A Comparison of the Downstream Predictability associated with ET and Baroclinic Cyclones

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ABSTRACT

The impact of the extratropical transition (ET) of tropical cyclones and baroclinic cyclogenesis in the Western North Pacific (WNP), Atlantic, and Southern Indian Ocean (SIO) basins on the predictability of the downstream midlatitude flow is assessed using 30 y of cases from the Global Ensemble Forecast System (GEFS) Reforecast Version 2. In all three basins, ET is associated with statistically larger 500 hPa geopotential height forecast standard deviation (SD) compared to the forecast climatology. The higher SD values originate from where the TC enters the midlatitudes and spreads downstream at the group velocity of the associated wave packet. Of the three basins, WNP ET is associated with the largest amplitude and longest-lasting SD anomalies. Forecasts initialized 2-4 d prior to the onset of ET have larger SD anomalies compared to forecasts initialized during or after the onset of ET. By contrast, the region of positive SD anomaly associated with winter baroclinic cyclones is confined to the upstream trough, with fall cyclones exhibiting some downstream propagation characteristics similar to ET. ET cases with the larger downstream SD anomaly are characterized by a more amplified ridge downstream of the TC as it enters the midlatitudes. By contrast, ET cases with large TC position variability at the onset of ET or upper-tropospheric PV advection by the irrotational wind are not characterized by significantly larger downstream SD.
1. Introduction

Tropical Cyclones (TCs) that move poleward into the midlatitude storm track often undergo the process of extratropical transition (ET), which can lead to the rapid amplification of the downstream midlatitude flow (e.g., Jones et al. 2003; Agusti-Panareda et al. 2004; Harr and Dea 2009; Grams et al. 2013a; Archambault et al. 2013). The amplification of the downstream flow is often attributed to the development of a ridge immediately downstream of the TC (e.g., Bosart and Dean 1991; Harr and Elsberry 2000; Riemer et al. 2008; Riemer and Jones 2010; Torn 2010), which in turn acts like an impulsive disturbance to the midlatitude waveguide (e.g., Torn and Hakim 2015), leading to the development of a Rossby wave packet associated with new ridges and troughs downstream (e.g., Simmons and Hoskins 1979; Chang and Orlanski 1993; Quinting and Jones 2016). The amplification of this ridge can be explained via the destruction of potential vorticity (PV) by latent heat release associated with the TC and along the baroclinic zone that often forms on the poleward side of the TC (e.g., Torn 2010), as well as the indirect impact of the latent heat release, which is the advection of PV by the irrotational wind (e.g., Riemer and Jones 2010; Archambault et al. 2013, 2015).

The above processes may allow a TC interacting with the midlatitudes to introduce significant uncertainty into the downstream state. Indeed, there have been several case studies that document large downstream forecast errors and/or uncertainty that have originated with upstream ET events (e.g., Jones et al. 2003; Harr et al. 2008; Anwender et al. 2008; Reynolds et al. 2009; Anwender et al. 2010; Pantillon et al. 2013). More recently, Aiyyer (2015) and Quinting and Jones (2016) used a large climatology of forecasts (i.e., reforecasts) to show that recurving TCs are associated with an anomalous “plume” of higher forecast standard deviation that originates at the longitude where the TC enters the midlatitudes and spreads downstream; therefore, the impact of ET is not
just limited to a few case studies. In addition, Quinting and Jones (2016) show that ET events in the western North Pacific lead to the largest downstream standard deviation impact, with cases with a Rossby wave packet exhibiting larger forecast standard deviation compared to cases without a Rossby wave packet. Forecast uncertainty can arise from a number of different processes, including how the ET system phases with the midlatitude flow, including the upstream trough (e.g., Klein et al. 2002; Harr et al. 2008; Anwender et al. 2008; Torn and Hakim 2009; Riemer and Jones 2010; Grams et al. 2013b; Riemer and Jones 2014; Grams et al. 2015) or via diabatic processes either through direct processes or via upper-level divergent outflow that can impact the waveguide (e.g., Riemer et al. 2008; Harr and Dea 2009; Hodyss and Hendricks 2010; Torn 2010; Keller et al. 2014; Archambault et al. 2015).

Waveguide perturbations and the subsequent amplification of the downstream midlatitude flow is not limited to instances where TCs enter the midlatitudes. Baroclinic cyclogenesis is often accompanied by an impulsive disturbance along the midlatitude waveguide that originate from upstream eddies that subsequently amplify the downstream flow (e.g., Orlanski and Katzfey 1991; Chang 1993, 1999; Hakim 2003; Souders et al. 2014), which often occurs in conjunction with baroclinic conversion and/or latent heat release, particularly in the warm sector of cyclones (e.g., Wernli 1997; Joos and Wernli 2012; Chagnon et al. 2013; Martinez-Alvarado et al. 2014; Teubler and Riemer 2016). Moreover, Torn and Hakim (2015) showed that on average baroclinic cyclones were associated with somewhat weaker wave packets in the western North Pacific than ET, while baroclinic cyclones in the Atlantic were associated with stronger and longer-lasting packets. During these baroclinic cyclone events, downstream forecast uncertainty could originate from the upstream feature as well as how the process of cyclogenesis amplifies the wave packet. Similar to ET, downstream forecast errors have been shown to exhibit sensitivity to upstream features, such as a trough (e.g., Langland et al. 2002; Majumdar et al. 2010; Zheng et al. 2013; Chang et al. 4
2013) and/or to uncertainty in latent heat release within the poleward flow of heat and moisture downstream of the surface cyclone (e.g., Bosart and Lackmann 1995; Dickinson et al. 1997; Zhang et al. 2003; Dirren et al. 2003; Davies and Didone 2013; Rodwell et al. 2013; Lamberson et al. 2016). Moreover, midlatitude forecast errors tend to propagate similar to a Rossby wave packet over relatively long distances (e.g., Szunyogh et al. 2000; Langland et al. 2002; Hakim 2005), though it is intriguing that Grazzini and Vitart (2015) showed that cases with wave packets tend to be associated with smaller medium-range forecast errors than cases without one.

