Diagnosis of the Source of GFS Medium-Range Track Errors in Hurricane Sandy (2012)

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ABSTRACT

Medium-range forecasts of Hurricane Sandy’s track were characterized by widely diverging solutions, with some suggesting that Sandy would make landfall over the mid-Atlantic region of the United States, while others forecast the storm to move due east to the north of Bermuda. Here, dynamical processes responsible for the eastward-tracking forecasts are diagnosed using an 80-member ensemble of experimental Global Forecast System (GFS) forecasts initialized five days prior to landfall. Comparing the ensemble members with tracks to the east against those with tracks to the west indicates that the eastern members were characterized by a lower-amplitude upper-tropospheric anticyclone on the poleward side of Sandy during the first 24 h of the forecast, which in turn was associated with a westerly perturbation steering wind. The amplification of this ridge in each set of members was modulated by differences in the advection of potential vorticity (PV) by the irrotational wind associated with Sandy’s secondary circulation and isentropic lift along a warm front that formed on the poleward side of Sandy. The amplitude of the irrotational wind in this region was proportional to the 0-h water vapor mixing ratio, and to a lesser extent the 0-h upper-tropospheric horizontal divergence. These two quantities modulated the vertical profile of grid-scale condensation within the model and subsequent upper-tropospheric divergence. The results from this study suggest that additional observations within regions of large-scale precipitation outside the tropical cyclone (TC) core could benefit TC track forecasts, particularly when the TC is located near an upper-tropospheric PV gradient.

1. Introduction

Hurricane Sandy (2012) was one of the costliest natural disasters in U.S. history due to its expansive wind field, large storm surge, and unusual track into the population-dense Northeast corridor (Blake et al. 2013). Although some operational numerical weather prediction systems suggested that Sandy could make landfall over the U.S. East Coast up to eight days in advance, there was considerable variability in Sandy’s position between models and even within ensemble members of the same model, with some forecasts correctly predicting landfall over New Jersey, while others predicted the storm would take an easterly track into the open Atlantic Ocean near Bermuda. Four days prior to landfall, most operational deterministic and ensemble prediction systems indicated Sandy would make landfall over the United States; however, the easterly track solution was still present in some forecasts. The large variability in Sandy’s track forecast suggests that this forecast could have been particularly sensitive to the combination of initial condition and model

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error; therefore, it is of interest to understand what processes contributed to the track variability in this case. The processes that govern the motion of tropical cyclones (TCs) and the sources of motion error have received considerable attention. In general, TC motion is primarily driven by the environmental flow (e.g., Chan and Gray 1982; Holland 1983; Chan et al. 2002, and others), although there is significant case-to-case variability in how to define that steering flow (e.g., George and Gray 1976; Hong and Neumann 1986; Velden and Leslie 1991; Aberson and DeMaria 1994, and others). In addition, others have shown that errors in the placement or structure of synoptic-scale features give rise to TC position errors (e.g., Carr and Elsberry 2000; Wu et al. 2004). Sensitivity analysis provides a method of determining where in the initial conditions small changes can have a substantial impact on a forecast metric, such as TC position. Although there can be disagreement between the various methods regarding what aspect of the initial conditions the track forecast is most sensitive to (Majumdar et al. 2006; Wu et al. 2009), TC position forecasts are often sensitive to remote weaknesses in the subtropical ridge, transient midlatitude troughs, the subtropical jet, as well as the steering flow near the TC itself (e.g., Majumdar et al. 2006; Peng and Reynolds 2006; Wu et al. 2007a; Chen et al. 2009; Wu et al. 2009). Moreover, assimilating additional observations that better define these features often leads to improved track forecasts (e.g., Aberson 2003; Wu et al. 2007b; Aberson 2010; Jung et al. 2010; Chou et al. 2011; Weissmann et al. 2011; Jung et al. 2012).

Although many of these previous studies have highlighted the importance of initially remote features, there has been less work on how the TC can modify its environment and, hence, feed back onto its motion (e.g., Wu and Emanuel 1995a,b; Henderson et al. 1999; Anwender et al. 2008; Harr et al. 2008). In the case of Sandy, this could have been particularly important given the proximity of the TC to midlatitude features, including troughs and fronts, throughout its life. As a consequence, it is worthwhile to consider Sandy within the paradigm of a TC undergoing extratropical transition (ET; Jones et al. 2003), whereby a warm-core cyclone acquires the properties of a cold-core system. The process of ET is often accompanied by the development of a downstream anticyclone, which can influence the surrounding environment and steering flow. The modification of the environment by the cyclone can be achieved through either adiabatic action at a distance by the TC winds (e.g., Ferreira and Schubert 1999; Riemer et al. 2008) or through the diabatically driven divergent outflow from either the tropical cyclone itself or the baroclinic zone that often forms on the poleward side (e.g., Bosart and Dean 1991; Harr and Elsberry 2000; Riemer et al. 2008; Riemer and Jones 2010; Torn 2010; Grams et al. 2013). This downstream ridge is due to the diabatic destruction of potential vorticity (PV; e.g., Riemer and Jones 2010), or the advection of low PV air by the irrotational wind (e.g., Archambault et al. 2013). Previous work has suggested that the amplitude of the downstream ridge is a function of the magnitude of the frontogenesis in the poleward baroclinic zone, or the poleward horizontal flux of moisture into the front; both of these quantities can modulate the amount of latent heat release and, hence, the amount of divergent outflow and PV destruction (e.g., Torn 2010). As a consequence, it is possible that errors associated with any of these processes could feed back onto the larger-scale flow and impact the track forecast.

