector. A key distinction between intensifying and non-intensifying TC’s is the presence of deep convection in the upshear left (USL) quadrant (Stevenson et al. 2014; Zagrodnik and Jiang 2014; Onderlinde and Nolan 2014, 2016; Rogers et al. 2015; Chen and Gopal 2015; Alvey et al. 2015;). Deep convection in the upshear quadrants indicates a greater symmetry of diabatic heating around the storm center, which is more efficient at spinning up the vortex (Nolan et al. 2007; Jiang 2012). If the convection is associated with a midlevel vortex that is also located upshear of the low-level vortex, then advection of this midlevel feature by the mid- to upper-level flow can also lead to vortex alignment and intensification (Stevenson et al. 2014).

A variety of mechanisms have been proposed to explain the noted radial and azimuthal distributions of deep convection. Strong radial inflow in the PBL that transports high angular momentum faster than it is lost to surface friction has been identified as a key mechanism for TC intensification (Smith et al. 2009; Montgomery and Smith 2012; Montgomery et al. 2014; Sanger et al. 2014). In addition to the inward transport of angular momentum, strong PBL inflow can determine where convection is forced relative to the RMW through convergence in the PBL (Rogers et al. 2013b, 2015). In a related sense, the outer-core distribution of inertial stability may impact the radial profile of convergence, since higher inertial stability provides a greater resistance to radial displacements and a reduced mass flux to the inner core. Rogers et al. (2015) showed this relationship by comparing lower-tropospheric inertial stability between an intensifying and a steady-state TC. Convective downdrafts can modulate the buoyancy distribution and impact both the radial and azimuthal distribution of deep convection (Riemer et al. 2010; Molinari et al. 2013; Zhang et al. 2013). Additionally, the vertical profile of equivalent potential temperature and relative humidity in the local environment within which the convection occurs can play a role. Finally, differences in slope between the updraft surfaces and angular momentum surfaces, reflective of the tangential wind field and intensity of the convection, helps determine whether or not the peak diabatic heating in the convection remains inside the high-inertial stability environment within the RMW (Hazelton et al. 2015; Rogers et al. 2015).

The structure and evolution of the thermodynamic and kinematic fields of TCs in sheared environments clearly play a role in governing the distribution of inner-core deep convection and the intensity evolution. Observations of these relationships, particularly over deep layers of the troposphere, have been limited, however. Halverson et al. (2006) showed a comprehensive analysis of the warm core structure of Hurricane Erin (2001), using primarily high-altitude dropsondes to assess the structure and evolution of the vortex in the presence of vertical shear. Many of the structures mentioned above, including the distribution of deep convection, are found in this paper, however, this data only spans one day and does not include information on the three-dimensional kinematic structure that can be provided by airborne Doppler measurements.

Part I of this two-part case study provided a synoptic overview of Edouard and related the vortex-scale thermodynamic changes Edouard experienced to the precipitation evolution described using data from infrared and passive microwave sensors. In addition to mean thermodynamic profiles from each Global Hawk flight, which emphasized the importance of moistening and humidification of the middle troposphere in the initially drier upshear regions of the inner core, a detailed analysis was also provided for two typically under-sampled periods; one early when Edouard was slowly intensifying as a weak tropical storm, and the other whenEdouard was rapidly dissipating under high vertical wind shear and relative low sea surface temperatures. Part II of this study describes a 2-day period (between those two periods analyzed in Part I), in which Edouard was intensifying to peak intensity and another day when it was beginning to weaken from that peak. Using the airborne Doppler radar from the NOAA P-3 aircraft, combined with Global Hawk and P-3 dropsondes, this part focuses on the kinematic structure and distribution of deep convection, as well as the mechanisms that dictate its radial and azimuthal distribution.

**2. Aircraft sampling and datasets used**

As discussed in Part I, multiple NOAA and NASA aircraft flew missions within and around Edouard between 12 and 19 Sept. Part II focuses on the coincident AV6/NOAA43 missions on 14 and 16 Sept., when Edouard was steadily intensifying (14 Sept.) and beginning to weaken after peak intensity (16 Sept.). Figure 1 shows the flight tracks of the aircraft sampling Edouard during these two days. On 14 Sept. NOAA43 flew a figure-4 pattern followed by a series of eyewall penetrations on the northeast side of the inner core (Uhlhorn et al. 2015). AV6 flew a combination of a lawnmower survey pattern during the first half of its mission, followed by a butterfly-type storm-relative pattern during the second half of its mission. On 16 Sept. NOAA43 flew a repeated series of eyewall passes, while NOAA42 flew a partial figure-4 pattern followed by a series of spirals while deploying the low-altitude Coyote unmanned aerial system (Uhlhorn et al. 2015). Only NOAA43 is considered for this day. During this time AV6 flew a rotating butterfly pattern for its entire mission, providing extensive azimuthal coverage with its ~500 km radial legs centered on the storm.

In addition to the dropsondes in Part I, Part II features three-dimensional analyses of wind and reflectivity from the airborne Doppler radar on the P-3. The analyses use a variational method (Gamache 1997; Reasor et al. 2009) to obtain grids with a 2 km x 2 km horizontal and a 0.5 km vertical spacing. The individual radial passes, which generally take ~1 h to complete, are merged into a single analysis that represents the conditions of the vortex over the time scale of the flight pattern (~3 h for these missions). The merged analyses are sufficient for the depiction of more slowly-evolving vortex-scale parameters that require greater spatial coverage, e.g., azimuthally-averaged fields and horizontal flow. For the smaller-scale and more rapidly-evolving convective-scale fields, individual radial passes are used. A similar analysis approach was followed in Rogers et al. (2012, 2013b, 2015).

**3. Vortex-scale structure and distribution of deep convection**

Figure 2 shows the Doppler-derived storm-relative winds at 2 and 8 km altitude for the NOAA43 missions on 14 and 16 Sept. The 2-km RMW is ~30 km on both days. The intensification of the storm between the two days is evident, with peak winds increasing from 35 m s-1 on 14 Sept. and expanding to a broader area of > 40 m s-1 winds on 16 Sept. A visual inspection of the 8 km flow on 14 Sept. indicates that, unlike on 12 Sept. (cf. Part I), the displacement between the lower and upper levels is greatly reduced. However, there is still an indication of a ~3 km displacement at 8 km toward the northeast on this day. The 8-km flow on 16 Sept. also indicates a displacement, this time toward the northwest with a magnitude of ~5 km. While these values may seem small, they are nonetheless ~10-15% of the RMW, which can be significant in terms of establishing asymmetries in radial flow fields. Such a tilt amplitude would likely not prevent intensification, however, and Edouard steadily intensified on 14 Sept. in the presence of light to moderate (~ 6 m s-1) shear, reaching peak intensity by 16 Sept. (cf. Fig. 2 from Part I).

A plot of reflectivity from the lower fuselage (LF) radar on NOAA43 for both 14 and 16 Sept. is shown in Fig. 3. Both days contain a maximum in reflectivity on the southwest side of the eyewall, however, 14 Sept. has a more widespread distribution of reflectivity at other locations around the storm. On the contrary, the eastern eyewall is essentially devoid of precipitation on 16 Sept (note that the apparent reflectivity on the north, northeast, and east side of the image in Fig. 3c is sea clutter and is thus spurious). With respect to shear orientation, the location of the maximum LF reflectivity is primarily found to the left of the shear vector both days (cf. Fig. 3b,d). Most of the high reflectivity is located in the downshear left (DSL) quadrant, but some precipitation is also found upshear left (USL), particularly on 14 Sept.

While the LF images provide information on the horizontal distribution of precipitation, they do not provide any information on its vertical structure, an important component to infer locations of deep convection. Figure 4 shows plots of the echo top (defined as the height of the 20 dBZ surface) and peak vertical velocity in the 8-16 km layer from individual passes of the tail Doppler radar on 14 and 16 Sept. On 14 Sept., a ~25 km x 25 km area of deep convection in the SW side of the eyewall (DSL and USL) has maximum echo tops exceeding 16 km. Elsewhere in the domain the only other echoes reaching 10 km altitude are a small area ~80-100 km from the center on the west side of the storm. The strongest upward motion in the upper troposphere approaches 8 m s-1, coincident with the highest echo tops. In contrast, the pass on 16 Sept. (Fig. 4b) shows that the maximum echo tops are only 12 km. A very small region of echo tops > 10 km is observed on the SW (DSL, USL) side within an otherwise broad region of echo tops between 6 and 8 km. Consistent with this reduced echo top height, the peak updrafts during the pass are < 5 m s-1, with only limited, isolated areas exceeding 3 m s-1.

