We would like to thank Brian Tang for carefully studying our manuscript and for his insightful comments. Our response to each of his detailed questions and suggestions, marked in blue, is provided below.

1 Choice of Experiments and Model Setup

Wind shear is switched on while the vortex is still intensifying up to its maximum intensity. Some suggestions for future experiments are to assess the sensitivity of the intensity evolution to the time when wind shear is applied. The two limiting cases are to switch on the wind shear once the TC reaches a quasi-steady state (i.e. its maximum intensity) and to switch on the wind shear right at the onset of the experiment. These follow up experiments would assess whether the weakening and subsequent strengthening observed in all of the current experiments is a robust feature.

We agree fully that the interaction of an intensifying TC with vertical wind shear examined here does not cover the full spectrum of possible TC–shear interactions. Brian Tang has nicely sketched the limiting cases of this spectrum.

The interaction of a TC with shear during genesis, however, can be expected to show very complex behaviour (e.g. Molinari et al. 2004) and can be considered a research area of its own right. We consider the comparison of the shear interaction of quasi-steady TCs and intensifying TCs to be an interesting extension of our current line of research. We plan to perform additional experiments with a weaker TC in future research. In the current experimental design, the TC reaches an 'equilibrium' intensity of ~100 m/s and may be expected to be extremely resilient. As a first step in the requested direction, however, we have performed one such numerical experiment at a reduced vortex intensity, i.e., we imposed the shear (15m/s) at 30 h corresponding to a time when the TC consolidated after the first episode of rapid intensification. All salient features found in the previous experiments have been confirmed in this experiment. A more detailed description of this experiment is contained in our response to Dr. John Molinari.

Although radiation likely doesn't play a strong role in the mechanism detailed in this study as the time scales of the phenomena at play are short (e.g. advection of a downdraft parcel to the radius of maximum winds), a simple radiation scheme in form of a slow relaxation to the initial state may prove useful in another regard in that it would help maintain the environmental shear in the near storm environment. The authors note that the shear evaluated in a 120km radius in the 20m/s shear case decreases considerably after 12 hours. One wonders if the tropical cyclone is in effect buffering the shear or that the shear over a larger region is being obliterated as the near field temperature gradient is diminished without any compensating forcing to restore the synoptic scale gradient.

We consider the axisymmetrization of the thermal and horizontal momentum fields associated with the imposed vertical shear in the near-core region to be physical processes. We agree (and state in section 6) that the forcing for the shear might be underestimated. Although the current experimental setup might benefit from a more realistic shear forcing, we have strong reservations with providing this forcing by a relaxation scheme towards a background temperature gradient in thermal wind balance with the zonal shear. While the balanced temperature gradient associated with the shear profile is very small (O(1K/1000km)), the temperature perturbations associated with the induced asymmetries are significantly larger. A relaxation towards the background state would artificially damp the dynamically-induced asymmetries. In this study we have purposely chosen to refrain from artificially modifying the storm asymmetries.

2 Wavenumber One Pattern and Tilt of TC

In figure 17, the authors show a prominent wavenumber one pattern in the vertical velocities extending out to approximately 120km. Associated with this is a wavenumber one pattern in the relative vorticity of the same sign (as implied by Ekman pumping). However if this wavenumber one standing vortex Rossby wave pattern is forced by the tilt of the TC and the tilt direction is generally toward the south (figure 14) at this time, why is the positive anomaly in the NW semi-circle of the storm? Heuristically, if one has a two layer model with separate PV anomalies in each layer in which the upper layer PV anomaly is displaced to the south, the projection of that anomaly on the lower layer should result in a positive anomaly that is likewise in the southern semicircle. If so, is convective coupling playing a dominant role making causality hard to attribute?