This study evaluates the sources of the eastward turn in Hurricane Sandy’s track forecast initialized five days prior to landfall (0000 UTC 25 October 2012). Rather than diagnosing the source of the track variability by comparing forecasts from two or more models, which could be characterized by different initial conditions and model formulations, this study employs a large ensemble of forecasts from an experimental version of the National Centers for Environmental Prediction (NCEP) Global Forecast System (GFS), which was run as part of the real-time Hurricane Forecast Improvement Project (HFIP) Demonstration Project (Gall et al. 2013). The processes responsible for the erroneous eastward tracks are diagnosed by comparing composites of ensemble members that predict Sandy turning east out into the open Atlantic Ocean against the composite of members that contained the observed westward turn (e.g., Reinecke and Durran 2009), as well as the ensemble-based sensitivity method (e.g., Ancell and Hakim 2007; Torn and Hakim 2008). Although previous studies have shown that the eastern track forecasts were related to the zonal component of the steering wind (e.g., Munsell and Zhang 2014), amplitude of the ridge on the poleward side of Sandy (e.g., Magnusson et al. 2014), or the choice of cumulus parameterization (e.g., Bassill 2014), this paper provides a more detailed description of the source of the differences.

The remainder of the paper proceeds as follows. Section 2 describes the dataset and methods used in this study. A short overview of Sandy is provided in section 3, while the dynamical processes responsible for the track differences are explored in section 4. A summary and conclusions are given in section 5.

2. Model description

As part of the National Oceanic and Atmospheric Administration (NOAA) HFIP real-time demonstration project, 80-member ensemble forecasts out to five days
were produced using an experimental higher-resolution version of the NCEP GFS model at a spectral triangular truncation at wavenumber 382 with 64 vertical levels (T382L64, 40-km horizontal resolution), which was run in near–real time from 1 June to 1 November 2012. The ensemble used the operational model for 2012, but at higher resolution (the operational version was T254L42, ~55 km at that time). (A history of GFS model changes can be found online at http://www.nco.ncep.noaa.gov/pmb/changes.)

In addition, forecasts were initialized with a higher-resolution version of the three-dimensional ensemble-variational hybrid data assimilation system that became operational at NCEP in May 2012 (see http://www.nws.noaa.gov/os/notification/tin12-22gfs_hybridaab.htm for details on the May 2012 GFS upgrade). An 80-member ensemble Kalman filter assimilation system provided a background ensemble that was partly used to estimate the background error covariance in the variational assimilation system. The total background error covariance was made up of a blend of 25% three-dimensional variational data assimilation (3D-Var) static background error covariance and 75% ensemble-based covariance. The variational system updated a high-resolution (T878, 20-km horizontal resolution) control forecast, and the analysis ensemble was produced by updating the ensemble perturbations and centering them around the control analysis. The configuration of the experimental analysis system was identical to that run in NCEP operations, except that the experimental system was run at higher resolution (T878 for the control forecast and T382 for the ensemble, as opposed to T574/T254), and tropical cyclone relocation (Liu et al. 2000) was not employed. The interested reader is directed to Hamill et al. (2011) and Wang et al. (2013) for additional details on the model and data assimilation system. The remainder of this study focuses on forecasts initialized at 0000 UTC 25 October. Ensemble forecasts initialized 12 h prior to and after this time were characterized by qualitatively similar results (not shown).

3. Overview of Sandy

Sandy originated from an African easterly wave that moved off Africa on 11 October and subsequently weakened as it moved across the Atlantic basin (Blake et al. 2013). Upon reaching the central Caribbean on 20 October, the wave entered an environment with deep moisture and lower vertical wind shear. Sandy was declared a tropical depression at 1800 UTC 22 October, subsequently underwent rapid intensification, and reached its maximum intensity at 0600 UTC 25 October, just as the cyclone made landfall over Cuba. The focus of this study is on forecasts initialized at 0000 UTC 25 October, which was chosen based on the large variability in position forecasts and because Sandy made landfall within five days.\(^1\) Subsequent initialization times were characterized by less position variability at landfall time.

Although Sandy was a category-2 TC at the forecast initialization time, the TC was associated with asymmetries in the nearby lower-tropospheric thermal and outflow pattern. Figure 1a shows that Sandy had a stronger 850-hPa geopotential height gradient, and in turn stronger geostrophic winds, on the eastern side of the TC compared to the western side. On the northeast side of Sandy, the inferred geostrophic winds turned sharply from southerly (parallel to the temperature gradient vector) to easterly north of Cuba (perpendicular to the temperature gradient vector), implying a confluent wind pattern, which led to an area of 850-hPa frontogenesis on the order of 0.6 K (100 km)\(^{-1}\) at 24°N, 76°W. Taking a cross section along 76°W (Fig. 2) indicates that this region of frontogenesis sloped poleward with height due to the interaction of Sandy with an upper-tropospheric trough to its west, such that the maximum frontogenesis is at 400 hPa. Moreover, the region between 21° and 25°N and between 850 and 200 hPa is characterized by isentropic lift, as evidenced by the positive meridional wind within a region where the isentropic surface slopes upward toward the north. Much of this area is near saturation; therefore, there was a broad region of precipitation to the north and east of Sandy with precipitation rates greater than 4 mm (3 h)\(^{-1}\), with the largest precipitation associated with two maxima: one just north of the 0000 UTC position of Sandy and another 300 km to the north of the TC center (Fig. 3). This asymmetric distribution of temperature and formation of fronts is more typical of a TC at the onset of extratropical transition, rather than a symmetric TC (e.g., Klein et al. 2000; Jones et al. 2003); however, Sandy had yet to begin the process of transition at this time.

At the same time, the PV distribution on the 340-K isentropic surface\(^2\) (Fig. 4a) was characterized by higher potential vorticity associated with a longwave trough over the western United States (not shown), a ridge downstream over the eastern half of the United States, and a second deep trough over the western Atlantic. In addition, there was a narrow corridor of higher PV extending from the western Atlantic trough that terminated over the eastern Gulf of Mexico, the result of

\(^1\) At this time, NHC issued 5-day track forecasts.