The center passes shown in Fig. 4 clearly demonstrate differences in the distribution of deep convection on 14 and 16 Sept., with a larger amount of higher echo tops and stronger updrafts on 14 Sept. compared with 16 Sept. However, Fig. 4 only shows one pass for each flight. A more comprehensive depiction of the distribution of deep convection is shown in Fig. 5, which presents locations in the Doppler analysis where the peak 8-16 km vertical velocity exceeds 3 and 5 m s-1 for all center passes during the 14 and 16 Sept. missions, normalized by the RMW at 2-km altitude. A broad region of updrafts > 3 and 5 m s-1, generally at and inside the low-level RMW, is concentrated in the DSL and (especially) the USL quadrants on 14 Sept. A significant amount of the strongest updrafts are located in the 0.5-0.75 x RMW band. Such a radial distribution of deep convection is consistent with intensifying TCs as shown in Rogers et al. (2013b, 2015), Stevenson et al. (2014), and Susca-Lopata et al. (2015). Contrary to 14 Sept., 16 Sept. has a much smaller region of grid points > 3 m s-1, primarily found in the DSR and DSL quadrants, with a very small area found USL. For the most part these updrafts are found at and just inside the low-level RMW; however, no updrafts > 5 m s-1 are observed, a result consistent with the lack of high echo tops for 16 Sept. shown in Fig. 4.

As Figs. 3-5 indicate, the majority of the precipitation and deep convection is located in the DSL and USL quadrants for both days. However, the precipitation can be traced back to the shear-relative quadrants upwind. For example, Fig. 6 shows LF images at ~10-25 minute intervals for a ~1.5 h period on 14 Sept. The letters “A” and “B” denote trackable locations of reflectivity cores (or groups of cores) from LF animations. Beginning at 1628 UTC, a group of three reflectivity cores (“A”) in the DSR quadrant (cf. Figs. 3,5), translate into the DSL quadrant over ~30 minutes, at which point they merge into a single large core by 1700. At the same time another small core develops in the upwind portion of the downshear right (DSR) quadrant (“B”). This core is also tracked around the storm, reaching its largest size by 1722 UTC as it passes into the DSL quadrant (the same time that “A” has begun to dissipate). The “B” core remains coherent for the next 20-25 minutes, where it is clearly seen in the DSL and USL quadrants by 1745 UTC. This sequence of events describes a typical evolution of the reflectivity cores – initiation in the DSR quadrant, consolidation in the DSL quadrant, and maturation (and in some cases dissipation) in the USL quadrant – a result also found in Black et al. (2002) and DeHart et al. (2014). This azimuthal evolution of the reflectivity cores is similar on 16 Sept. (not shown); however, the vertical structure of the precipitation, as manifested by the radial and azimuthal distribution of deep convection (cf. Figs. 4-5), reveals differences between the two periods. Specifically, the presence of deep convection on 14 Sept., including in the USL quadrant, distinguishes that intensification period from the weakening period on 16 Sept., when there was limited deep convection DSL, and a lack of convection USL.

**4. Mechanisms underlying distribution of deep convection**

*a) Azimuthal distribution*

This section will discuss some mechanisms controlling the distribution of deep convection and the likely impact of the distribution on the resultant intensity evolution of Edouard. Figure 7 shows azimuth-height plots of the storm-relative radial flow averaged in a 20-km radial band centered on the low-level RMW for these two days. Considerable azimuthal wavenumber-1 asymmetries of radial flow are seen on both days, likely driven by a combination of storm motion and vertical shear (Zhang and Uhlhorn 2012) and the vortex tilt toward the northeast (14 Sept.) and northwest (16 Sept.; cf. Fig. 2). On 14 Sept. a deep layer of radial inflow is found on the north and west side of the storm. This location of peak inflow is roughly in the forward quadrant relative to storm motion, and it covers both the DSL and DSR quadrants, with strongest low-level inflow DSR. Above 4 km inflow continues, but it begins to rotate around the storm downwind, through the DSL and into the USL quadrant by 6-8 km altitude. This inflow aloft may be related to the displacement of the vortex toward the northeast at 8 km altitude (cf. Fig. 2), and it is likely also related to the deep convection shown in Figs. 4-5. Above 10 km strong outflow is seen in the USL and USR quadrants, reaching a peak magnitude at 14-16 km. This feature is again likely tied to the deep convection seen in Figs. 4-5. The azimuthal structure of the inflow on 16 Sept. has similarities to that on 14 Sept. in the lower troposphere, with peak inflow located in the forward and DSR quadrants. By contrast to 14 Sept., however, some low-level inflow extends into the USR quadrant. This could reflect the fact that the shear and motion vectors are orthogonal to one another on 16 Sept., whereas the vectors are parallel on 14 Sept. Forcing of the inflow from storm motion on the forward side (16 Sept.) may be enough to counteract the tendency of shear induced inflow DSL and USL. Above 8 km there is an indication of outflow on the north side consistent with a tilt toward the northwest on this day. However, the strength of the radial flow is reduced, likely because of the lack of significant deep convection on this day. Despite these differences between the two days, the pattern of low-level inflow at the eyewall maximized DSR is generally consistent with past studies involving radar, dropsonde, flight-level, and surface wind measurements (e.g., Black et al. 2002, Reasor et al. 2013, Zhang et al. 2013). With the strongest, deepest inflow, it is this quadrant that is the source quadrant for precipitation in the eyewall, including the deep convection on 14 Sept.

The azimuthal variation of eyewall radial flow shown in Fig. 7 suggests a kinematic forcing for the observed azimuthal distribution of precipitation, including deep convection. Additional, or complementary, mechanisms for this azimuthal distribution can be identified based on the thermodynamic environment (where “environment” is defined as the innermost 200 km) described in part I. For example, Fig. 12 from part I shows that, on 14 Sept., there is a distinct maximum in equivalent potential temperature (θE) to the right of the shear vector, maximized in the DSR quadrant, and a minimum to the left of the shear vector. Such an azimuthal distribution of low θE air is consistent with downdrafts from precipitation and deep convection at and inside the low-level RMW in the DSL and USL quadrants (cf. Fig. 6 from part I, Figs. 3-6 here) cooling and drying the PBL (cf. Figs. 9, 11, 12 from part I). With sea-surface temperature (SST) values generally > 27⁰C on 14 Sept. (cf. Fig 4 from part I), the downdraft-driven low-θE air can be recharged by surface fluxes as it travels through the USL and USR quadrants (Molinari et al. 2013, Zhang et al. 2013). Once the air reaches the DSR quadrant, θE has fully recovered, as revealed by the θE profiles and CAPE values DSR (cf. Figs. 9 and 10 from Part I). That instability, combined with the radial inflow at the eyewall maximized DSR, are favorable conditions for the development of deep convection seen in the DSL and USL quadrants.

On 16 Sept., by contrast, there is a region of pronounced cooling of SST (< 24⁰C) in the southeast part of the domain (cf. Fig. 4 from part I), along the right rear portion of the track, in the USL and USR quadrants. These low SST values prevent the recharge of downdraft-cooled air by surface fluxes. This results in low θE values in the lowest 500 m DSR and DSL, as shown in Fig. 9 from part I. CAPE values on this day are also nearly uniformly low, as nearly all dropsondes DSR and DSL have CAPE values < 500 J kg-1 (cf. Fig. 11 from Part I). Despite this surface-based stabilization on 16 Sept., θE profiles still show some instability in the lower troposphere between 900 and 600 hPa (cf. Fig. 9 from part I). Precipitation is thus observed DSL and USL, but no deep convection is present.

*b) Radial distribution*

Some potential mechanisms underlying the observed radial distribution are considered next. Figure 8 shows the distribution of dropsondes released from NOAA42, NOAA43, and AV6 within a 12-h window from 12 UTC 16 Sept. to 00 UTC 17 Sept., after Edouard had reached peak intensity and was weakening. These sondes are used to calculate planetary boundary layer (PBL) kinematic properties shown in Fig. 9, similar to an analysis done in Hurricane Earl, which was rapidly intensifying at the time (2010; Montgomery et al. 2014; Rogers et al. 2015). The largest number of sondes are released inside 50 km radius, at and near the RMW. However, a fairly large number of sondes are also released at outer radii (Fig. 8b). This radial distribution of sondes is comparable to that shown in the analyses of Hurricane Earl’s PBL structure. Figure 9 shows radius-height profiles of azimuthally-averaged radial flow, agradient wind wind (defined as the difference between the total wind and the gradient flow), and divergence on 16 Sept. Similar to Earl (Rogers et al. 2015), inflow is maximized in the lowest few hundred meters. Both Earl and Edouard produced regions of significant supergradient flow at and inside the RMW, above 100 m altitude. Unlike Earl, however, the radial location of the peak inflow is 2-2.5 x RMW for Edouard, whereas for Earl the location was at ~1.5-2 x RMW. Furthermore, the radial location of the strongest convergence in Edouard is at ~1.5 x RMW, whereas for Earl that peak was located inside the RMW. The radial location of the peak convergence outside the RMW for Edouard lends support to the idea that this PBL convergence is a key determinant of the radial location of precipitation and deep convection, and may help to explain why most of the strongest updrafts on 16 Sept. shown in Fig. 5, while still primarily located at and inside the low-level RMW, are located radially outward from those on 14 Sept. Unfortunately, there was not an adequate distribution of dropsondes outside the RMW to perform a similar PBL calculation on 14 Sept.

Figure 10 examines the radial distribution of lower tropospheric kinematic forcing from the airborne Doppler radar1 within the DSR quadrant, in order to focus on the forcing mechanisms in the quadrant where convection is primarily initiated (cf. Figs. 6,7, Black et al.