The purpose of this study is to expand upon the Aiyyer (2015) and Quinting and Jones (2016) studies that documented the downstream predictability ET in the western North Pacific (WNP), Atlantic, and Southern Indian Oceans (SIO) by comparing the resulting 500 hPa geopotential height standard deviation (SD) anomaly to baroclinic cyclones that occur in those same basins using a multi-year dataset of ensemble forecasts and employing the null hypothesis. For both ET and baroclinic cyclones separately, the null hypothesis is that the SD for these cases is indistinguishable from a random set of forecasts from the same time of year. In addition, the null hypothesis that the ensemble standard deviation downstream of ET is indistinguishable from baroclinic cyclones is also evaluated, which might be expected given that Torn and Hakim (2015) found that ET and baroclinic cyclones have similar downstream midlatitude responses. The role of forecast lead time is also evaluated by repeating the analysis for forecasts initialized at various times before, during, and after either the onset of transition, or cyclogenesis, respectively. Finally, this study will evaluate whether several dynamical processes that have been linked to the amplitude of the ET downstream response impact the downstream predictability by subdividing the ET cases based on these processes and evaluating whether one set of cases is characterized by a larger downstream standard deviation.
The remainder of the paper proceeds as follows. Section 2 describes the datasets and methods used in this particular study. Composite SD anomalies for ET and baroclinic cyclones in each basin are presented in section 3. In section 4, the ET cases are subdivided by various properties to understand what processes give rise to larger downstream forecast SD. A summary and conclusions in section 5.

2. Methods

The role of ET and baroclinic cyclones on downstream forecast predictability is assessed by compositing ensemble forecasts from a large number of events over multiple years. The ensemble forecasts are taken from the Global Ensemble Forecast System (GEFS) Reforecast Version 2 dataset (Hamill et al. 2013), which consists of an 11-member ensemble (control + 10 perturbed members) initialized each day at 0000 UTC from 1985-2014. This dataset is particularly well-suited for this type of study because it contains a large number of events, and the model and ensemble initialization technique are consistent during the entire period; therefore, it removes the possibility that improvements in the modeling and/or data assimilation system over time could be responsible for the differences.

Throughout this paper, 500 hPa geopotential height SD is used as a proxy for midlatitude predictability. This choice is made because this field has a long history of being used to measure midlatitude forecast performance, particularly at the synoptic scale, which is the focus here. In order to provide an appropriate benchmark to test the above hypotheses, the ensemble SD for the cases of interest is compared to the GEFS Reforecast 500 hPa geopotential height SD “climatology” from the same 1985-2014 period, which is a function of horizontal location, forecast lead time, and day of the year. The forecast climatology is computed by considering all forecasts initialized within ± 14 days of the specific day of interest during the 30-yr period (i.e., a forecast
initialized on 1 September would be compared to a climatology of all forecasts initialized between 18 August - 15 September). Throughout this study, composite normalized forecast SD anomalies are employed, which allows for a quantitative comparison despite differences in geographical locations, forecast lead times, and times of the year. These composite anomalies are computed for a set of forecasts via:

\[ \delta \sigma(i,t,f) = \frac{\sigma(i,t,f) - \sigma_{climo}(i,t,f)}{\sigma}(1) \]

where \( \sigma \) is the mean ensemble standard deviation at horizontal location \( i \), day of the year \( d \), and forecast lead time \( f \) for the cases considered in the composite, \( \sigma_{climo} \) refers to the mean climatological forecast SD for the day of the year corresponding to each case in the composite, and \( \sigma \) is the standard deviation of the climatological forecast SD for the cases considered. The statistical significance of the anomalies is obtained via bootstrap resampling whereby alternative days are randomly selected from the forecast climatology within \( \pm 14 \) days of each day of the year used in the case composites. An anomaly is deemed statistically significant if the normalized SD falls outside the 95% confidence bounds of the distribution of normalized SD (obtained by resampling 2000 times) from random cases.

Baroclinic cyclones are identified using the method first outlined in Hakim (2003) and is summarized here. Here, baroclinic cyclones are local maxima in 1000 hPa geostrophic relative vorticity that exceed \( 10^{-4} \) s\(^{-1} \) in the 2.5\(^\circ\) resolution Climate Forecast System Reanalysis (CFSR) dataset (Saha et al. 2010). Multiple additional criteria are employed to ensure that the cases used are characterized by individual cyclone events in a certain geographical region. Any cyclone that is within 25\(^\circ\) of another identified cyclone is removed from consideration. To ensure that the same event is not identified twice, any cyclone within 15\(^\circ\) latitude and longitude of a previously identified cyclone during the previous 24 h is also removed. Finally, cyclones are only considered if they meet
the above criteria at either 0000, 0600, or 1800 UTC. This choice is made because the reforecast dataset consists of forecasts initialized at 0000 UTC. One aspect of this work is to evaluate the SD anomalies for forecasts initialized at various lag times relative to the cyclone identification time. For cyclones that first meet the criteria at 0600 (1800) UTC, the initialization time closest to this is 6 h prior (after). For cyclones that meet the criteria at 1200 UTC, it is unclear whether to use the initialization time 12 h prior to or 12 h after; therefore, these cyclones are excluded. For these composites, “t = 0” refers to the time when the cyclone exceeds the above critical value because it is meant to approximate when the cyclone reaches the mature stage. Table 1 gives the geographical domain that contains the maximum number of cyclones during the winter period for each of the basins and the number of cases.