\(^2\) This isentropic level corresponds to roughly 250 hPa to the north of Sandy and intersected the tropopause along the subtropical waveguide.
earlier anticyclonic wave breaking over the western Atlantic Ocean. This region of high PV was associated with a tropopause-based trough; the inferred southerly winds in between this trough and the ridge to the east are consistent with the poleward movement of Sandy. Unlike a classical TC that has maximum divergence and irrotational wind vectors emanating from near the TC center, the greatest divergence at this time was shifted to the north of Sandy’s center and slightly north of the aforementioned region of heavy precipitation associated with the warm front (Fig. 3), while the largest irrotational wind vectors were to the north of this front and 600 km to the north of Sandy. Essentially, this pattern of divergence and irrotational wind vectors represented the superposition of the divergence associated with the convection associated near Sandy’s core, and upward motion associated with the isentropic lift to the north of Sandy and the developing warm front (cf. Fig. 2). To the west and north of Sandy, the irrotational wind vectors were parallel to the PV gradient, which led to PV advection up to \(24\) (potential vorticity units) \(\text{PVU} (1 \text{ PVU} = 10^{-6} \text{ K kg}^{-1} \text{ m}^2 \text{ s}^{-1})\) day\(^{-1}\) along the eastern edge of the trough along the 1.5-PVU contour from 85\(^\circ\) to 65\(^\circ\)W (Fig. 4a). Previous studies suggest that advection of PV by the irrotational wind can be an effective method of amplifying the downstream ridge (e.g., Riemer and Jones 2010; Archambault et al. 2013).

Over the next 24 h, Sandy moved poleward across Cuba and was over the Bahamas by 0000 UTC 26 October (Fig. 1b). At this time, both the horizontal temperature and geopotential height gradients were increasing on the northeast side of the Sandy, which led to a more
extensive region of frontogenesis with a larger maximum value \[3 K \text{ (100 km)}^{-1} \text{ (3 h)}^{-1}\]. As a consequence, one might expect the greater frontogenesis to be associated with greater forcing for vertical motion and upper-tropospheric divergence. Indeed, the 340-K irrotational wind vectors exceeded 10 m s\(^{-1}\) over much of the region north of Sandy (Fig. 4b). Along the northern and western edge of the ridge, the greater magnitude irrotational winds were associated with up to \(-8 \text{ PVU day}^{-1}\) PV advection, which was more than twice the value from 24 h prior. At the southern end of the aforementioned eastern Gulf of Mexico trough (in between Cuba and the Yucatan Peninsula), the PV began to cyclonically wrap up, which likely helped turn Sandy to the west toward Florida during this period. Farther upstream over North America, the longwave trough, denoted by the expansive region of high PV over the northern Rockies, shifted farther east, with the trough axis centered near 108°W (not shown).

By 0000 UTC 27 October, Sandy continued to move slightly west of north parallel to the coast of Florida and into an increasingly baroclinic environment (Fig. 1c). At this point, the 900–600-hPa thickness asymmetry, which was used by Hart (2003) to determine an objective definition of the onset of ET, was 19 m, which is indicative of a TC that has begun the process of ET. The region of maximum frontogenesis was then located on the northwestern side of the TC; however, frontogenesis in excess of \(0.5 K \text{ (100 km)}^{-1} \text{ (3 h)}^{-1}\) extended 800 km along the warm front located at roughly along 30°N. On the 340-K surface, the ridge on the poleward side of Sandy continued to amplify, where the greatest negative PV advection by the irrotational wind was to the west of the TC over Georgia, with comparatively smaller PV advection along the northern edge of the ridge (Fig. 4c). The persistent negative PV advection to the northwest of Sandy forced the aforementioned PV strip to move westward with time and led to the cyclonic wrapping of the southern edge of that trough around the southern side of the TC, which in turn allowed Sandy to move farther west. The upstream midlatitude trough axis moved to 98°W by this time and subsequently combined with the PV strip after this time. Beyond this time, the progression of this trough was slowed by the divergent outflow from Sandy (e.g., Pantillon et al. 2014).

From 0000 UTC 27 October to 0000 UTC 29 October (21 h prior to landfall), Sandy moved parallel to the U.S. coastline, turning west thereafter. Over this period, the cyclonic circulation associated with Sandy increased considerably through mainly baroclinic processes (e.g., Galarneau et al. 2013), with well-defined warm and cold fronts on the northeast and southwestern sides, respectively (Fig. 1d). The rapid westward motion was a consequence of the interaction of Sandy with the aforementioned large midlatitude trough moving southeast into the southeastern United States and cyclonically wrapping around Sandy’s circulation (Fig. 4d). This westward motion and wrap-up of the trough was characteristic
of an LC2 or cyclonic wave breaking, while the PV distribution over the Canadian Maritime provinces was more characteristic of an LC1 or anticyclonic wave breaking (e.g., Thornicroft et al. 1993). Moreover, the irrotational wind vectors were far displaced from the center of the TC, which further prevented the trough from progressing to the east (e.g., Pantillon et al. 2014).

4. Forecast sensitivity

Similar to other operational modeling systems, including the European Centre for Medium-Range Weather Forecasts (ECMWF) ensemble prediction system (Magnusson et al. 2014), experimental GFS ensemble forecasts initialized at 0000 UTC 25 October exhibited a significant amount of uncertainty regarding the evolution of Sandy (Fig. 5). Over the first 72 h of the forecast, most of the position variability was in the across-track direction, yet all of the members captured both the initial westward, then eastward turn of Sandy, even though the best-track position was on the eastern edge of the ensemble over the first 24 h. By 96 h, the ensemble was spread out from northwest to southeast, with some members showing Sandy off the North Carolina coast, while others had Sandy within 500 km of Bermuda. At 120 h (3 h after landfall), many of the ensemble members had Sandy making landfall somewhere over the eastern United States though with slight timing differences, while other members showed Sandy moving northeast toward the Canadian Maritime or due east along 32°N. In general, the track forecasts from the operational GFS ensemble system were qualitatively similar, with some members showing a westward turn, while others had the eastern solution (not shown). Based on Fig. 4 and Fig. 5a, the
lack of a westward turn in some members was likely due to Sandy being too far away from the midlatitude trough once the trough reached the U.S. East Coast.