2002; Reasor et al. 2013; DeHart et al. 2014). The top panel (Figs. 10a,b) shows a radius-height plot of radial and tangential velocity in the DSR quadrant for both 14 and 16 Sept. On 14 Sept., inflow extends inward to ~15 km radius, which is well inside the RMW in that quadrant. By contrast, on 16 Sept. the inflow only reaches ~30 km radius, essentiallystopping at the RMW. The divergence and vertical velocity fields in Figs. 10c,d, as expected, show convergence maximized well inside the RMW on 14 Sept., near ~10-15 km. This peak convergence forces an area of upward motion greater than 1.5 m s-1, centered at 10 km radius. On 16 Sept. the peak convergence has shifted radially outward, covering a radial band 30-40 km from the center. The strongest upward motion is of comparable magnitude to 14 Sept., but again is shifted outward to a radius of ~30 km. The inertial stability shown in Figs. 10e,f is higher on 16 Sept. than on 14 Sept. For example, at r=40 km and z=3 km (outside the RMW on both days), the inertial stability is nearly twice as large on 16 Sept. than on 14 Sept.

The mechanism to explain the differences in the radial profile of convergence and ascent from the PBL on the two days examined here remains elusive. Differences in inertial stability could be one explanation, since from a balanced perspective inertial

stability provides a greater resistance to radial displacements. A similar relationship between outer-core inertial stability and radial flow was seen in Rogers et al. (2013b; 2015). This explanation is likely valid above the frictional boundary layer (~2 km altitude), where the flow is in approximate gradient wind balance. However, within the frictional boundary layer, where gradient wind balance is not satisfied, the role of inertial stability in constraining inflow is likely invalid (e.g., Smith et al. 2015, Kilroy et al. 2015). By contrast, Miyamoto and Takemi (2015) emphasize the importance of the vortex Rossby number in determining the radial location of peak PBL convergence. They show that as storms intensify the radius of peak convergence in the PBL shifts radially outward as the Rossby number (and inertial stability) increases. This relationship may explain why the PBL convergence is farther from the center on 16 Sept. than on 14 Sept. In this sense increased inertial stability in the PBL could serve as a self-limiting mechanism, essentially moving the radial location of peak convergence to a larger distance from the center. The linkage between the TC PBL and convection is an area of active research, and the results shown here provide an important observational benchmark for these studies. *c) Impact of spatial and temporal distributions on intensity evolution*

The quadrant-averaged distribution of vertical velocity and inertial stability are shown in the DSL and USL quadrants for the merged analyses for 14 and 16 Sept. (Fig. 11). These quadrants are located downwind of where convection is primarily initiated. On 14 Sept., an area of upward motion is seen at 4 km altitude at a radius of 15 km in the DSL quadrant (Fig. 11b). Further aloft this upward motion exceeds 2.5 m s-1 (in a quadrant-averaged sense) inside the local RMW at that altitude. This strong upward motion also coincides with high inertial stability in the inner core. Continuing downwind into the USL quadrant (Fig. 11a), the strong upward motion continues above 10 km altitude, however, there is now also a broad region of subsidence below. Such a distribution of upward motion in the quadrant average is consistent with the presence of deep convection in this quadrant (cf. Figs. 4,5). Additionally, the strong upward motion above 10 km USL inside the RMW is coincident with high inertial stability, similar to the DSL quadrant. On 16 Sept., there is an area of upward motion inside the local RMW in the DSL quadrant. However, this upward motion is weaker than on 14 Sept., and there is less overlap with the region of high inertial stability in this quadrant. Looking at the USL quadrant on 16 Sept., a small area of weak upward motion is again consistent with the lack of deep convection here (cf. Figs. 4,5).

The intensification of Edouard on 14 Sept. may be tied to the fact that de strong is co-located with high inertial stability inside the RMW in the DSL and USL quadrants. Such a distribution of heating on 16 Sept. An alternate interpretation of the importance of the radial location of diabatic heating and vortex spin-up is offered in Smith et al. (2015), Kilroy et al. (2015), and Smith and Montgomery (2016). They make the point that, rather than making arguments regarding heating efficiency, the role of deep convection can be more succinctly understood simply by considering that angular momentum surfaces are advected inward (outward) in the frictional boundary layer when deep convection occurs inside (outside) the RMW. The role of the PBL in TC intensification in this context is thus more direct, rather than playing a more indirect role via convergence forcing deep convection, which then may or may not intensify the vortex depending on the location of the deep convection relative to the high inertial stability region of the inner core. Regardless of which role the PBL plays in intensification, i.e., a direct role involving advection of angular momentum surfaces in the frictional boundary layer vs. an indirect role involving forcing of deep convection via PBL convergence, both mechanisms require a favorable location of deep convection inside the RMW.

**5. Discussion and concluding remarks**

The datasets analyzed in this two-part study, which include high-altitude dropsondes from the Global Hawk and airborne Doppler radar and dropsondes from the WP-3D, provide a unique opportunity to describe the thermodynamic and kinematic structure of the inner core and near environment of Edouard as it underwent several different stages of its intensity evolution. Despite the uniqueness of the datasets described here, there are areas for further inquiry, as well as a need for even better datasets. For example, it is still not clear from this study whether it is the distribution of deep or shallow/moderate convection that is more important for intensification. The results of this study seem to suggest that they may play an equally important role, since both experience an increase in occurrence upshear during intensification. One question to consider is whether the forcing mechanisms for deep convection discussed here, e.g., PBL convergence, outer-core inertial stability, also play a role in governing both the radial and azimuthal distribution of shallow/moderate convection. Another question to consider is other modes of precipitation, such as stratiform precipitation. For example, the vertical velocity in the upshear left quadrant on 14 September indicates a clear region of deep-layer subsidence below 8 km, overlain by updrafts above 10 km. Is that stratiform precipitation, or does it simply reflect the azimuthal trajectory of strong updrafts initiated downshear right?

Additional datasets are needed to answer these questions. Combined satellite and aircraft datasets, nearly coincident in space and concurrent in time, can provide a broader context for the vortex and precipitation structures, and their relationship with intensity change, vertical shear, and storm motion. Numerical models can also help to address these questions. Trajectory calculations can be made to identify the source regions of the precipitation and better characterize the precipitation as deep convection, shallow/moderate convection, stratiform, etc. One important question to consider when using numerical models is how well the models represent the structures and processes, e.g., PBL, microphysics, air-sea interaction, that are important in intensity change. To perform these evaluations new datasets are also needed. In particular, PBL measurements from dropsondes and low-level unmanned systems (Cione et al. 2015; Uhlhorn et al. 2015) are needed, as are precipitation and microphysics measurements from airborne and ground-based radar and satellites. Polarimetric radar, whether ground-based (Melnikov et al. 2011) or airborne (Bluestein et al. 2014), is another promising dataset for microphysical evaluations. Ocean measurements from airborne expendables and dropsondes capable of measuring the sea-surface temperature along with air temperature (Uhlhorn et al. 2015) can provide measurements at the air-sea interface. Finally, aircraft missions that specifically target sheared storms are needed to sample these fields and augment the existing database of measurements.

For the Edouard analysis shown here, a common aspect of the discussion in both parts I and II pertains to the role of symmetry in Edouard’s intensification. From these observations, symmetry applies to both the thermodynamic and precipitation structure, spanning a broad domain within and around the storm. Precipitation as indicated by the passive microwave satellite data showed a distinct tendency to transition from a less symmetric state during the slowly-intensifying stage on 12 September, to a more symmetric state during the near-rapid intensification on 14 September, to a period of maximum symmetry once the storm had reached peak intensity on 16 September. A similar evolution in symmetry was seen in the profiles of relative humidity and θE between 12–16 September. The notion of symmetry can be defined relative to the shear vector, which was of moderate intensity from the southwest (on 12 September) to the southeast (on 14–16 September). Within this framework, changes in symmetry are most closely tied to changes in the upshear quadrants on these days; i.e., the most significant changes in humidity, θE, precipitation (broadly speaking), and deep convection (specifically) were seen upshear. These observations lead to the first hypothesis to arise from this work:

*Hypothesis 1:*

* *Increased symmetry of precipitation and deep convection, particularly when present in the upshear quadrants, contributes to a greater azimuthal coverage of diabatic heating and is a favorable configuration for vortex spin-up.*