The difference in timing between ET and winter cyclones can result in different background states (i.e., baroclinicity, waveguide structure, etc.), which in turn could impact the downstream predictability of these events. The potential role of season is addressed by applying the same algorithm to baroclinic cyclones that occur during same months where ET is most likely to occur (August, September and October in WNP and Atlantic and January-March in the SIO). In each basin, the maximum number of cyclones within any $5^\circ \times 10^\circ$ box during the autumn is much smaller than what occurs during the winter months, which does not provide a robust set of cases to evaluate the downstream predictability hypotheses. As a consequence, the autumn baroclinic cyclone composites are constructed by identifying all events that occur within a particular latitude band (given in Table 1) and shifting the grids to a common longitude, which is defined here as the longitude of the cyclone when it first exceeds the criteria. This procedure is identical to what was used in Torn and Hakim (2015).

ET cyclones were identified using methods outlined in Torn and Hakim (2015) and are summarized here. Starting from the set of all TCs contained in the Joint Typhoon Warning Center
(JTWC) WNP and SIO best track data and the National Hurricane Center (NHC) Atlantic data from 1985-2014, a TC was considered a candidate for ET if the track acquired a western component of motion (i.e., recurved) and moved poleward of 20°. For the remaining TCs, the CFSR TC position is determined by finding the minimum in 1000 hPa geopotential height each 6 h, whereby the the asymmetry parameter from the Hart (2003) cyclone phase space\(^1\) is calculated. As in Hart (2003), the onset of transition is defined as when the asymmetry parameter exceeds 10 m, which indicates the formation of thermal asymmetries associated with the cyclone, and is hereafter referred to as “t = 0”. Similar to baroclinic cyclones, only ET events that occur at 0000, 0600, or 1800 UTC are considered because of the ambiguity of assigning the closest initialization time for cases that meet the criteria at 1200 UTC. Similar to ASO baroclinic cyclones, the number of cases in any 5° × 10° box is relatively small; therefore all ET cases that occur within a latitude band are considered (bounds given in Table 1). The latitude band used in each basin contains the largest number of cases. As a consequence, all fields are shifted to a common longitude, which is defined as the TC longitude at the onset of transition. It is worth pointing out that although both baroclinic cyclones and ET can produce a downstream predictability response, the most appropriate choice for t = 0 for comparing these two phenomenon is not obvious. As was shown in Torn and Hakim (2015) and will be demonstrated later, compositing ET cases based on the onset of transition appears to produce higher amplitude anomalies than composites based on the recurvature time (e.g., Aiyyer 2015; Quinting and Jones 2016). For consistency with past work, the Hakim (2003) definition of t = 0 is used for midlatitude cyclones since it clearly identifies a downstream signal.

\(^{1}\)The asymmetry parameter is defined as the difference between the 600-900 hPa thickness averaged over a semicircle to the right of the track and left of the track.
3. Results

a. Western North Pacific (WNP)

Composites of ET in the WNP indicate a clear signal of increased ensemble SD that originates from the TC and subsequently moves downstream. Fig. 1 shows the normalized 500 hPa height forecast SD anomaly for ET cases initialized 2 d prior to the onset of transition. At 0 h, positive forecast SD anomalies are tied to the 500 hPa reflection of the TC near 25\degree N, 145\degree E and negative SD anomalies co-located with the anticyclone to the northeast (Fig. 1a). Furthermore, the upstream midlatitudes is characterized by SD differences near zero, suggesting that the typical forecast exhibits little evidence of greater than typical forecast uncertainty originating upstream during ET. By 48 h (the onset of ET), the positive SD anomaly translates northward, but is still mainly confined to the area immediately surrounding the TC (located at 32\degree N, 147\degree E), whereas there is a region of statistically lower SD to the south and east of the TC associated with anticyclones (Fig. 1b). Following the onset of transition, large regions of statistically higher SD appear downstream from where the TC enters the midlatitudes (150\degree E). At 96 h (48 h after the onset of transition), normalized SD anomalies exceeding 0.6 are present both at the mean location of the transitioning TC, but also with the downstream trough located near 160\degree W (Fig. 1c). Beyond that time, the region of statistically higher SD expands across the Pacific Basin, reaching central North America (90\degree W) by 144 h (96 h after onset of ET; Fig. 1d) and east of North America by 192 h (Fig. 1e) before becoming mostly insignificant at 240 h (Fig. 1f). Overall, these results suggest that the downstream 500 hPa height forecasts are significantly more uncertain than a typical forecast following the time when a TC enters the midlatitudes and the process of ET acts as a source of uncertainty.

\[^2\]To facilitate comparison, the geography on the ET figures is oriented such that the cyclone position at \( t = 0 \) matches the winter cyclone position at \( t = 0 \).
downstream forecast uncertainty, rather than a feature that amplifies upstream midlatitude forecast uncertainty.