The tilt structure for a dry vortex is nicely depicted in Fig. 16b. of our manuscript. Clearly, the tilt is to the right, while high vorticity values at low levels are found left of the centre. The following provides a heuristic explanation for this behaviour: Consider an upright stationary vortex that is suddenly exposed to linear easterly shear. Barring the initial excitation and outward propagation of intertia-gravity waves that help enable the attainment of a quasi-balanced evolution (Shapiro and Montgomery 1993, Jones 1995), the vortex will first tilt to the West with height and also start moving to the West. At low levels (z<5 km), the storm relative flow is then from the West. At upper levels (z>7 km) the storm relative flow is from the East. By differential advection, the low-level vorticity will be found predominantly East of the centre and the upper-level vorticity to the West. Subsequently, the upper and lower vorticity anomalies interact by their 'induced circulations' at lower and upper levels, respectively. The horizontal flow associated with the anomalies leads to a mutual cyclonic displacement of the anomalies. Thus, the vortex tilt begins to rotate cyclonically. In the left-of-shear direction, the differential advection by the vertical shear and the mutual advection of the anomalies tend to cancel and a tilt equilibrium can be found. Jones (1995) studied this process using PV invertibility concepts and Reasor, Montgomery and Grasso (2004) extended these ideas by

formulating and solving a forced-damped oscillator model using a quasi-balanced vortex Rossby wave framework. It is indeed these dry vortex dynamics concepts on which we base our assertion that the vorticity anomaly can be considered to be the forcing for convection rather than being a consequence of a convective asymmetry that is 'forced' by other means. In order to make the vortex dynamics more accessible to the general community, we have added a bit more discussion on the vortex behaviour based on Jones (1995) in Sec. 5.1.

The lowest 4km of the inner core is tilted much more strongly than the remainder of the vortex above in figure 16. One might suspect this strong tilting of the inner core to be associated with a strong baroclinic generation of eddy kinetic energy (Wang, 2001). For future work, it would be interesting to investigate thoroughly what processes are responsible for the excitation of vortex Rossby waves in a sheared TC.

We agree that an assessment of the relative importance of dry, balanced tilt dynamics as compared to further contributions to the tilt evolution and vortex Rossby wave excitation could be an interesting venue for future research.

3 Theta_e diagnostics

The areal extent to which the boundary layer θe is perturbed seems to be quite an important factor in separating weakening from intensifying storms in the shear experiments. A broad, but overwhelming injection of low θe air into the boundary layer versus localized, but nevertheless, intense downdrafts of low θe air have entirely different effects. The former appears to primarily affect the low frequency variability reducing the storm intensity on time scales of greater than several hours, whereas the latter seems to play a role in the high frequency variability.
In the Hovmoller diagrams presented in figure 13, the contrast in areal coverage the downward flux of θe is quite distinguishable during the first 30 hours of each experiment with the 20m/s shear experiment having a more sustained area of downward fluxes between radii of 50-150km. It would be interesting to see a time series of the area integrated downdraft θe flux through the top of the boundary layer in a disk within 150km of the center to assess whether there is indeed a large difference between each experiment.

The requested area-integrated downward flux of low theta_e (within 150 km radius) is depicted in the figure below. This is the radial integration of the values depicted in Fig. 13. As suggested by the reviewer, this plot shows that the area-integrated downward flux increases with shear, at least if one applies a visual temporal smoothing. The area-integrated value of the downward flux of low theta_e, however, does not correlate well with the intensity evolution of the storm (see e.g. Fig. 3 or Fig. 13). In the original version of our manuscript we hypothesised in Sec. 5.2.2 (see also Fig. 11) that the spatial structure of the downdrafts plays an important role. The banded structure of the downdrafts can be expected to exert a persistent theta_e diminution of the inflowing air. A trajectory analysis is underway
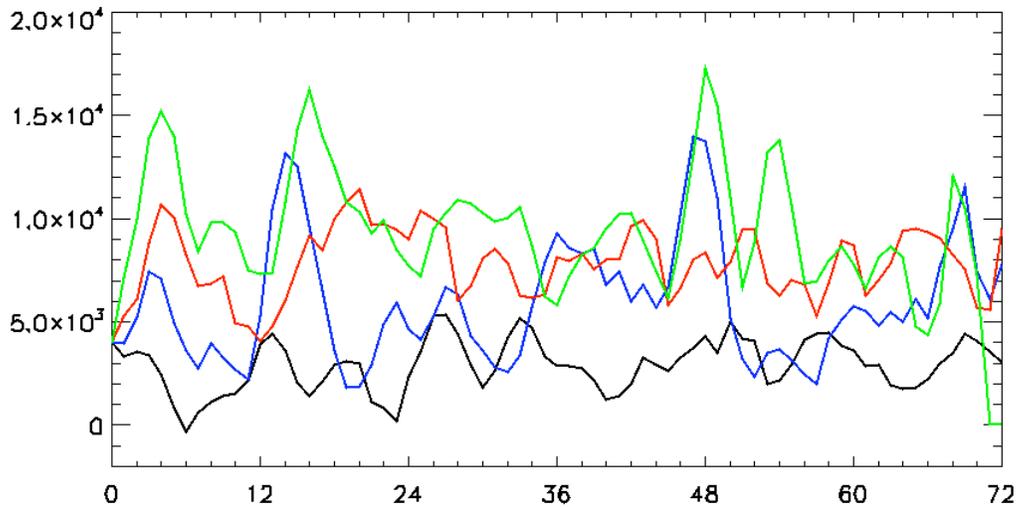
Time evolution of area integrated (out to 150 km radius) downward flux of theta_e (in Kms^{-1}km^2), black = no_shear, blue = 10mps, red = 15mps, and green = 20mps.. For visual clarity the 3 h moving average is shown.