Several possible mechanisms can modulate the distribution of precipitation upshear. Entrainment of dry air into the inner core can impact the strength of convection, either by forcing downdrafts through evaporation in the lower troposphere or reducing updraft strength through detrainment in the middle/upper troposphere. The source of the dry air upshear in the case of Edouard did not appear to be from the environment, as the local environment did not have any pronounced regions of dry air. Rather, dry air appeared to be driven by subsidence, especially during the first few days Edouard was sampled. Differences in the azimuthal distributions of CAPE can also modulate the precipitation distribution. CAPE, as well as vertical profiles of θE, showed that the most unstable quadrant was the downshear right quadrant on 14 September. While humidity distributions can certainly impact stability through entrainment processes, another key element is the lower boundary condition. In the case of Edouard, the presence of warm water around the storm on 14 September likely played a significant role in creating an unstable environment downshear right, and, when combined with the maximum eyewall radial inflow in that quadrant, provided a favorable environment for the initiation and maintenance of deep convection downwind (i.e., downshear left and upshear left). The sea-surface cooling seen two days later, partially induced by Edouard itself, likewise likely played a role in limiting the destabilization downshear right, which is why there was no deep convection seen on that day. The development of the upper-level warm core also played a role in constraining CAPE, especially as the warm core developed later in the storm’s lifecycle. Tilt of the vortex can drive asymmetries in the low-level convergence at the eyewall (Reasor et al. 2009, Reasor and Eastin 2012, Reasor et al. 2013). Tilt has also been shown to drive low-level convergence outside the eyewall on the downshear side, leading to downdraft cooling at large radii and weakening the precipitation at the eyewall (e.g., Riemer et al. 2010). Displacement of the circulation center thus can clearly hinder intensification. While Edouard did show some tilt (at least on 14-16 Sept.; cf. Fig. 2), tilt is not considered to play a significant role in modulating the intensity change. Finally, environmental helicity (Onderlinde and Nolan 2014, 2016) has been shown to be a parameter that can control the longevity and robustness of convection in tropical cyclones, particularly upshear. Onderlinde and Nolan (2016) related PBL recovery to storm-relative helicity, showing that the boundary layer downwind of deep convection recovered more quickly for storms encountering positive helicity. This was not investigated here, but future work should include a consideration of helicity in modulating the azimuthal distribution of precipitation.

The radial distribution of deep convection and its relationship with the intensity evolution of Edouard is consistent with previous studies (e.g., Rogers et al. 2013b, 2015; Susca-Lopata et al. 2015). The radial location of peak PBL convergence while Edouard was weakening on 16 Sept, and its comparison with the radial location in Hurricane Earl (2010) when it was intensifying, supports the idea that PBL convergence is a key forcing mechanism governing the radial distribution of deep convection and the subsequent vortex response. Additionally, by considering the PBL convergence and inertial stability in a shear-relative framework, the radial distribution of deep convection can be examined within the context of the azimuthal variations. Such an analysis suggests that it may be necessary to focus on forcing mechanisms in specific quadrants, such as the downshear right quadrant.

To the extent that TC intensification is driven by precipitation processes, a useful framework for interpreting the role of precipitation processes is to consider the spatial and temporal structure and evolution of both the forcing mechanisms and the local environment of the precipitation. Kinematic properties such as PBL convergence and outer-core inertial stability, and thermodynamic properties such as low-level destabilization through surface fluxes, provide the forcing mechanism for precipitation. Thermodynamic fields like relative humidity and the upper-level warm core, as well as kinematic fields such as inner-core inertial stability and angular momentum distributions, provide the local environment that supports (or suppresses) precipitation and governs the response of the vortex to the diabatic heating from the precipitation. These two considerations – the forcing mechanisms and local environment of the precipitation – control the radial and azimuthal distribution of precipitation, which determines whether or not the TC will intensify. Both of these considerations are strongly dependent on the shear, as well as the vortex response to the shear. This leads to the second hypothesis:

*Hypothesis 2:*

* *Storms that are intensifying are characterized by a maximum in boundary-layer convergence in the downshear-right quadrant that is inside the local radius of maximum winds, precipitation and deep convection downshear and upshear left, and diabatic heating co-located with high inertial stability inside the local radius of maximum wind, particularly upshear.*

Thus, the example of Edouard confirms that dry air and asymmetrical precipitation distribution, both closely related to shear, can hinder intensification. But this is already well known. Perhaps is time to propose (as suggested by Kowch and Emanuel 2015) that there is nothing special about “rapid” intensification, and to stop asking the question “why do tropical cyclones intensify?”. This leads to the final hypothesis:

*Hypothesis 3:*

* *All tropical cyclones over a warm ocean will intensify, perhaps toward their maximum potential intensity, unless hindered by vertical shear, dry air, and/or asymmetrical precipitation distribution.*

The challenge of course is to identify where and when these negative influences (both environmental and vortex-scale) will be significant enough to hinder intensification.