In order to facilitate a quantitative comparison between different initialization times, basins, and cyclones, normalized SD differences are averaged over the latitude band characterized by the largest SD anomalies (30-60°N; Fig. 2c). Forecasts initialized 2 d prior to the onset of ET (same as described above) are characterized by a “plume” of higher SD that originates from where the TC enters the midlatitudes and extends downstream, which is similar to the distribution of downstream forecast SD seen in the case studies described by Harr et al. (2008) and Anwender et al. (2008) and is largely consistent with the climatologies presented in Aiyyer (2015) and Quinting and Jones (2016). As a general rule, the downstream SD becomes statistically higher than a typical forecast following the passage of the composite analysis Rossby Wave packet, which is denoted by tracing the maxima in the 500 hPa meridional wind amplitude. This result implies that higher forecast uncertainty is not associated with the arrival of the wave packet, but with how the individual ridges and troughs evolve after the peak of the wave packet passes. Tracing the maximum in normalized SD implies that forecast uncertainty is spreading downstream at 20 m s⁻¹, which is consistent with the WNP ET cyclone group velocity obtained in Torn and Hakim (2015).

Although forecasts initialized at various times prior to the onset of ET are also characterized by positive SD anomalies, the amplitude of SD plume is a function of the initialization time. Whereas forecasts initialized 4, 3, and 2 d prior to the onset of ET have maximum SD anomalies between 0.3-0.4, the maximum SD anomalies for forecasts initialized closer to the onset of ET are comparatively smaller. In particular, forecasts initialized 1 d prior to ET onset have a SD plume; however, the differences are maximized where the TC enters the midlatitudes (Fig. 2d). Beyond 180°, the normalized SD is generally below 0.3, which is smaller than earlier initialization times. By contrast, forecasts initialized at 0 (Fig. 2e) and 1 d after the onset (Fig. 2f) are characterized
by relatively small regions where the downstream SD is statistically different from zero. In the forecasts initialized at the onset of ET, the largest SD differences occur where the TC enters the midlatitudes, while the differences downstream generally do not exceed 0.2. Moreover, forecasts initialized 1 d after ET onset generally exhibit small, statistically insignificant differences in the downstream SD. This result suggests that forecasts downstream of ET are most unpredictable when initialized prior to the onset of ET. Once the process of ET begins and the TC begins to interact with the midlatitude flow, the downstream forecast SD is not distinguishable from a typical forecast.

The resulting downstream SD anomaly exhibits some sensitivity to the choice of \( t = 0 \). Fig. 3 shows the normalized SD anomalies averaged between 30-60\(^\circ\)N for various initialization times prior to and during the recurvature point, which was used as \( t = 0 \) within Aiyyer (2015); Quinting and Jones (2016). Torn and Hakim (2015) found that recurvature tends to occur on average 1-2 d prior to the onset of ET; therefore, forecasts initialized 1 d prior to recurvature (Fig. 3b) are roughly equivalent to forecasts initialized 2 or 3 d prior to the onset of ET (Fig. 2b,c). Although the SD anomalies based on recurvature time exhibit the same plume shape as SD anomalies based on the onset of ET, the amplitude of the downstream SD anomalies from the recurvature-based composite are generally lower. For example, the downstream SD anomalies are generally 30% lower with forecasts initialized 1 d prior to recurvature compared to forecasts initialized 2 d prior to ET onset, which likely occurs because the recurvature time does not necessarily denote when the TC starts to interact with the midlatitudes. As a consequence, the composite signal can be washed out, yielding lower amplitude signals (e.g., Torn and Hakim 2015). Nevertheless, this result indicates that ET onset is more appropriate time to use to measure the downstream SD anomaly associated with ET.

Although previous work has demonstrated that winter baroclinic cyclones exhibit downstream Rossby wave propagation (i.e., akin to ET), there is little evidence of a plume of higher SD orig-
inating from the cyclogenesis location. Fig. 4 displays a Hovmoller diagram of SD anomaly for various initialization lag times with respect to the baroclinic cyclone $t = 0$ time (see above for definition) averaged between 30-60°N. Most initialization times are characterized by a region of positive SD anomalies that originates upstream of the cyclone’s $t = 0$ location and appears to be associated with the upstream trough involved in cyclogenesis, suggesting that the precursor troughs are relatively uncertain relative to the forecast climatology. With the exception of forecasts initialized 4 d prior to the cyclone, each initialization time is characterized by a region of negative SD anomaly co-located with the ridge immediately downstream of the cyclone, suggesting that the downstream ridge is characterized by lower uncertainty relative to climatology. Unlike ET, locations that are downstream of the cyclogenesis location are not characterized by a positive SD anomaly that originates from where the cyclone develops. Following $t = 0$, the only locations with a positive SD anomaly is associated with cyclone’s trough. Remarkably, locations east of 180° are characterized by negative SD anomalies following the passage of the wave packet (denoted by the meridional wind anomalies) for forecasts initialized -2, -1, +0 and 1 d after the cyclone development time. This result suggest that forecasts over the eastern Pacific Ocean are more predictable following western Pacific cyclogenesis event. One possible reason for this result is the structure of the midlatitude waveguide in the Pacific Basin during this time of the year. During the winter, the speed of the midlatitude jet decreases in the central Pacific basin, which results in frequent Rossby wave breaking. As a consequence, the uncertainty that might be introduced by cyclogenesis cannot propagate as far to the east.