To understand what processes led to the divergence in forecast tracks, ensemble members were separated into subgroups depending on their track evolution. The first subgroup, denoted “ExtremeEast,” was the eight members with the largest southeastward displacement along the leading eigenvector of position variability (Hamill et al. 2011) at 96 and 120 h; these members are shown in Fig. 5b. For the remainder of this paper, this subset of members will be compared to the “AllWest” members, which is defined as all members to the west of the ensemble mean position between 96 and 120 h (Fig. 5c).

The dynamical processes responsible for the difference in Sandy’s track evolution were diagnosed by comparing the mean of the ExtremeEast members against the AllWest subset at various lead times. Rather than computing the absolute difference, this study evaluates the normalized difference (i.e., standardized anomaly) between two subsets via

$$\Delta x_i = \frac{x_{i}^{\text{ExtremeEast}} - x_{i}^{\text{AllWest}}}{\sigma_{x_i}}$$

where $x_{i}^{\text{ExtremeEast}}$ ($x_{i}^{\text{AllWest}}$) denotes the mean of the $i$th state variable for the ExtremeEast (AllWest) ensemble members and $\sigma_{x_i}$ is the ensemble standard deviation of $x_i$ computed from all members. Dividing by the ensemble spread standardizes the differences and permits comparison among different vertical levels, fields, and times. Moreover, this choice also aids in the interpretation on a single vertical level because it downplays the differences along large gradients, which can have large absolute differences due to displacement errors. The statistical significance of the composite differences was assessed using a bootstrap resampling without replacement. In particular, two subsets of ensemble members, equal in size to the ExtremeEast and AllWest members, were randomly selected from the 80-member ensemble, and the difference between the ensemble mean of the two subsets was calculated. This process was repeated 10000 times to obtain the 95% confidence bounds on the composite difference. It is worth noting that the same set of resampled ensemble members was used for each horizontal location and field, thus providing some measure of the statistical significance of individual locations, while also taking into account horizontal correlations.

Before determining what processes contributed to the eastern tracks during Sandy, it is necessary to determine what vertical level(s) constituted Sandy’s steering wind.
Here, the steering wind at any given time was determined using the Galarneau and Davis (2013) algorithm, which involves matching the TC motion with a mean vector wind averaged over a horizontal area and vertical layer. In essence, this method involves computing the vorticity and divergence associated with the TC vortex within a given distance of the TC center, then applying the Poisson equation to solve for the streamfunction and velocity potential, and hence the nondivergent and irrotational winds, respectively, associated with the TC vortex at each pressure level. The environmental wind vector at any location is then defined as the difference between the total wind vector and the TC vortex nondivergent and irrotational vector. An area-averaged environmental wind was then computed by averaging the environmental wind within a certain radius and vertical depths starting at 850 hPa. This process was repeated for different vertical depths and radii to determine the steering level, which was defined as the area-average wind that most closely matches the TC motion between 612 h.

In theory, the steering wind could be a function of lead time and ensemble member, which could complicate the ability to evaluate the differences in steering wind between the ExtremeEast and AllWest members. Based on Fig. 5b, the ExtremeEast members were at the eastern end of the ensemble members within the first 48 h; therefore, it is important to define a steering flow for that time period. The optimal steering flow for the first 48 h was determined by computing the mean-absolute difference between Sandy’s forecast motion and the area-average environmental wind, computed with various radii and vertical depths, then averaging over all ensemble members and repeated every 6 h between 12 and 48 h. Figure 6a shows that the lowest mean-absolute difference was obtained for a TC removal radius of 4° and the 850–300-hPa layer wind. As a consequence, the steering wind was defined as the 850–300-hPa layer-average wind, which is similar to what was found in Galarneau and Davis (2013) for Hurricane Earl. The mean vector wind difference for this layer was 1.0 m s⁻¹, while the time-averaged speed of Sandy during the same time period was 5.8 m s⁻¹, suggesting that this layer-averaged wind does a reasonable job describing the motion of Sandy. It is worth noting that the optimal vertical depth and radius was relatively insensitive to computing the difference over shorter periods of time (i.e., 12–24 h, 24–36 h, etc.; not shown). Moreover, the optimal depth and radius for the ExtremeEast members (Fig. 6b) and the AllWest members (Fig. 6c) were identical; therefore, these two subsets of members did not experience a deeper or shallower steering flow relative to each other. This might have been the case if the ExtremeEast members were weaker or stronger than other members of the ensemble; however, the mean minimum SLP and maximum wind speed of the ExtremeEast members was not statistically different from the AllWest members until 84 h (not shown).

For the remainder of this section, the meteorological rationale for the eastern tracks is evaluated by computing the difference between the ExtremeEast and

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3 The 0- and 6-h values were not used so that the calculation would not be contaminated by initialization issues.
AllWest subsets of ensemble members. Before exploring the differences in individual fields, the lead time when the across-track distance in the ExtremeEast subset became statistically distinct from the AllWest subset was determined. Figure 7a shows the mean across-track distance of these two subsets as a function of forecast hour. The across-track distance in the ExtremeEast members increased from 0 to 75 km during the first 24 h of the forecast, then increased by another 100 km from 24 to 48 h. Although the across-track difference decreased from 48 to 60 h as Sandy turned from northwest to northeast, the ExtremeEast position increased quickly thereafter, suggesting there might be multiple reasons for the position differences. The difference in across-track distance between the ExtremeEast and AllWest subsets exceeded the bounds of the 95% confidence interval starting at 12 h; thus, it appears the two subsets became distinct early in the forecast.