to substantiate this hypothesis. It is therefore not surprising to us that the area-integrated downward flux does not show a clear correlation with the intensity evolution.

Also in figure 13, there are often transient upward spikes in intensity that very nearly coincide with the downdrafts. Although only strictly valid in axisymmetric, slantwise neutral TC theory, the radial gradient of θe is proportional to the tangential wind speed. The downdrafts could be locally increasing this gradient resulting in these spikes in intensity. An alternative dynamic view is that the downdrafts are locally increasing the inflow resulting in a larger Coriolis torque and a corresponding increase in the tangential wind speed.

Although we could not identify the times in Fig. 13 to which the reviewer refers, we are intrigued with the suggestion that divergent radial inflow associated with the downdrafts might lead to a <u>transient</u> spin up by increased radial advection of angular momentum. As the reviewer has mentioned, this argument is valid only in an azimuthal mean sense. After the proposed transient intensification, of course, downdrafts and the associated low-level outflow divergence would be expected to spin down the low-level vortex. Furthermore, the proposed transient spin-up mechanism might not be very pronounced. The downdrafts arguably favour to spread radially outward, rather than inward, because of the high Rossby elasticity associated with the strong inner core PV gradient that acts as a barrier for enhanced local inflow; or in the famous words of Michael McIntyre (1993): "the elastic eye-wall PV barrier".

In figure 18, there is a rather large vertical θe gradient in the lowest 1km coincident with the strong downdrafts on the south side of the storm. The downdrafts are obviously not only modifying the temperature and moisture of the boundary layer,
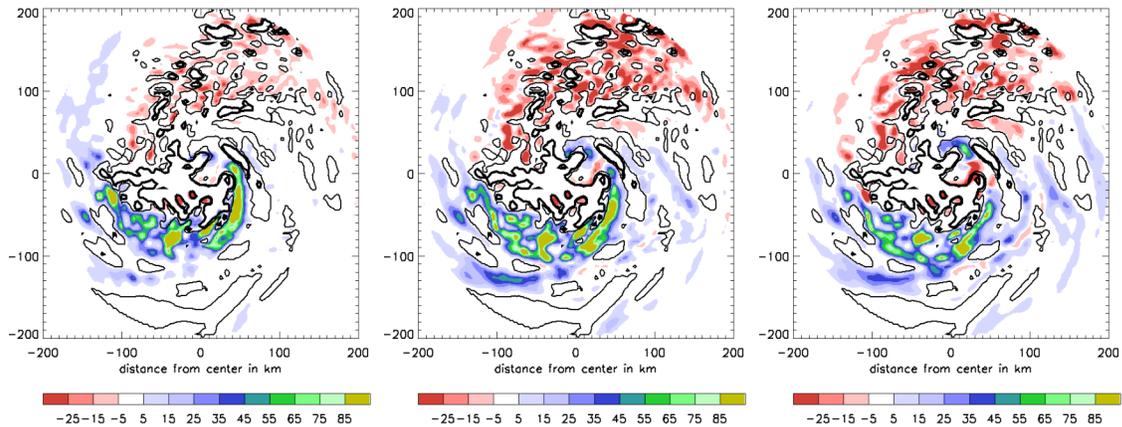
Same as Fig. 11 of our manuscript, but for a height of the inflow layer of 1km (left), 1.5km (middle), and 2 km (right).

but also of the boundary layer height and near surface stability. How are the authors defining the top of the boundary layer since it varies with space and time?