One method of trying to quantify the role of the waveguide structure on downstream predictability is to repeat the above calculations for baroclinic cyclones that occur during August-October, which is the climatological maximum in WNP ET. Composite SD anomalies for these cases exhibit properties that appear to be a mixture of the ET and winter baroclinic cyclone composites.
Unlike the baroclinic winter cyclones, there is little evidence of a positive SD anomaly associated with the upstream precursor trough at any lead time. Instead, the region of positive SD anomaly appears to develop within the upstream trough around the cyclone $t = 0$ time, with a region of negative SD anomaly co-located with the ridge. This combination of SD differences suggests that upstream forecasts prior to cyclogenesis have similar predictability to a forecast during that time of year, with anomalously higher SD associated with cyclone development. Moreover, most initialization times are characterized by positive SD anomalies downstream of where cyclogenesis takes place, similar to the downstream plume seen in ET forecasts, but with a smaller amplitude. This region of increased downstream SD is most pronounced for forecasts initialized 0-2 d prior to $t = 0$ (Fig. 5c-e). Moreover, the positive SD anomaly appears to move at the group velocity of the mean wave packet and is positive following the passage of the wave packet peak (similar to ET). Overall, this result suggests that baroclinic cyclogenesis during August-October is associated with decreased downstream predictability, though not to the same extent as ET.

**b. Atlantic**

Downstream forecasts following ET and baroclinic cyclone development in other basins exhibit similar behavior to the WNP. For brevity, only the forecasts initialized 2 d prior to and at $t = 0$ are shown here; results for other times are qualitatively consistent (not shown). For ET, forecasts initialized at these two different times are characterized by positive SD anomaly starting at the longitude and times where the TC enters the midlatitudes ($60^\circ$W) that subsequently spreads downstream following the passage of the Rossby wave packet peak (Fig. 6a,b), such that forecasts over Europe ($0-30^\circ$E) are characterized by an SD anomaly $> 0.2$ between 2-6 d after ET onset. In comparison to the WNP results, the region of positive SD anomaly does not extend as far downstream from where ET occurs, which is likely due the fact that the composite wave packet
associated with Atlantic ET events does not propagate downstream as far as the WNP composite
wave packet (e.g., Torn and Hakim 2015). Furthermore, the amplitude of the downstream SD
anomaly for Atlantic ET is generally lower than for WNP ET, suggesting that on average WNP
ET events introduces more forecast uncertainty than Atlantic ET. Finally, it appears that forecasts
initialized 2 d prior to and during the onset of ET have comparable amplitude SD anomalies, thus
downstream forecasts initialized closer to the onset of ET are not necessarily more predictable in
this basin, though forecasts initialized 1 d after the onset of ET have negligible SD anomalies (not
shown).

Baroclinic cyclone forecasts in the Atlantic exhibit similar downstream SD evolution to what is
seen for WNP. For both lead times, the positive SD anomalies during winter baroclinic cycloge-
nesis follows the cyclone’s upstream trough originates 45° upstream of the cyclogenesis location
(Fig. 6c,d). This positive SD anomaly region becomes generally statistically insignificant within
48 h of t = 0 for both initialization times, suggesting that baroclinic cyclone development in the
Atlantic does not lead to a longer-lasting increase in forecast uncertainty. Instead, the regions of
both positive and negative SD anomalies are associated with the ridges and troughs within the
wave packet, with mainly insignificant anomalies outside of it. Finally, August-October baroclinic
cyclones exhibit similar results to their WNP counterparts, with the positive SD anomalies mainly
tied to the amplification of the upstream trough, but negative SD anomalies associated with the
downstream ridge that is statistically significant for 6 d after the cyclone develops and appears to
be tied to the onset of blocking at 30°W (Fig. 6e,f). Both initialization times exhibit a region of
positive SD anomaly east of 0°, which is beyond the extent of the mean wave packet.
c. Southern Indian (SIO)

ET forecasts in the Southern Indian Ocean are also characterized by positive SD anomalies downstream of ET, despite that ET occurs less frequently in this basin. Unlike the other two basins, forecasts initialized 2 d prior to ET onset appear to have two origins of higher forecast SD that merge at the location of ET (Fig. 7a). One of these regions occurs as the TC enters the midlatitudes (i.e., 80°E from 0-48 h), while the other is associated with an upstream trough that moves eastward starting at 50°E at 12 h toward the TC. This result suggests that SIO ET events have an upstream source of uncertainty, which is not as obvious for the other two basins. Following ET, downstream forecasts are characterized by normalized SD anomalies above 0.3, though the signal becomes statistically insignificant 4 d after the onset of ET at roughly 70° downstream of where ET begins.

Forecasts initialized at the onset of transition generally have smaller SD anomalies downstream of where the TC enters the midlatitudes (Fig. 7b). While there is an extensive area of positive SD anomalies between 96-192 h, these are likely a consequence of the small number of ET events that make up this composite.