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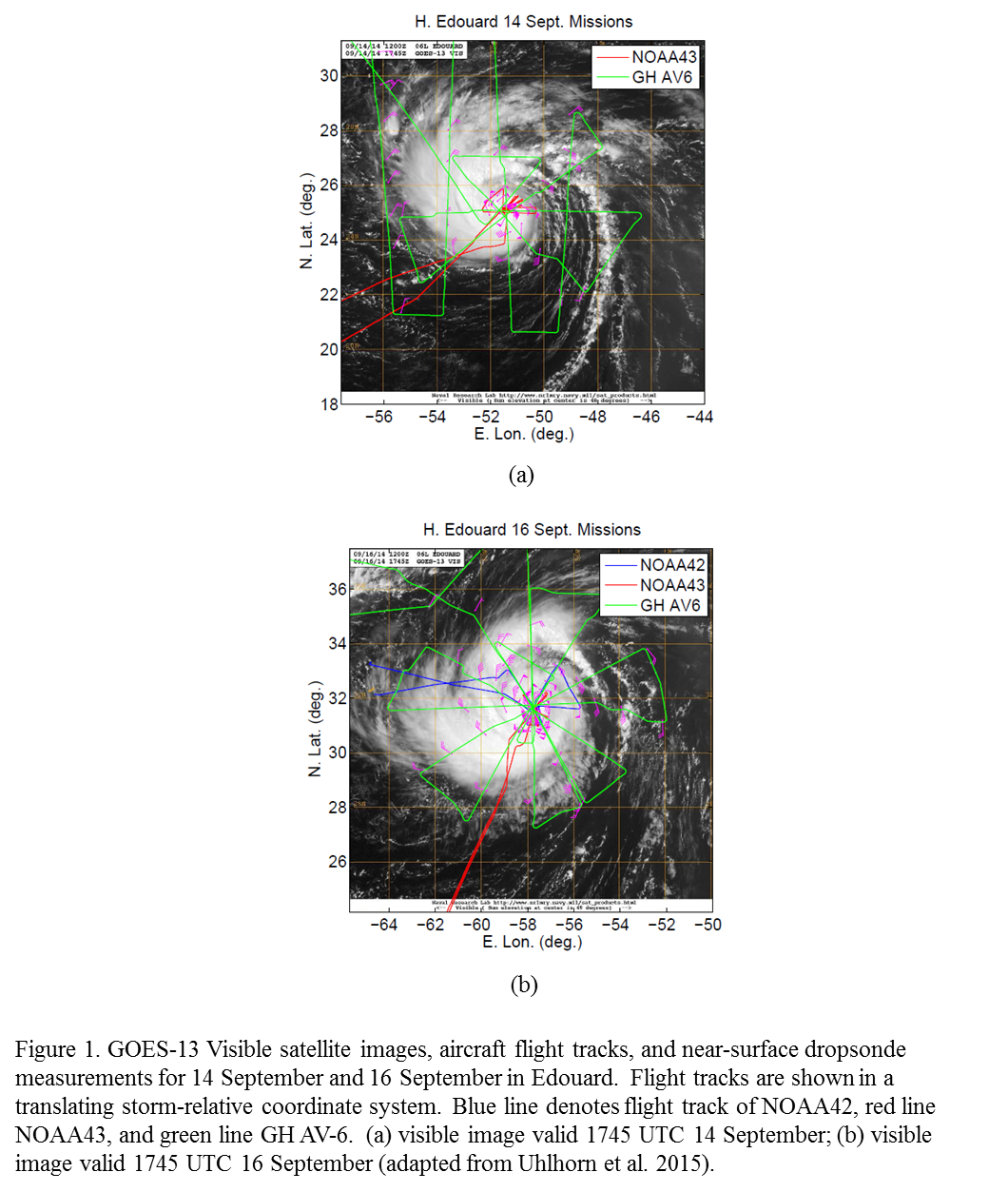
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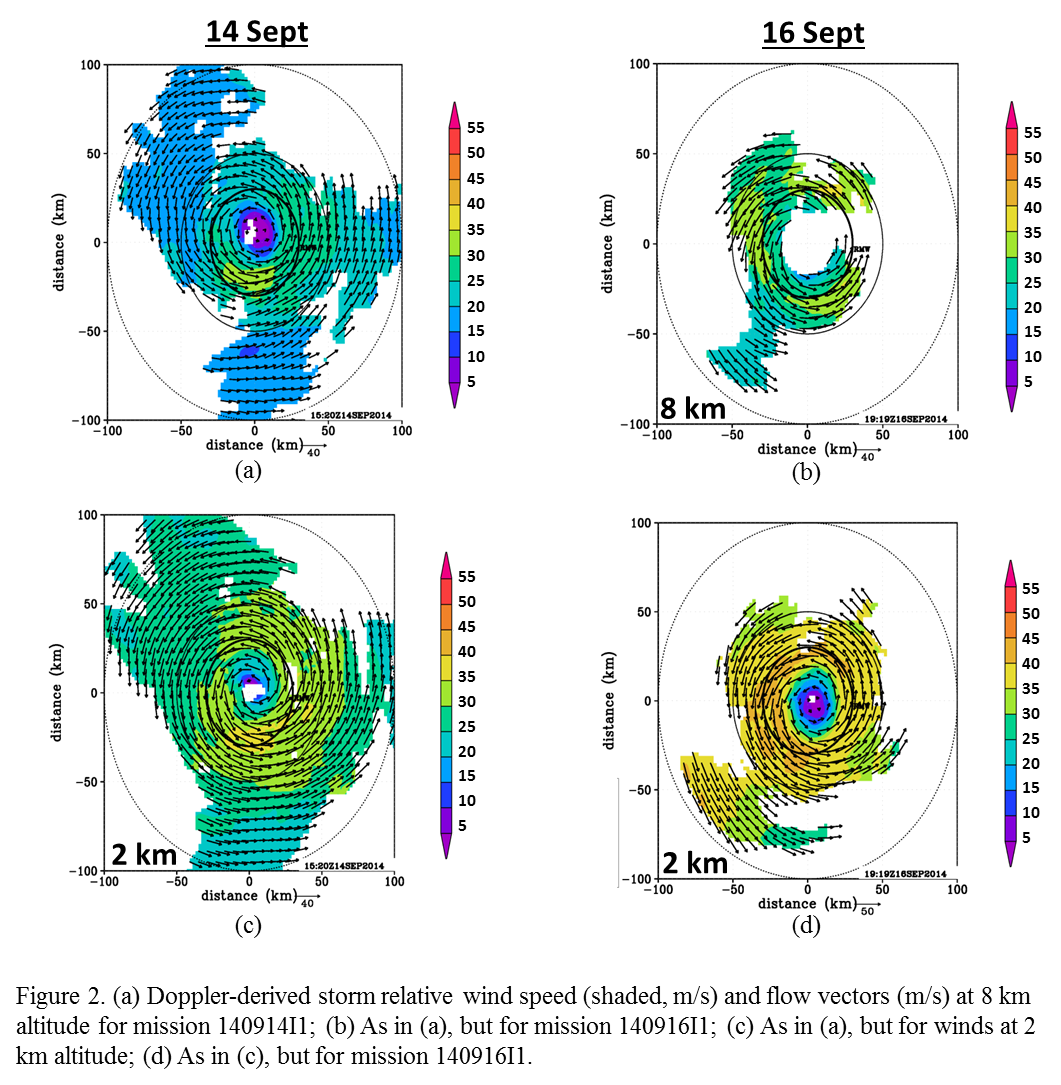
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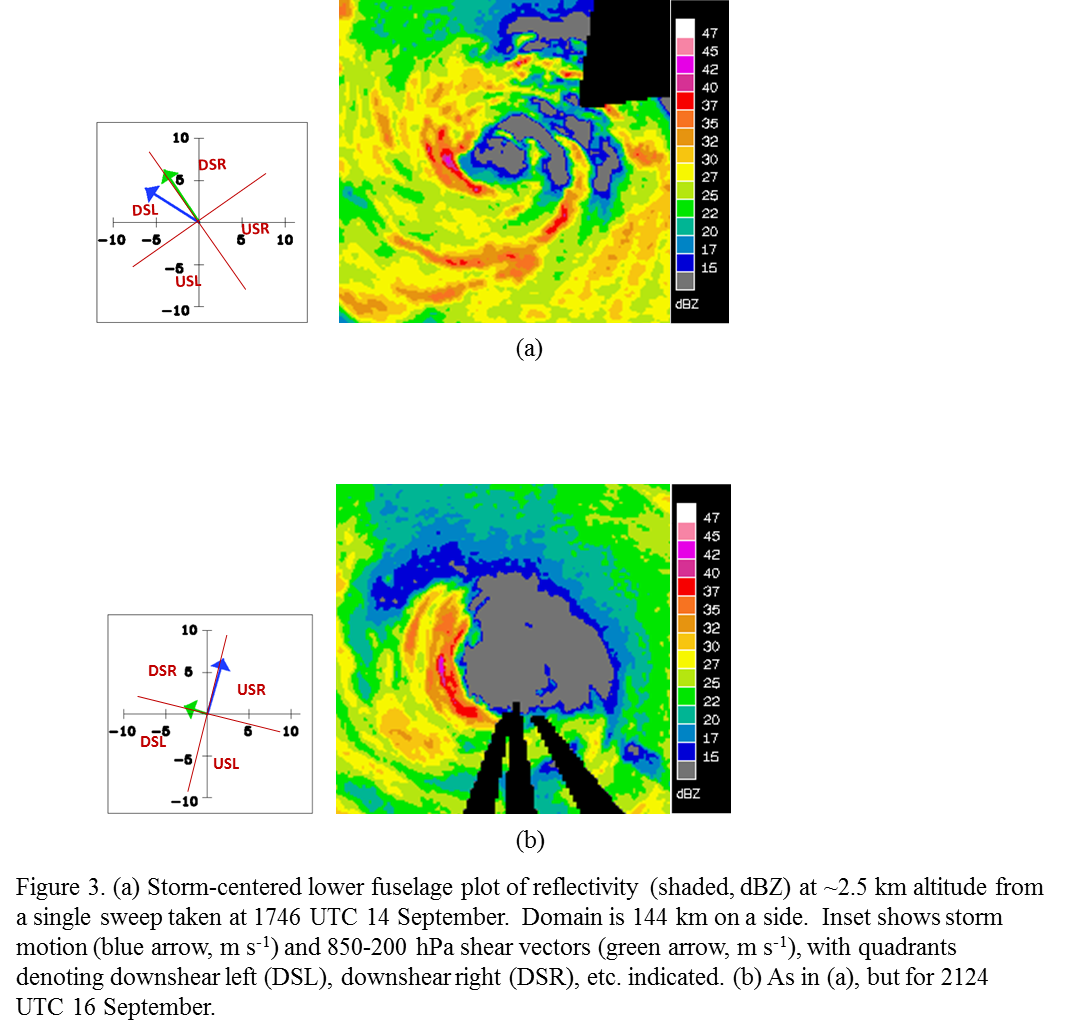