The boundary layer (BL) is here defined as the height of the axisymmetric inflow layer, approx. at 1.5 km (see section 4.2.1). The BL theta_e, depicted in e.g. Fig. 7, is averaged over the lowest 1km. The patterns of BL theta_e and the downward flux of theta_e are vertically coherent and not sensitive to the choice of the BL height between 1 km and 2 km (exemplified for the downward flux of theta_e in the figure above).

4 Relationship with Prior Studies and Observations

The effect of wind shear on TCs was also a subject of a paper by Frank and Ritchie (2001) in which they hypothesize that wind shear acts to weaken the storm in a "top-down" fashion. Asymmetries are hypothesized to dilute potential vorticity and entropy at upper levels first and then descend with time. The hypothesis presented in the current paper is clearly a different one, whereby the entropy sink is due to convective fluxes into the boundary layer. It would be interesting for the authors to compare and contrast their results with those of Frank and Ritchie.

Frank and Ritchie's hypothesis of the upper-level warm core erosion is based on an Eulerian diagnostic, namely the radial eddy fluxes of theta_e. Our reservations with an Eulerian viewpoint is indicated in the introduction. We believe that air mass exchange/ mixing in TCs needs to be examined by Lagrangian means. A respective trajectory analysis is in preparation. We do not yet have results to quantitatively assess the importance of possible upper-level warm core erosion.

In our experiment the low-level wind maximum decreases significantly before the wind profiles above the boundary layer change considerably (see end of Section 4.1). Apparently, the vortex weakens at low-level first, as opposed to a 'top-down decay' process. This strongly indicates that boundary layer processes – the

significant depression of BL theta_e – governs the intensity evolution and that mid- to upper-level processes are of secondary importance. We have correspondingly modified the last paragraph of section 4.1 and have added a statement in the 2nd paragraph of the conclusions to provide a closer comparison with previous work.

The authors point to ancillary evidence of the same downdraft mechanism occurring in two recent hurricanes: Omar in 2008 and Erin in 2001. Was there evidence in either reconnaissance, best track record, or other objective/subjective estimates of intensity of subsequent weakening after these arc clouds or bands of strong divergence were noted?

According to the NHC webpage, Omar started to rapidly decay 12 h after the time considered in Fig. 21: from 115 to 70 kt in 12 h. Until that time, NHC determined intensification from 80 to 115 kt. A time lag of O(6 h) between the first occurrence of strong downdrafts and significant weakening is found also in our experiments, in particular in the 10mps case. Erin weakened from 100-105kt to 80kt in the subsequent 12h-18h after the time shown in Fig. 22 (again, according to the NHC webpage).

Care must be taken, however, when drawing conclusions from these observations because it is not known how the storm would have evolved without the occurrence of the indicated downdraft patterns. Moreover, there are some complications when comparing the intensity evolution in our experiments with the intensity evolution of real hurricanes as determined by NHC. In situ measurements of 1min sustained surface winds, the relevant quantity for intensity as defined by NHC, are rarely available. Storm intensity is therefore usually derived by NHC from other observations, inter alia storm structure, and consequently the forecasters' estimates of storm intensity sometimes lags behind the intensity evolution of the real storm. Finally, please recall that intensity by NHC is determined as the maximum wind speed, whereas our study uses the azimuthal average of the tangential winds as the intensity metric.

References:

McIntyre, M. E.: Isentropic distributions of potential vorticity and their relevance to tropical cyclone dynamics, Tropical Cyclone Disasters: Proceedings of ISCU/WMO International Symposium, edited by: Lighthill, J. Zhemin, Z., Holland, G. J., and Emanuel, K., Peking University Press, 143–156, 1993.

Molinari, J., D. Vollaro, and K.L. Corbosiero, 2004: Tropical cyclone formation in a sheared environment: A case study. *Journal of the Atmospheric Sciences*, **61**, 2493-2509.

Please refer to reference list in discussion paper for further references.