Baroclinic cyclones in this basin exhibit forecast SD anomalies that are consistent with other basins. In particular, the positive SD anomalies for winter baroclinic cyclones (here May-September) are mainly confined to the upstream trough associated with cyclogenesis (Fig. 7c-d), though there is also a positive SD region that appears between 150°E-180°, which is associated with the next downstream trough within the wave packet. This would appear to suggest that winter baroclinic cyclones in the SIO can lead to a limited reduction in downstream predictability following cyclogenesis, though the amplitude is much smaller than ET. Finally, baroclinic cyclone forecasts during the time period that overlaps with SIO ET (January-March) exhibit positive SD anomalies that maximize with the upstream trough and with the next downstream trough that is
part of the composite wave packet (Fig. 7e-f). This pattern is more consistent with ET systems than winter baroclinic cyclones. In addition, forecasts initialized 2 d prior to \( t = 0 \) are characterized by larger SD anomalies compared to forecasts initialized at \( t = 0 \).

4. ET Forecast Subsetting

The remainder of this paper is devoted to understanding why some ET cases are characterized by statistically higher downstream SD anomalies compared to other cases. This is accomplished by subdividing the forecasts based on various factors for which previous studies have suggested might impact the downstream predictability during ET and determining whether the SD anomalies are greater for one set of cases compared to another. For brevity, the discussion focuses on forecasts initialized 2 d prior to the onset of ET; other initialization times (i.e., -1 d, -3 d, -4 d) have qualitatively similar results (not shown).

A multitude of studies have documented how the relative configuration of the TC and midlatitude flow has a significant impact on TC and downstream flow evolution (e.g., Klein et al. 2002; Harr et al. 2008; Anwender et al. 2008; Torn and Hakim 2009). For example, if the TC moves poleward ahead of a midlatitude trough, it is more likely to reintensify as a baroclinic system, which in turn may lead to a more amplified flow (e.g., Klein et al. 2002; Torn and Hakim 2015; Keller et al. 2014; Quinting and Jones 2016). As a consequence, one potential hypothesis is that cases characterized by greater position variability at the onset of ET, which in turn could be associated with different trough-TC configurations, could be associated with a larger downstream SD anomaly compared to cases where the TC position variability at the onset of ET is lower. This hypothesis is tested by computing the 48-h TC position forecast standard deviation (i.e., at the onset of transition) for each ET case within a basin and compositing the difference in normalized SD anomaly for cases in the upper quartile of TC position variability and for the cases in the lower quartile of TC position.
variability. The statistical significance of the composite differences is evaluated using a bootstrap resampling method whereby two sets of cases equal in size to these two quartiles are randomly selected from all of the ET cases within the same basin and the normalized SD differences are computed. The null hypothesis of no difference between the two subsets of cases is rejected for a particular location and time if the composite SD difference between the large and small TC position cases is outside of the 95% confidence bounds obtained from 2000 random resamples.

For all three basins explored here, the cases characterized by larger TC position variability are not associated with statistically higher downstream SD anomalies (Fig. 8). With the exception of a 48 h period in the Atlantic basin, the SD difference between large and small TC position variability cases is not statistically significant within any of the longitudes and times characterized by positive SD anomalies averaged over all cases. Furthermore, the cases with larger TC position variability are not even statistically different from the cases with smaller TC position variability at the location and time where the TC enters the midlatitudes at the onset of transition (48 h), which might be expected if the TC position variability introduces significant variability on the downstream flow. Overall, this result suggests that having larger TC position variability does not necessarily lead to increased downstream forecast uncertainty.

TC position variability represents a proxy for how uncertainty associated with ET can impact the upper tropospheric flow and hence downstream predictability. An alternative is to compute forecast metrics that more directly measure how the TC interacts with the midlatitudes and stratify cases based on those characteristics. Archambault et al. (2013) demonstrated that ET events characterized by greater advection of PV by the irrotational wind, which is attributable to latent heating near the TC, are associated with a more amplified downstream flow. As a consequence, it is reasonable to hypothesize that cases with greater PV advection by the irrotational wind, which in turn are characterized by more amplified downstream flow, are associated with lower predictability.
This hypothesis is tested by computing the CFSR 200-300 hPa PV advection by the irrotational wind over a 15° box characterized by the largest value as described in Archambault et al. (2013) for each of the ET cases studied here and comparing the downstream SD anomalies for cases in the top quartile (i.e., cases with the largest PV advection by the irrotational wind) against the cases in the bottom quartile (i.e., cases with the smallest PV advection by the irrotational wind). If the above hypothesis is true, than the cases with larger PV advection by the irrotational wind should be characterized by statistically larger downstream SD anomalies.

Although cases with large PV advection by the irrotational wind are associated with higher SD anomalies compared to the small PV advection cases where the TC enters the midlatitudes, there are limited differences downstream. For WNP cases, the large PV advection cases are characterized by statistically larger SD anomalies between 36-108 h (-12 h to +60 h after ET onset) within 10° of where the TC enters the midlatitudes; however, locations and times downstream of that are statistically indistinguishable (Fig. 9a). This distribution suggests that higher PV advection by the irrotational wind introduces forecast uncertainty where the PV advection is occurring (typically where the TC enters the midlatitudes); however, the higher forecast SD in this region does not necessarily translate into larger downstream uncertainty than cases with lower PV advection by the irrotational wind. Instead, the region of statistically higher SD remains tied to the west side of the amplifying ridge (not shown). ET forecasts in the Atlantic basin exhibit similar behavior, with the largest SD differences associated with where the TC enters the midlatitudes (Fig. 9b). Finally, SIO cases with larger PV advection are not characterized by statistically higher SD anomalies where the TC enters the midlatitudes (Fig. 9c). The only region of statistically larger SD anomalies for large PV advection cases is on the eastern side of the ET SD plume, which may suggest that cases with larger PV advection are associated with higher SD further downstream; however, the differences may also be due to the small number of cases (15) in each quartile. Nevertheless,
these results suggest that cases characterized by larger upper-tropospheric PV advection by the irrotational wind are not necessarily associated with systematically higher downstream SD.