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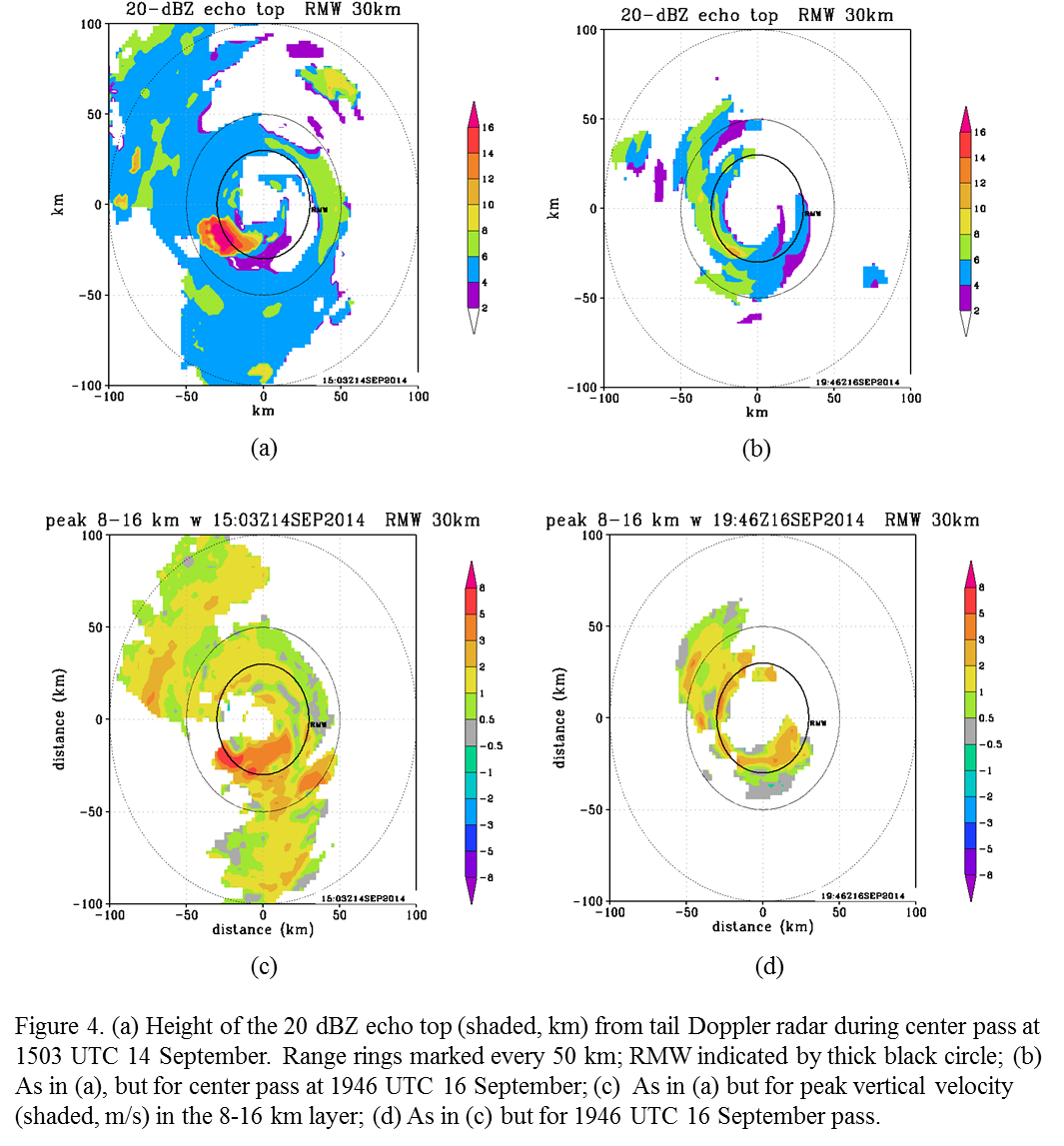
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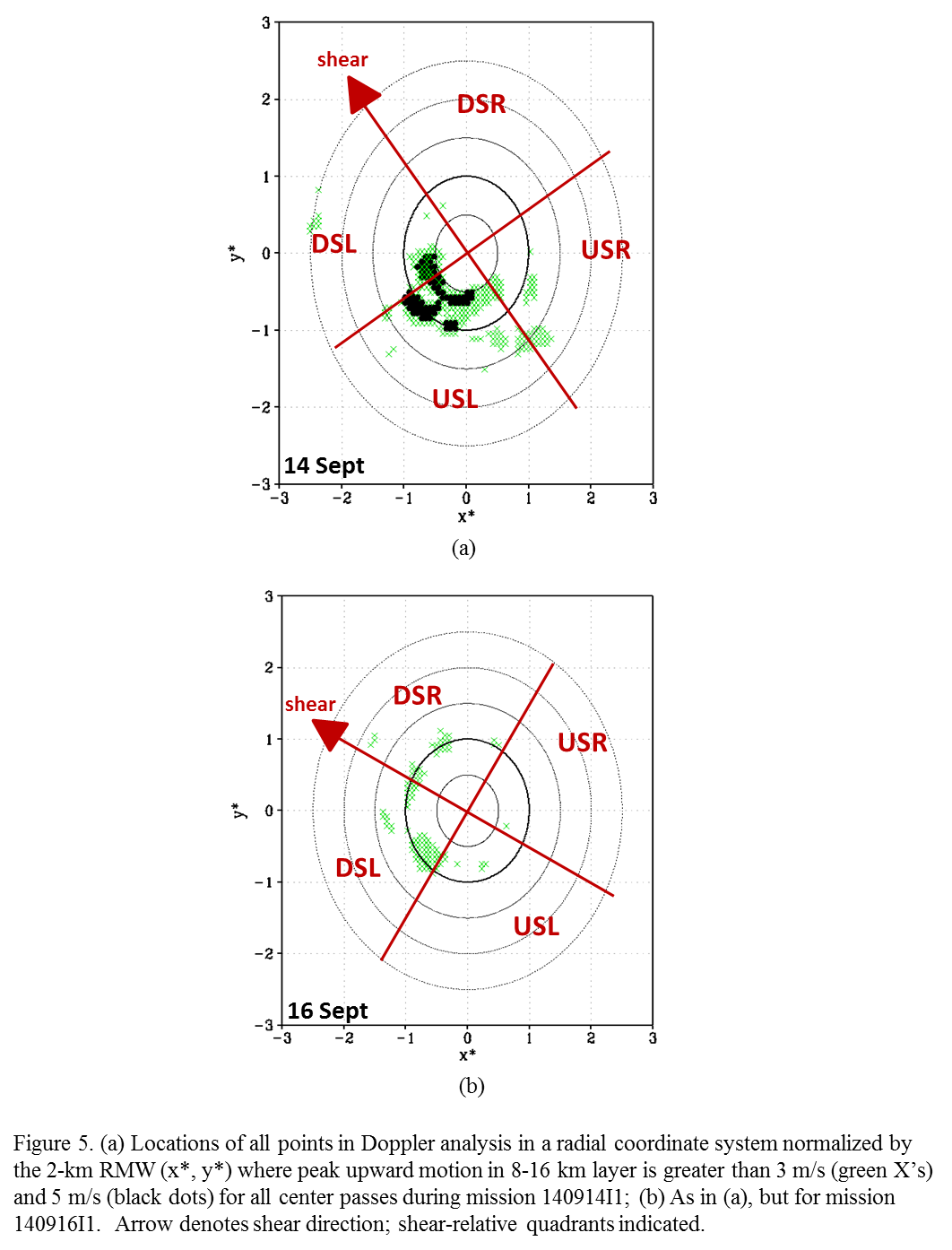
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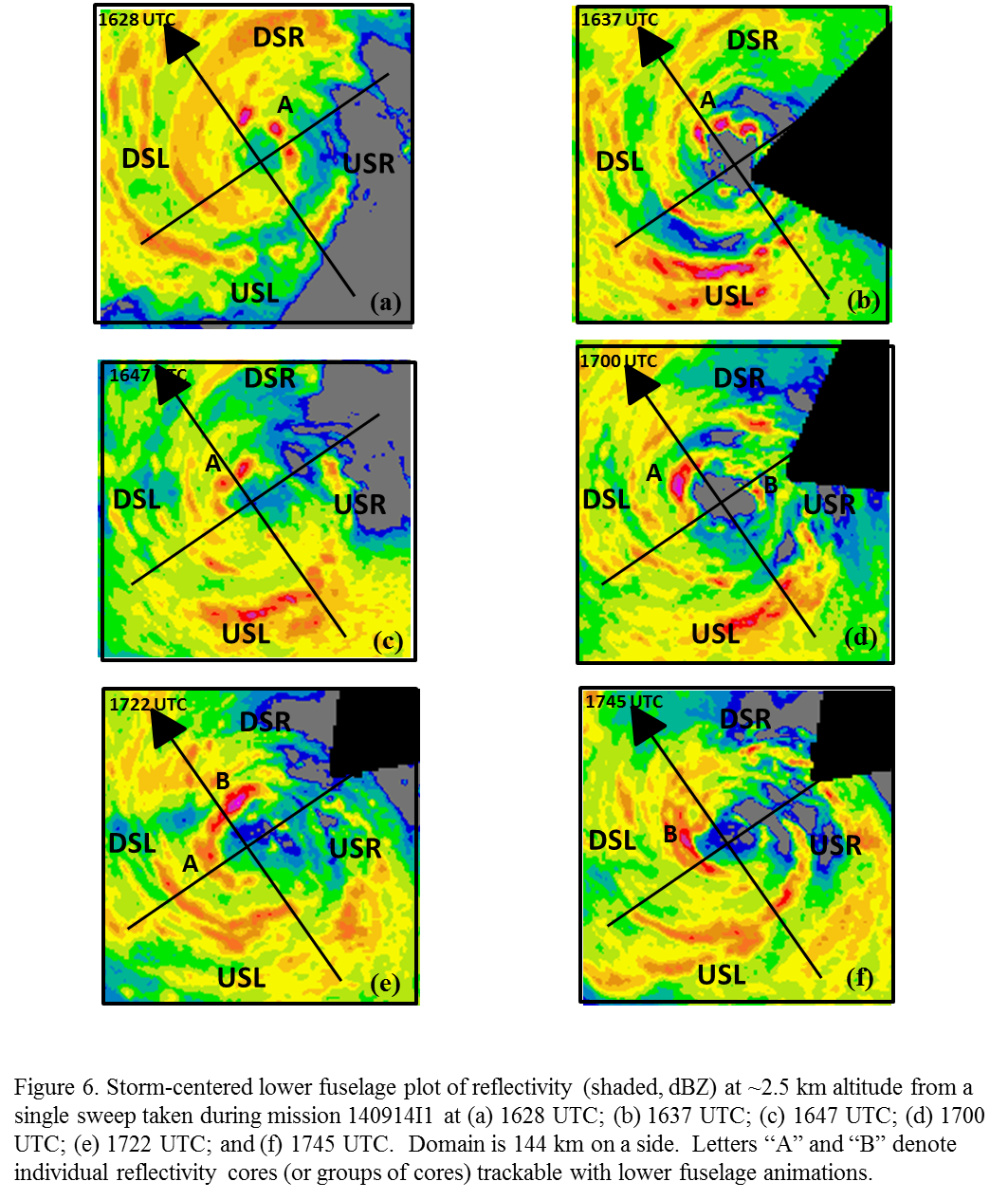