The difference in across-track distance between the ExtremeEast and AllWest members appear to originate from differences in the 0-h TC steering wind. Figure 7b shows that the zonal steering wind for the ExtremeEast members was 0.3 m s$^{-1}$ larger than the mean of the AllWest members during the first 12 h; however, this difference increased to 1.2 m s$^{-1}$ by 36 h (statistically significant from 6–36 h). The positive perturbation steering wind in the ExtremeEast subset would be expected to direct Sandy to the east with time, consistent with Munsell and Zhang (2014). During Sandy’s eastward turn, the steering wind difference became minimal again, then increased to over 5 m s$^{-1}$ beyond 60 h, suggesting that the ExtremeEast members experienced two distinct periods of perturbation westerly steering flow. It is worth noting that the variations in steering wind became less tied to the initial values over time. While the correlation between the ensemble estimates of the 0-h zonal steering wind and the ensemble estimates of the zonal steering wind prior to 12 h was greater than 0.75, the correlation remained below 0.2 for all subsequent lead times (not shown). This result suggests that the steering wind differences during the first 12 h were tied to the initial conditions; however, beyond that lead time, another process was driving the position divergence.

To evaluate what processes contributed to the position divergence beyond 24 h, it is necessary to explore the evolution of the horizontal differences in the steering wind. Figure 8 shows the horizontal distribution of the mean 850–300-hPa wind vectors and the difference in zonal wind between the ExtremeEast and AllWest members. In each of these figures, the TC vortex has been removed using the Galarnau and Davis (2013) TC removal method with a radius of 4$^\circ$. Subtracting out the TC vortex helps to remove the differences in steering wind related to systematic differences in the forecast TC position. At 0 h, the steering wind near Sandy was primarily east of south and was characterized by a statistically significant positive difference on the order of 0.9 standard deviations (0.5 m s$^{-1}$) along 17$^\circ$N in between

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4 Here, across track was defined with respect to the ensemble-mean motion, where positive is to the right of track.
80°–70°W (Fig. 8a). This region of perturbation westerly wind helps to explain the initial eastward motion of the ExtremeEast members relative to the AllWest members. Although there are other horizontal locations that were characterized by a statistically significant difference, these regions were not coherent with time and were, thus, were not considered further. At all lead times, the meridional wind differences were also sporadic in both space and time (not shown).

By 12 h, the zonal wind differences that were near Sandy at 0 h became statistically insignificant; however, a new region of statistically significant positive zonal wind differences on the order of 1.0 standard deviations (1.2 m s⁻¹) emerged that was centered near 28°N, 75°W, an area that Sandy moved into during the subsequent forecast (Fig. 8b). This region roughly corresponded to an upper-tropospheric ridge on the poleward side of Sandy, which suggests that the existence of the ExtremeEast members may be related to the variability in the ridge amplification. Beyond this lead time (24, 36 h), the region of positive zonal wind differences move poleward with Sandy and increased in magnitude with increasing lead time, such that the 36-h differences were 1.8 standard deviations (2.7 m s⁻¹; Figs. 8c,d). Subsequent lead times were characterized by an expanding region of statistically significant positive zonal wind differences (not shown), suggesting that Sandy moved into an environment with positive zonal wind anomalies in the ExtremeEast members, which moved Sandy away from the U.S. coastline and prevented the TC from interacting with the significant midlatitude trough discussed in section 3.

Given that Sandy’s track forecast in the ExtremeEast members appeared to be tied to the amplification of the downstream ridge, the remainder of this section focuses on understanding what processes modulated the ridge evolution. Although there are a number of different dynamical quantities that could be used to describe the ridge amplitude, this study evaluates the differences in the 340-K PV because this quantity is conserved under adiabatic and frictionless conditions (e.g., Hoskins et al. 1985) and has proved useful for diagnosing the sources and evolution of forecast errors (e.g., Davies and Didone 2013). Moreover, this isentropic level was characterized...
by the largest PV differences at 24 h and the variability of the PV at this level has a large impact on the 850–300-hPa steering flow.

Figure 9 shows the evolution of the ensemble mean and differences in 340-K PV as a function of lead time. At 0 h, there was no systematic difference between these two subsets of members (Fig. 9a); rather it appears a few random points passed statistical significance by chance. Beginning at 6 h, a region of statistically significant forecast differences in excess of 0.8 standard deviations (0.6 PVU) developed to the northwest of Sandy along the eastern edge of the corridor of higher PV associated with the trough over the Gulf of Mexico (Fig. 9b). At 6 hours later, the region of positive differences to the northwest of Sandy increased in area and magnitude, such that the maximum difference was 1.4 standard deviations (1.1 PVU; Fig. 9c), which suggests that the ExtremeEast members were characterized by less amplification of the ridge on the poleward side of Sandy. Moreover, this region of higher PV in the ExtremeEast members was on the northern periphery of the positive steering wind anomalies described previously (cf. Fig. 8b). By 24 h, the differences in PV exceeded two standard deviations (1.8 PVU) and were concentrated on the northwest side of the building anticyclone to the north of Sandy (Fig. 9d). Beyond 24 h, the region of positive differences continued to expand to the north and west of Sandy along the anticyclone, which was beginning to cyclonically wrap with the Gulf of Mexico trough (Figs. 9e,f). The negative–positive difference...
centered near 28°N, 78°W at 48 h appeared to be related to the more eastern position of Sandy in the Extreme-East members. Finally, the differences associated with the ridge expanded upstream into the midlatitudes with time, which allowed the upstream midlatitude trough to move farther east in the Extreme-East members.

Although the largest PV differences occurred around 250 hPa, which was above the vertical layer used to compute the steering flow (850–300 hPa), variations in the PV at this level were associated with perturbation zonal wind throughout the troposphere. The relationship between variations in the 340-K PV and the steering winds was determined by computing the leading empirical orthogonal function (EOF) of the ensemble 24-h 340-K PV field over the domain shown in Fig. 10a and regressing the resulting normalized principal components onto the ensemble 24-h zonal steering wind at each horizontal grid point. Figure 10a shows the regression of the first principal component onto the ensemble 340-K PV. This EOF explained 17% of the variance in PV over this domain, and was characterized by a negative–positive dipole that paralleled the large PV gradient from the Florida Panhandle through southern Virginia. This pattern essentially reflected the northwest extent of the amplifying ridge to the north of Sandy. This pattern also bears striking resemblance to the PV differences in Fig. 9d, thus the EOF identified here is of interest. It is worth noting that the elements of the principal components associated with the ExtremeEast members all exceed 0.9.