While it appears that cases with higher PV advection by the irrotational wind are not necessarily associated with higher downstream SD, another possibility is that cases characterized by larger uncertainty in the magnitude of the PV advection by the irrotational wind could be associated with higher downstream SD. This hypothesis is evaluated by computing the forecast SD of the PV advection by the irrotational wind for the same case-specific geographic area and time period used above and taking the difference between the composite of cases with the upper quartile of PV advection standard deviation and the composite of cases within the lower quartile of PV advection standard deviation.

Similar to the analysis PV advection metric, the upper quartile cases are not associated with greater downstream 500 hPa height SD. For the WP and SIO basins (Fig. 10a,c), the normalized SD difference is mainly confined to where the TC enters the midlatitudes, while there are no statistically significant differences in the Atlantic basin (Fig. 10b). Overall, this result suggests that cases characterized by large uncertainty in how the irrotational wind interacts with the midlatitude PV gradient do not necessarily lead to greater downstream forecast uncertainty.

The lack of significant differences suggests that downstream predictability during ET events are not primary dictated by properties of the TC and that cases with larger downstream forecast SD may instead be a function of the midlatitude flow configuration. This possibility is evaluated by computing a metric that measures the downstream SD anomaly (hereafter downstream SD metric) for each case and comparing the midlatitude flow properties for cases with large downstream SD metric with cases with small downstream SD metric. The downstream SD metric is computed by averaging the normalized SD within a basin-specific polygon (shown in Figs. 2c, Figs. 6a, Figs. 7a, for the WNP, Atlantic, and SIO basins, respectively), which generally encloses the plume of pos-
itive SD anomalies over all cases. Fig. 11a shows a histogram of the downstream SD metric for each basin. All three basins are characterized by a variety of metric values, including a significant number below zero, meaning that those cases are characterized by downstream SD values that are similar to or lower than a typical forecast for that location and time. In general, the WNP has more cases with downstream SD values above 0.6, which is indicative of a large reduction in downstream predictability. Similar to above, composites of the cases in the upper quartile of the downstream SD metric cases are compared to composites of the cases in the lower quartile of downstream SD metric cases. Fig. 11b shows the difference in normalized SD between the cases in the upper and lower quartile of the downstream SD metric for WNP cases. As desired, the region of statistically significant difference is confined to SD plume; therefore, this metric can discriminate between cases with large downstream SD versus cases that do not.

The relatively small number of cases and variety of midlatitude flow configurations associated with ET preclude using the traditional Hovmöller approach, which involves computing average quantities between often wide latitude bounds. Instead, the Martius et al. (2006) approach is adopted, which provides a method of determining properties along the midlatitude waveguide, whose position can vary as a function of longitude and time. Here, the waveguide position at each lead analysis time and longitude is determined by tracing the 2 PVU contour on the CFSR 340 K surface and applying a 7.5° longitude filter. Once the waveguide longitude is determined, modified Hovmöller diagrams are constructed by averaging CFSR quantities within ±10° of latitude of the waveguide position at each longitude and time. The benefit of this approach is that it can demonstrate differences within the time-evolving structure of the midlatitude waveguide without having to resort to wide latitude bounds. Indeed, the traditional 30° latitude Hovmöller generally yields statistically insignificant differences (not shown). The statistical significance of the differ-

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3This contour generally denoted the position of the midlatitude waveguide during a majority of cases.
ence between upper and lower quartile cases is established using the aforementioned bootstrap
resampling method.

Although the analysis differences between the large and small downstream SD metric cases
vary between the three basins, some common characteristics exist, particularly with respect to
the latitude of the waveguide. Fig. 12 provides a modified Hovmoller diagram of the waveguide-
relative standardized differences between the large and small downstream SD metric cases. In
the WNP, the large downstream SD metric cases are characterized by a more poleward waveguide
latitude from 150°E-180° starting at 0 h, which corresponds to the position of the downstream
ridge (Fig. 12a) and subsequently follows the plume of higher spread (c.f., Fig. 2c). This result
suggests that ET events that cases with larger downstream SD are characterized by a more am-
plified downstream anticyclone starting at the onset of ET. The stronger downstream anticyclone
is preceded by lower 250 hPa geopotential heights along the waveguide (i.e., a trough) between
120-145°E (i.e., upstream of where the TC enters the midlatitudes) up to 12 h prior to the onset
of ET (Fig. 12b). In addition, there are anomalously higher heights downstream of the composite
ridge between 160°E-180 within 12 h of ET onset. The combination of lower heights to the west,
higher heights to the east is associated with anomalous poleward meridional wind that would help
amplify the downstream ridge through the rotational wind, thereby leading to a stronger waveguide
perturbation and hence Rossby wave dispersion. This result is consistent with Quinting and Jones
(2016), who found that ET events associated with Rossby wave development were associated with
greater downstream SD than cases that did not.

Waveguide-relative differences between the large and small SD metric cases in the Atlantic and
SIO basins exhibit some similarity to the WNP. Namely, the waveguide latitude for the large SD
cases is statistically poleward in the downstream ridge starting either at or shortly after the onset
of ET, which indicates stronger ridge amplification and hence waveguide perturbations in these
cases (Fig. 12c,d). Unlike the WNP, there is little evidence of the upstream trough as both basins are characterized by statistically insignificant upstream 250 hPa height differences (not shown).