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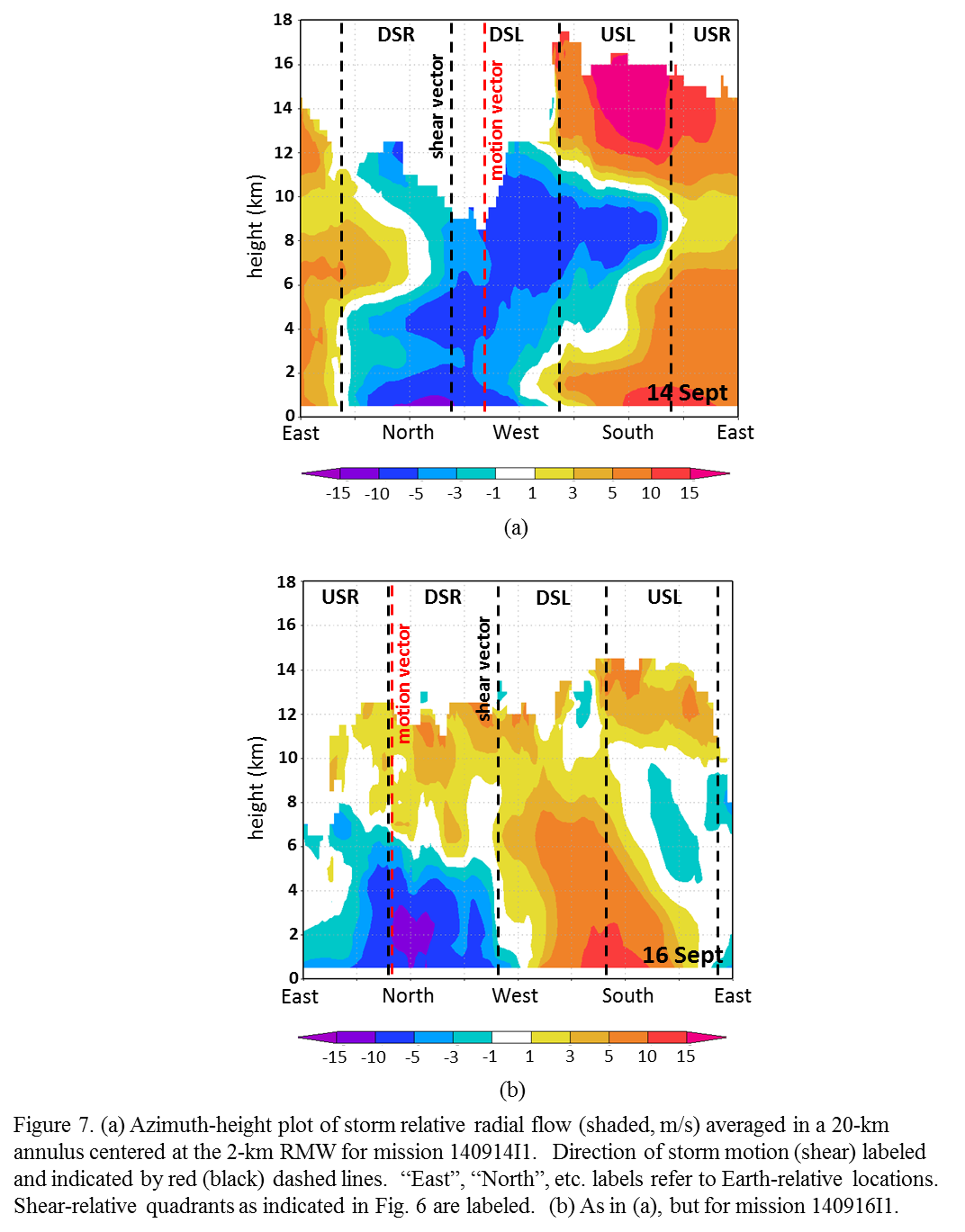
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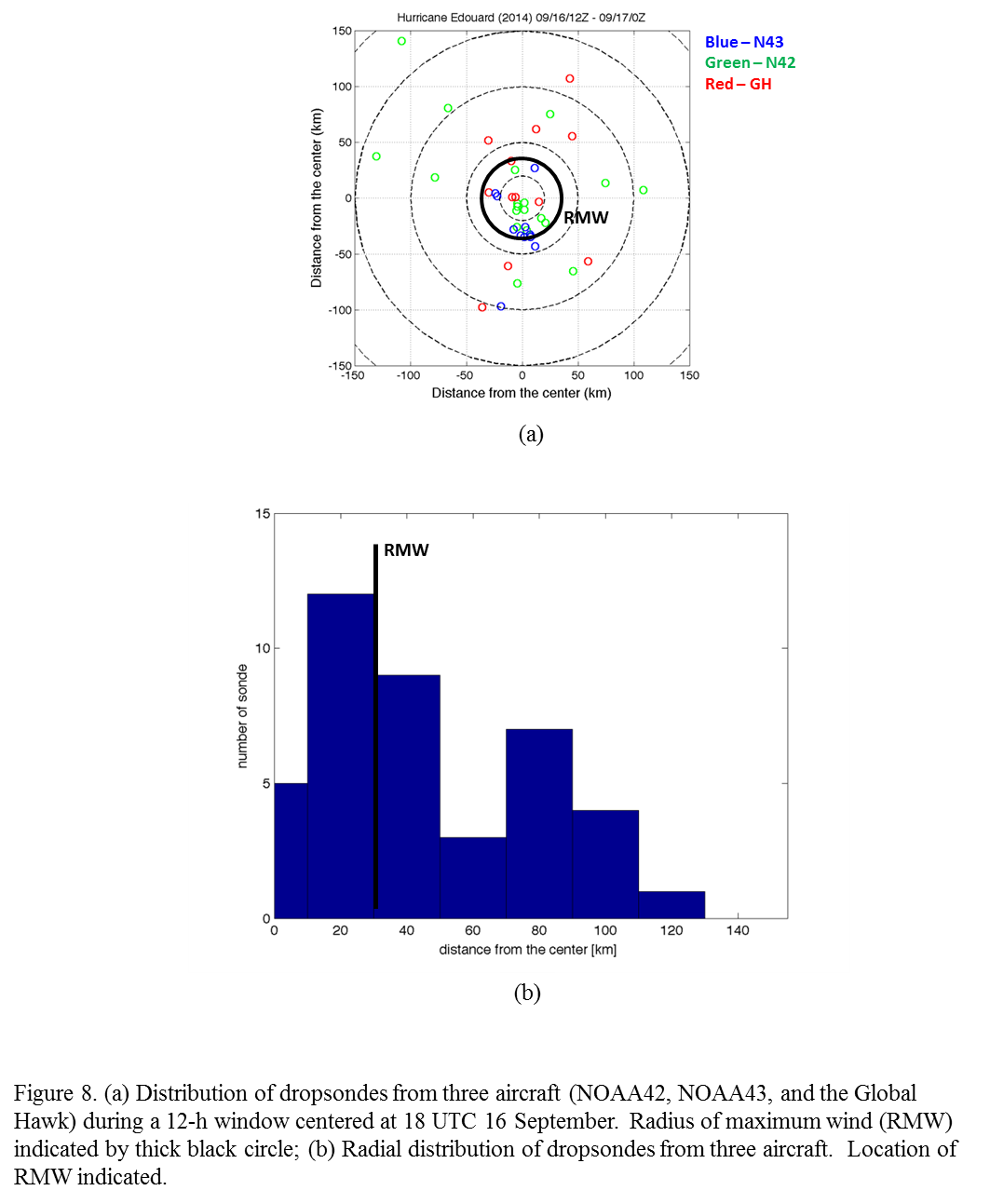
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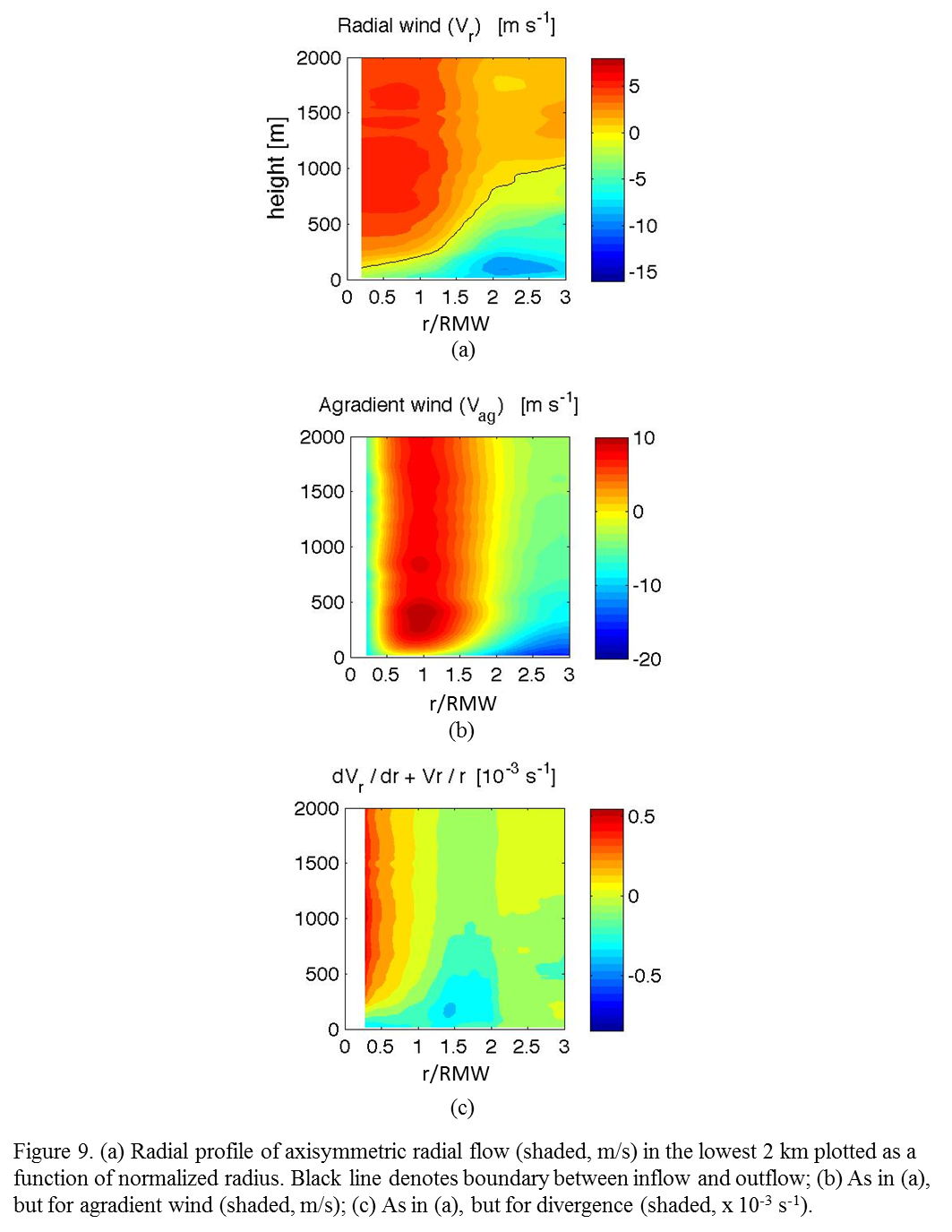