Regressing the principal components onto the zonal winds confirmed that variations in the PV at this level were associated with troposphere-deep changes in the zonal wind. Figure 10b shows the regression of the principal components of the PV EOF pattern onto the 24-h 850–300-hPa zonal wind. Here, the statistical significance of the regressed zonal wind was determined using the method outlined in Torn and Hakim (2008). Increasing the PV by 0.8 PVU to the south of the maximum in PV through the southeastern United States was associated with a >0.5 m s⁻¹ westerly wind perturbation to the east of Florida in between 83° and 70°W, which was mostly to the south of the PV EOF (cf. Fig. 10a), with smaller regions of negative perturbation zonal wind to the north over Virginia. This pattern of zonal wind perturbation was consistent with the notion that shifting the region of higher PV to the south was akin to having a positive PV anomaly at those locations, which in turn was associated with perturbation cyclonic winds. Taking a zonal cross section along 30°N and regressing the principal component onto the zonal component of the wind at various vertical levels indicates that the largest changes in the zonal wind (2 m s⁻¹) were near the level of the PV.

FIG. 10. (a) First EOF of the ensemble 24-h 340-K potential vorticity over this domain (shading, PVU). The contours are the ensemble-mean 340-K PV. (b) The regression of the principal components of the first EOF of 340-K PV onto the 24-h 850–300-hPa mean wind (shading, m s⁻¹), with vectors denoting the ensemble-mean wind. (c) The regression of the PV EOF principal components onto the zonal wind along the thick black line shown in (b) (m s⁻¹). The contours denote the ensemble-mean zonal wind, where the dashed values denote negative values.
difference (250–300 hPa); however, a one standard deviation change to the PV field was associated with a >0.75 m s$^{-1}$ change in the zonal wind down to 700 hPa (Fig. 10c). As a consequence, it appears that variations in the PV along the tropopause to the northwest of Sandy influenced the deeper-layer wind, which motivates determining what caused the differences in the PV evolution in this region.

During the first 24 h of this forecast, there was negligible precipitation along the northern edge of this ridge where the PV differences were observed, thus, it would appear that the amplification of this ridge was tied to advection of lower PV from the south, which could have resulted from differences in the irrotational wind between the ExtremeEast and AllWest members. Figure 11 shows that while the differences in the 0-h advection by the irrotational wind was noisy, by 6 h, a coherent region of negative differences on the order of 0.6 standard deviations (1.0 PVU day$^{-1}$) formed off the eastern coast of Florida near 30°N, 78°W, with a region of positive differences to its northwest of comparable magnitude. This dipole of PV advection differences was indicative of the PV advection occurring farther to the southeast in the ExtremeEast members, consistent with weaker ridge amplification. At 12 h, the ExtremeEast members were characterized by >1.2 standard deviation less (1.3 PVU day$^{-1}$) advection at a similar location, with smaller amplitude positive differences to the northwest (Fig. 11c). Moreover, averaging the PV advection in between 25°–35°N and 85°–75°W, which is akin to the interaction metric used in Archambault et al. (2013), indicates that the area-average PV advection in the ExtremeEast members was 0.9 standard deviations less than the AllWest members (not shown), suggesting that the ExtremeEast members had less forcing for ridge amplification.

Given that the 340-K PV differences between the ExtremeEast and AllWest members were not statistically significant at 0 h, this suggests that the differences in PV advection were due to the irrotational wind magnitude. Similar to PV advection, Fig. 12 indicated little systematic difference in 0-h 340-K irrotational wind magnitude between the ExtremeEast and AllWest members. Six hours later, large regions of negative irrotational wind differences appeared to the north and west of Sandy over northern Florida, which roughly corresponds to the region of lower PV advection (cf. Fig. 11b), with a maximum value of 1.2 standard deviations (1.4 m s$^{-1}$; Fig. 12b). After 12 h, a large region of negative irrotational wind magnitude differences on the order of 1.0 standard deviations existed from Florida along the East Coast of the United States toward North Carolina (Fig. 12c). This supports the idea that the ExtremeEast members were characterized by less advection of PV advection due to weaker irrotational wind differences through the forecast. Moreover, the lack of irrotational wind differences at 0 h suggests that some physical process was generating the wind difference during the first 6–12 h of the forecast.
An alternative method of obtaining the origin of the irrotational wind difference is to evaluate differences in the upper-tropospheric divergence, which has the benefits that it can describe differences in divergent vector wind without concern about direction, and that it is a prognostic variable in GFS. Figure 13 shows the differences in 340-K divergence averaged within 300 km of each grid point. At 0 h, the ExtremeEast members were characterized by a region of decreased divergence along the western and northern periphery of the region of maximum ensemble-mean divergence to the north of Sandy along 24°N. The maximum differences, which were on the order of 1.2 standard deviations (0.9 x 10^{-5} s^{-1}), also coincided with the northern section of the precipitation shield seen in Fig. 3 and were associated with the warm front. After 3 h, the region of statistically significant divergence differences increased in areal extent and magnitude while moving poleward by 50 km, which was consistent with the distance Sandy moved over the same period (Fig. 13b). Beyond 3 h, the negative divergence difference continued to propagate to the northwest (Figs. 13c,d), such that the region of maximum divergence difference is located to the south of the maximum irrotational wind magnitude difference (cf. Figs. 12c and 13d). This result suggests that the ExtremeEast members were characterized by less divergence in the initial conditions; however, the further decrease during the first 3 h suggests some other process was helping to weaken the divergence further.