5. Summary and Conclusions

This paper evaluates the hypothesis that TCs entering the midlatitudes and baroclinic cyclogenesis, both of which can act as perturbations on the midlatitude waveguide and introduce downstream Rossby wave dispersion, are generally associated with decreased downstream predictability. This hypothesis is evaluated using 30 y of cases taken from the GEFS Reforecast Dataset Version 2 and comparing the 500 hPa geopotential height ensemble SD for cases with one of these events against the forecast “climatology” for that particular location, lead time, and time of the year. For each type of event, forecast SD are composited for forecasts initialized at various lead times prior to, during, and after either the onset of ET, or based on when the baroclinic cyclogenesis point. The null hypothesis that forecasts of these events are statistically indistinguishable from a typical forecast is rejected if the composite forecast SD is statistically different from a random sample of forecasts for the same time of year and geographical location.

ET forecasts exhibit the previously-documented plume of anomalously high SD that originates from where the TC enters the midlatitudes and moves downstream at the group velocity of the associated wave packet. In general, the region of anomalously high SD occurs after the passage of the wave packet peak at a given location, suggesting that the packet itself is not characterized by greater uncertainty. In the WNP, there is no evidence of a region of higher SD that originates upstream of where the TC enters the midlatitudes, suggesting that, on average, the downstream SD plume is related to how the TC interacts with the midlatitude flow and is not due to the amplification of a pre-existing midlatitude forecast uncertainty. ET forecasts in the WNP basin appear to introduce the highest amplitude and longest-lasting downstream SD anomaly of the three basins.
studied here, such that the positive SD anomaly reaches the east coast of North America (150° of longitude) 4 d after the onset of ET, while Atlantic and SIO SD anomalies do not extend more than 80° downstream of ET. In general, forecasts initialized close to the onset of ET (-1, 0, +1 d) have smaller downstream SD anomalies relative to forecasts initialized 2, 3, or 4 d prior to the onset of ET, suggesting that forecasts have the greatest uncertainty before the TC interacts with the midlatitude flow. This may suggest that much of the downstream uncertainty results from the initiation of the ET-related wave packet. Once the wave packet structure is contained in the initial conditions, the downstream forecast state has similar predictability to a typical forecast for that location and time of year. This result would be consistent with Grazzini and Vitart (2015) who showed that Rossby wave packets were associated with higher than average medium range forecast skill.

Subdividing the ET cases based on characteristics associated with the TC, the midlatitude waveguide, and how these two features interact, suggest that less predictable ET cases are associated with certain midlatitude flow configurations. In all three basins, the cases with the largest downstream SD are characterized by a more amplified downstream anticyclone compared to the cases with the lowest downstream SD. Previous work has suggested that downstream Rossby waves originate with the downstream anticyclone (e.g., Torn and Hakim 2015); therefore, this result suggests that rapid anticyclogenesis may act to limit downstream predictability. In addition, the large downstream SD cases in the WNP are characterized by an upstream trough and downstream ridge at the onset of ET, which in combination with diabatic outflow, would help amplify the downstream ridge. By contrast, ET cases with large TC position variability at the onset of ET, large PV advection by the irrotational wind, or large uncertainty in the magnitude of the PV advection by the irrotational wind, do not exhibit greater downstream SD compared to the ET cases with small values. As a consequence, this result suggests that while large PV advection by the irrotational wind may produce a larger downstream response, that response is not necessarily less
predictable. It is also worth noting that not all ET cases are characterized by decreased downstream predictability. Roughly 25% of ET cases were characterized by normalized SD metric value below 0.0, meaning that the downstream SD is no different for those cases than one might expect if no ET was present.

In contrast to ET forecasts, baroclinic cyclogenesis has a much more limited impact on downstream predictability, even during the same time of the year as ET. Winter baroclinic cyclones are characterized by positive SD anomalies that originate upstream in the wave packet associated with cyclogenesis, with positive SD anomalies coinciding with and mostly limited to the upstream trough. One of the most interesting results of this study is that forecasts over the eastern North Pacific Basin are characterized by lower SD relative to climatology for several days following WNP baroclinic cyclogenesis. This result is likely due to the nature of the Pacific waveguide during cyclogenesis events, which is often characterized by wavebreaking in the central Pacific. This wavebreaking limits the motion of wave packets and hence the ability of forecast uncertainty to propagate downstream. In all three basins, baroclinic cyclones that occur at the same time of year as ET, have greater downstream SD anomalies than the winter, suggesting that baroclinic cyclogenesis during the non-peak seasons may be more likely to reduce downstream predictability than during the winter, which is often characterized by greater baroclinicity. It is unclear why this is the case, though it could be related to the structure of the midlatitude jet during this season. Future work will attempt to understand why midlatitude cyclones, which produce waveguide perturbations and wave packets akin to ET, do not produce a significant downstream reduction in predictability and why some cases are characterized by significantly higher downstream SD compared to other cases.
Acknowledgments. The paper benefited from discussions with Heather Archambault, Chris Davis, Pat Harr. GEFS Reforecast data was made available through the Physical Science Division of the NOAA Earth System Research Lab. This research is supported by NSF Award 1461753.

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<th>Number of Cases</th>
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<td>WNP Fall Baroclinic</td>
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<tr>
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