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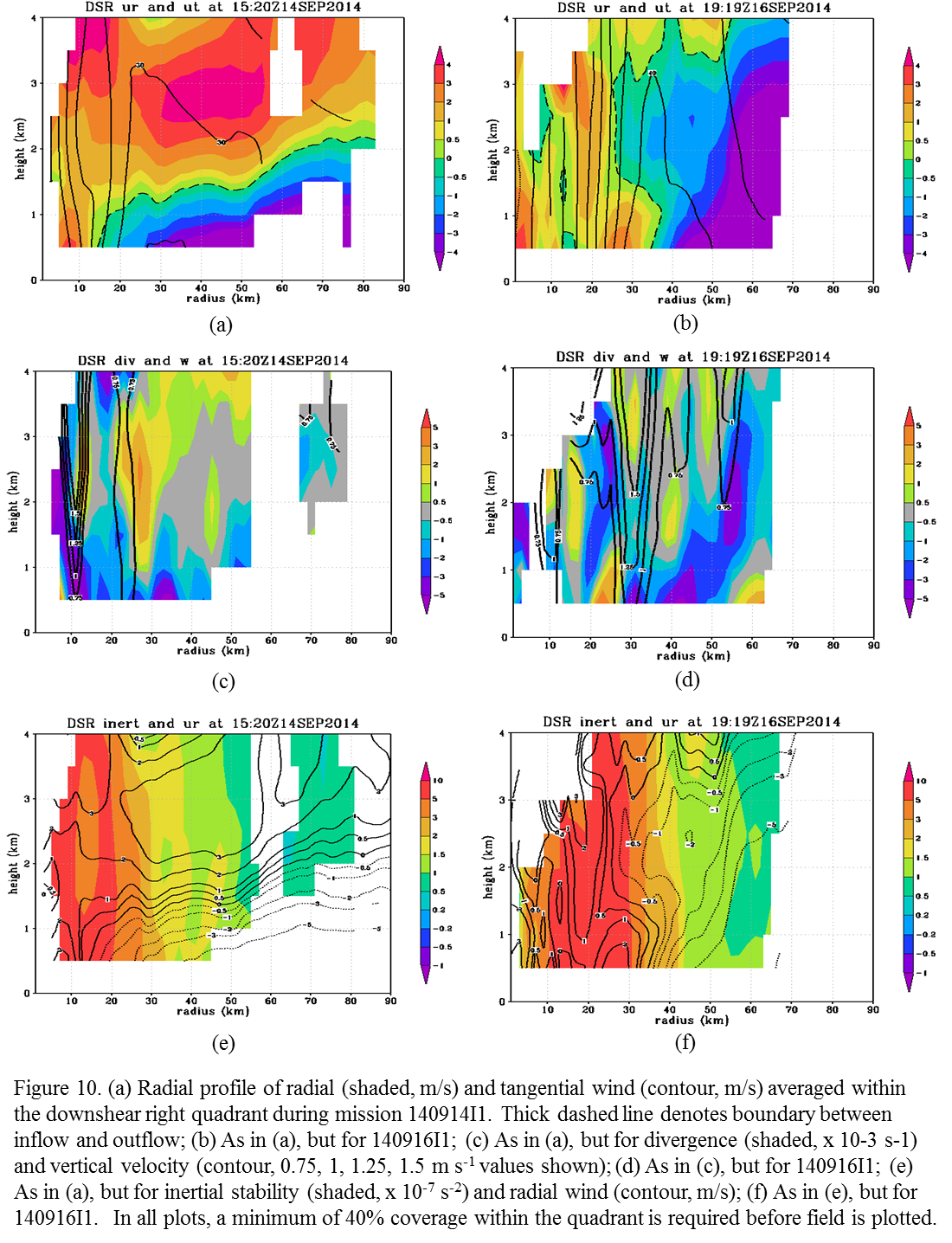
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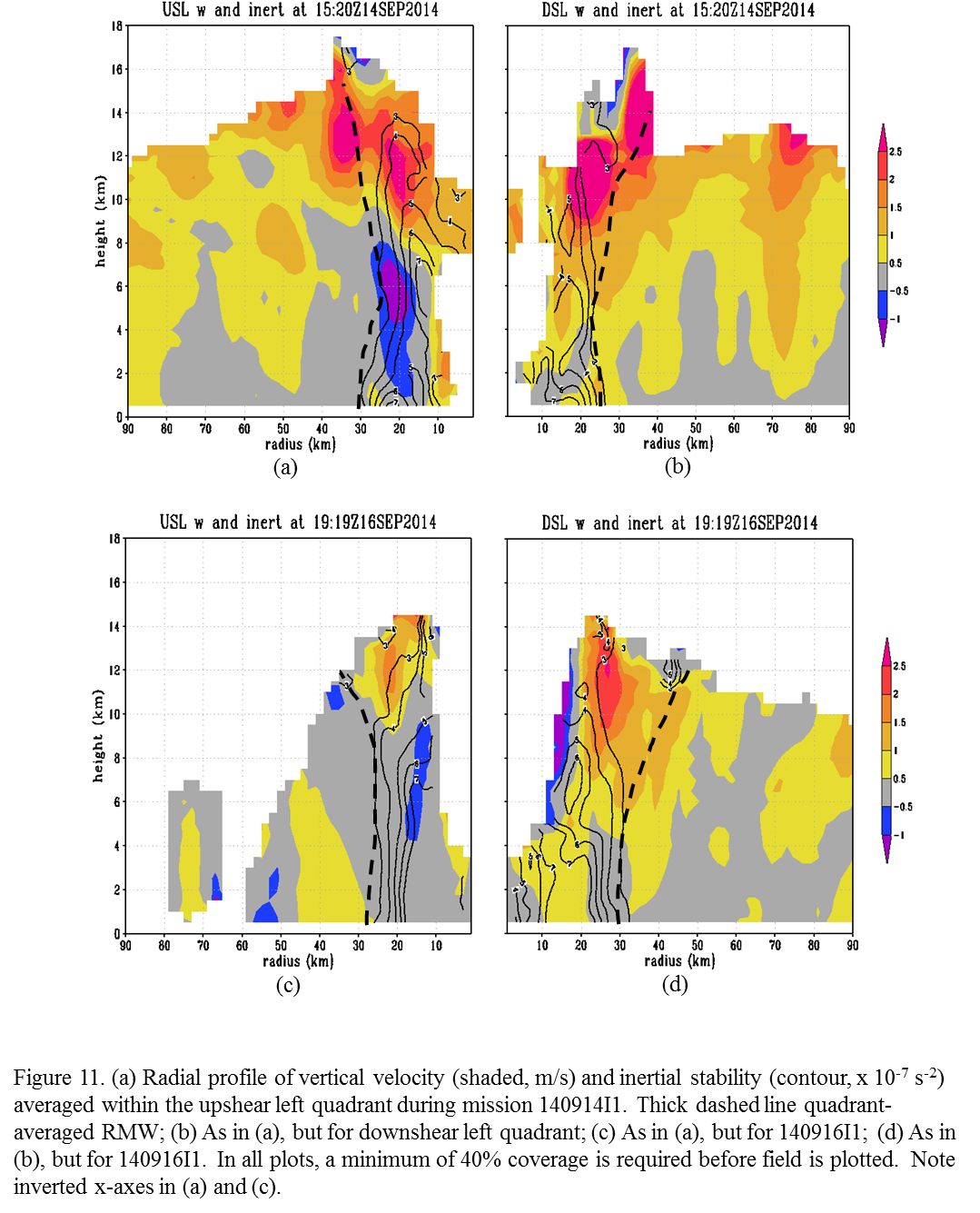
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