For many of the grid points where the 0-h 340-K divergence differences were statistically significant, the precipitation was a combination of the cumulus parameterization and grid-resolved processes (cf. Fig. 3); therefore, either model component could have been responsible for the divergence differences. To evaluate the relative role of these two aspects of the model, vertical profiles of area-averaged quantities that modulate grid-scale precipitation processes and the output from the cumulus parameterization were computed within the box shown in Fig. 13a, which includes the 0-h region of anomalous divergence in the ExtremeEast members. Over this box, the difference in 0–3-h area-average cumulus heat and momentum tendency was small (generally less than 0.3 standard deviations) and not statistically significant at any level (not shown); therefore, the differences must have been related to grid-scale processes.

At the grid scale, the condensational heating rate is proportional to the vertical moisture flux convergence, which depends on the vertical profile of the product of

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5 Recall that the divergence is the derivative of the irrotational wind.

6 Averaging within 300 km of each grid point helps to filter the noise that comes from taking derivatives, yet still retains the important signal.

7 Assuming that the parcel is at saturation, which is true in the region near the warm front (cf. Fig. 2).
the vertical motion and the water vapor mixing ratio. Figure 14 shows the vertical profile of 0-h divergence, vertical motion, and the water vapor mixing ratio averaged over the box shown in Fig. 13a. In the ensemble mean, there was upward vertical motion between 700 and 150 hPa, with a maximum value near 400 hPa (Fig. 14b), with divergence above the vertical motion maximum and convergence below (Fig. 14a). The ExtremeEast members were characterized by less vertical motion throughout the column, with the differences maximized at 400 hPa (1.3 standard deviations, 0.17 Pa s$^{-1}$); the differences were statistically significant from 450 to 300 hPa. For divergence, the ExtremeEast members were characterized by anomalous divergence between 600 and 500 hPa and anomalous convergence between 300 and 250 hPa on the order of one standard deviation (both layers statistically significant); this pattern is opposite of the ensemble-mean pattern. In addition, the ExtremeEast members had two layers where the water vapor mixing ratio was at least one standard deviation less than the AllWest members and the differences were statistically significant: 1000–800 and 500–300 hPa (Fig. 14c). The combination of the lower water vapor mixing ratio in the same location as lower vertical motion in the ExtremeEast members led to the ExtremeEast members having roughly half the vertical moisture flux of the AllWest members (not shown); therefore, the vertical gradient was lower, leading to less latent heat release and decreased divergence higher in the troposphere. Moreover, the 0–3-h precipitation averaged over this box, which is a proxy for the amount of latent heat release, was 5.7 mm for the AllWest members, compared to 3.7 mm for the ExtremeEast members (the difference is statistically significant at the 95% confidence level), which further suggests the ExtremeEast members were characterized by less latent heat release. Three hours later, qualitatively similar differences were obtained for a box shifted to the north by 0.5° (not shown); therefore, the ExtremeEast members were characterized by lower latent heat release in this

Fig. 13. Normalized difference in the (a) 0-, (b) 3-, (c) 6-, and (d) 12-h 340-K divergence averaged within 300 km of each horizontal grid point between the ExtremeEast and AllWest members (shading; units: standardized anomaly). Dotted regions indicate where the difference is statistically significant at the 95% confidence level. The contours denote the ensemble-mean 340-K divergence (10$^{-5}$ s$^{-1}$).
region during the first few hours of the forecast, which
was associated with less irrotational wind in the upper
troposphere.

Similar to the 340-K divergence differences seen in
Fig. 13a, the ExtremeEast members were characterized
by lower midtropospheric water vapor mixing ratio over
a larger horizontal area. Figure 15 indicates that the 0-h
450-hPa water vapor mixing ratio was more than 0.8
standard deviations (0.3 g kg$^{-1}$) lower in the ExtremeEast
members along 24°N (the warm side of the midtropospheric
warm front). These differences farther expanded to the
north and west of Sandy by 3 h (Fig. 15b). As a conse-
quence, it appears that errors in either the moisture
or divergent wind field could be responsible for the
subsequent divergence differences.

Although the physical mechanism described in this sec-
tion is physically plausible, the results presented so far
do not directly imply that variability in the 0-h midtropospheric
water vapor, or upper-tropospheric divergence directly
impact the subsequent divergent outflow and structure of
the ridge poleward of Sandy. To isolate the role of 450-hPa
water vapor mixing ratio and 250-hPa divergence varia-
tions on the evolution of the ridge in the forecast, the sen-
sitivity of the divergence, 340-K irrotational wind, and PV
forecasts was calculated from the ensemble sta-
tistics. Here the sensitivity was computed via

$$\frac{\partial x_i}{\partial J} = \frac{\text{cov}(x_i, J)}{\text{var}(J)},$$

(2)
where $\mathbf{x}_i$ was the ensemble estimate of either the 340-K divergence, irrotational wind magnitude, or PV at each horizontal location $i$ and lead time, $\mathbf{J}$ was either the ensemble estimate of the 0-h 450-hPa water vapor mixing ratio averaged over the box shown in Fig. 15a, or the 250-hPa divergence averaged over the box in Fig. 13a. Regions of nonzero $\partial x_i / \partial l$ indicate where errors in the 0-h water vapor mixing ratio or divergence to the south of the midtropospheric warm front produced a nonzero impact on the subsequent irrotational wind and PV forecast. Statistical significance was determined using the method outlined in Torn and Hakim (2008). It is worth pointing out that these two metrics were weakly correlated ($r^2 = 0.15$); therefore, these two metrics represent relatively distinct analysis errors.

The sensitivity calculations support the aforementioned hypothesis that decreasing the 0-h upper-tropospheric water vapor mixing ratio on the warm side of the front led to a less amplified ridge in the forecast. In particular, increasing (decreasing) the 0-h mixing ratio by one standard deviation (0.14 g kg$^{-1}$) was associated with an up to $4.0 \times 10^{-5}$ (0.8 standard deviation) increase (decrease) in 3-h divergence at 25$^\circ$N along the poleward side of the region of maximum divergence (Fig. 16a). This region roughly corresponded to the location of negative divergence differences in the ExtremeEast members (Fig. 13b). In addition, the 0-h water vapor change led to up to 1.0 m s$^{-1}$ (0.7 standard deviation) increase (decrease) in the 12-h irrotational wind magnitude between 30$^\circ$ and 35$^\circ$N to the north of the baroclinic zone (Fig. 16c) and a 0.4-PVU reduction (0.6 standard deviations) in 340-K PV along the northern edge of the ridge (Fig. 16e), which indicated that increasing (decreasing) the water vapor mixing ratio led to a more (less) amplified ridge in the forecast. These results strongly imply that variability in the upper-tropospheric water vapor mixing ratio along the baroclinic zone to the north of Sandy was associated with differences in the upper-tropospheric divergence and the subsequent amplitude of the ridge poleward of Sandy, leading to modifications of the steering flow.

In comparison to the water vapor mixing ratio, one standard deviation changes to the 0-h 250-hPa divergence produced somewhat smaller impacts on the PV. A one standard deviation change to the 0-h divergence ($1.1 \times 10^{-5}$ s$^{-1}$) was associated with a $0.5 \times 10^{-5}$ change in 3-h divergence at individual grid points over a region slightly to the west of the sensitivity region for 450-hPa water vapor (Fig. 16b). At 12 h, the sensitivity of the irrotational wind to the divergence was $0.1$–$0.2$ m s$^{-1}$ per standard deviation lower than the sensitivity to water vapor (Fig. 16d), while the 12-h 340-K PV sensitivity was $20$–$25$% lower (Fig. 16f). Moreover, increasing the 450-hPa water vapor mixing ratio by one standard deviation was associated with a 0.11-PVU decrease in the 12-h PV averaged over the polygon that encompasses the northern edge of the ridge (showed in Fig. 16e), compared to a 0.074-PVU decrease for a one standard deviation increase in the 250-hPa divergence (difference statistically significant at the 95% confidence level). As a consequence, it appears that the ridge amplitude was more sensitive to one standard deviation changes in the initial-time water vapor compared to divergence.

5. Summary and conclusions

This manuscript explored the source of the erroneous eastward track of Hurricane Sandy that was present within an 80-member GFS ensemble initialized five days prior to landfall. The origin of these outlier solutions was determined by comparing the mean of the subset of ensemble members that forecasted Sandy to be at the far eastern side of the envelope against the ensemble mean of the subset of ensemble members that predicted Sandy to the west of the ensemble mean position. The forecast differences were then compared over several lead times to determine the dynamical processes responsible for that behavior.

The ensemble members that predicted Sandy would move out to sea away from the United States were characterized by differences in physical processes along a region of isentropic lift and a developing warm front to the north of Sandy. The eastern members had lower water vapor mixing ratio and less divergence on the warm side of the warm front; therefore, this region was characterized by weaker vertical moisture flux convergence and, hence, upper-tropospheric latent heat release. This lower heating rate fed back to produce even weaker upward vertical motion and less divergence in the upper troposphere. The divergent wind acted to amplify the ridge to the north of Sandy via advection of lower PV; thus, the lower irrotational wind speed in the eastern forecasts produced a less amplified ridge, which was associated with a perturbation westerly wind throughout much of the troposphere that eventually pushed Sandy away from the United States. Regression calculations suggest that one standard deviation changes to the area-averaged 0-h water vapor mixing ratio (0.14 g kg$^{-1}$) produce a larger impact on the PV forecast compared to one standard deviation changes to the area-averaged 0-h divergence ($1.1 \times 10^{-5}$ s$^{-1}$), thus, the forecast was slightly more sensitive to the initial moisture profile.

The results of this study highlight an important source of uncertainty in the midlatitude flow, which have received relatively little attention outside of the ET
literature. In particular, these results imply that the specification of the thermodynamic profile within the warm sector of the cyclone can have a significant influence on the depth and strength of vertical motion and, thus, the upper-tropospheric divergence, which can perturb the midlatitude waveguide and propagate downstream (e.g., Bosart and Lackmann 1995; Dickinson et al. 1997; Riemer et al. 2008; Torn 2010; Grams et al. 2011).

Future work will evaluate the importance of upscale error growth in the warm sector of a cyclone over a larger set of cases.

In addition, the conclusions of this study complement the Bassill (2014) result that different cumulus parameterizations can give rise to the westward and eastward forecast tracks. In this study, the eastward track resulted from the lack of divergent outflow due to the specification of the thermodynamic profile and divergence within precipitating regions to the north of Sandy. Varying the
cumulus parameterizations can also produce different divergence profiles due to how the individual schemes handle processes like entrainment, detrainment, or the trigger. Taken together, these two studies suggest that variability in the divergent outflow can have a significant impact on track forecasts when an upper-tropospheric trough is located nearby.

Although there was near-continuous sampling of Sandy by aircraft four days prior to landfall, these results suggest that synoptic sampling of the region to the north of Sandy could have helped decrease the variability associated with forecasts initialized at earlier times. Operational model forecasts could have been characterized by large errors in the region of water vapor sensitivity because the extensive cirrus canopy to the north of Sandy would limit the amount of satellite radiation data that could have been assimilated. One method of obtaining observations would have been to task the NOAA G-IV to deploy dropwindsondes on the northern side of Sandy in the region of water vapor and divergence sensitivity. At individual grid points, the 0-h ensemble standard deviation within the region of 500-hPa water vapor mixing ratio sensitivity was 0.3 g kg$^{-1}$, which is comparable to the observation error standard deviation for dropwindsonde moisture observations within the NCEP GFS system. Instead, the first mission with this aircraft took off at 1800 UTC 26 October when most operational models had a fairly accurate forecast of the landfall position. These results suggest that earlier aircraft missions could have been beneficial and that future missions should consider sampling the nascent warm sector of poleward-moving TCs. Although it is not possible to test this hypothesis with observations, an alternative method of evaluating the hypothesis in future work could be to perturb the initial conditions in GFS within the regions of high sensitivity and observe the outcome in terms of Sandy’s forecast.

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