The Impact of Initial Condition and Warm Conveyor Belt Forecast Uncertainty on Variability in the Downstream Waveguide in an ECWMF Case Study

JEREMY D. BERMAN AND RYAN D. TORN

Department of Atmospheric and Environmental Sciences, University at Albany, State University of New York, Albany, New York

(Manuscript received 17 September 2018, in final form 9 August 2019)

Abstract

Perturbations to the potential vorticity (PV) waveguide, which can result from latent heat release within the warm conveyor belt (WCB) of midlatitude cyclones, can lead to the downstream radiation of Rossby waves, and in turn high-impact weather events. Previous studies have hypothesized that forecast uncertainty associated with diabatic heating in WCBs can result in large downstream forecast variability; however, these studies have not established a direct connection between the two. This study evaluates the potential impact of latent heating variability in the WCB on subsequent downstream forecasts by applying the ensemble-based sensitivity method to European Centre for Medium-Range Weather Forecasts (ECMWF) ensemble forecasts of a cyclogenesis event over the North Atlantic. For this case, ensemble members with a more amplified ridge are associated with greater negative PV advection by the irrotational wind, which is associated with stronger lower-tropospheric southerly moisture transport east of the upstream cyclone in the WCB. This transport is sensitive to the pressure trough to the south of the cyclone along the cold front, which in turn is modulated by earlier differences in the motion of the air masses on either side of the front. The position of the cold air behind the front is modulated by upstream tropopause-based PV anomalies, such that a deeper pressure trough is associated with a more progressive flow pattern, originating from Rossby wave breaking over the North Pacific. Overall, these results suggest that more accurate forecasts of upstream PV anomalies and WCBs may reduce forecast uncertainty in the downstream waveguide.

1. Introduction

Synoptic-scale Rossby waves are ubiquitous features of the extratropical atmospheric circulation that are associated with surface cyclones and anticyclones (e.g., Hoskins et al. 1985), and act as precursors to high-impact weather events (e.g., Chaboureau and Claud 2006; Martius et al. 2006). Rossby waves propagate along horizontal gradients of potential vorticity (PV) that act as waveguides in the sub tropics and extratropics, and are often collocated with the jet (Hoskins and Ambrizzi 1993; Schwierz et al. 2004; Martius et al. 2010). Perturbations of the waveguide can induce the downstream dispersion of Rossby waves (i.e., downstream development; Simmons and Hoskins 1979; Chang 1993) in the form of wave packets (e.g., Simmons and Hoskins 1979; Hakim 2003; Riemer et al. 2008), which in turn can amplify the downstream waveguide (e.g., Chang and Orlanski 1993).

One of the most prominent ways to perturb the midlatitude waveguide is via the divergent outflow associated with latent heat release in the warm conveyor belt (WCB) of extratropical cyclones (e.g., Wernli and Davies 1997; Riemer et al. 2008). The WCB (e.g., Browning 1971; Harrold 1973) is one of three prominent Lagrangian airstreams from the conveyor belt model of extratropical cyclones (Carlson 1980). The WCB transports relatively warm and moist air poleward and upward, and is the primary cloud- and precipitation-producing flow structure associated with extratropical cyclones (Browning 1986, 1990). The rising motion in WCBs is associated with the amplification of an upper-tropospheric ridge downstream of the surface cyclone (e.g., Wernli 1997; Joos and Wernli 2012; Chagnon et al. 2013). The amplification of this ridge can occur in response to several dynamical mechanisms: 1) downstream development when an upstream trough is more amplified than a downstream

Corresponding author: Jeremy Berman, jdberman@albany.edu

DOI: 10.1175/MWR-D-18-0333.1

© 2019 American Meteorological Society. For information regarding reuse of this content and general copyright information, consult the AMS Copyright Policy (www.ametsoc.org/PUBSReuseLicenses).
trough, 2) the adiabatic transport of lower PV air from the lower to the upper troposphere or the poleward transport of lower PV air, 3) the diabatic reduction of PV above the level of maximum diabatic heating associated with latent heat release (e.g., Riemer and Jones 2010), and 4) the advection of lower PV air by the irrotational outflow, which can amplify the ridge and tighten the PV gradient (e.g., Davis et al. 1993; Henderson et al. 2010; Archambault et al. 2013). Recent case studies suggest that ridge amplitude is more strongly modulated by upper-tropospheric divergent outflow than by direct diabatic PV modification (e.g., Teubler and Riemer 2016). In addition, Baumgart et al. (2018) highlighted the relative importance of individual processes on PV forecast error growth by applying the PV tendency partitioning technique of Teubler and Riemer (2016) to ECMWF forecasts. For this case, they found that PV error growth has the largest contribution from tropopause-based PV anomalies in contrast to the smaller contribution from upper-tropospheric divergent outflow.

Several studies have suggested that forecast errors associated with the diabatic heating in WCBs can amplify and propagate downstream along the waveguide (e.g., Bosart and Lackmann 1995; Davies and Didone 2013), which in turn lead to the reduction of practical predictability (Lorenz 1969) downstream. Rodwell et al. (2013) and Rodwell et al. (2018) showed that poor medium-range forecasts over Europe originated with errors in the diabatic heating associated with mesoscale convective systems over North America. Moreover, Lamberson et al. (2016) highlighted how errors in subsynoptic features originating from the WCB in the eastern Pacific lead to European Centre for Medium-Range Weather Forecasts (ECMWF) downstream cyclone forecast variability over Europe.

In addition to the above error growth examples, errors associated with diabatic heating in WCBs can also impact the divergent outflow, which in turn can amplify the nearby waveguide. For instance, forecast uncertainty associated with the divergent outflow within a WCB-like region of Hurricane Sandy (2012) yielded differences in the structure of the upper-level ridge on the poleward side of the TC, which in turn impacted subsequent position forecasts (Torn et al. 2015). The diabatic heating and relative humidity errors were hypothesized to alter the vertical profiles of divergence and vertical velocity in the WCB region, which then modulated the structure of the divergent outflow that was associated with the ridge amplification. The aforementioned studies suggest the importance of the WCB to downstream forecasts; however, there is not necessarily an established clear connection between the two in the scientific literature.

One method to link the WCB to downstream forecasts is to use ensemble-based sensitivity analysis (Ancell and Hakim 2007; Torn and Hakim 2008), which has successfully identified sources of forecast variability for features such as midlatitude cyclones (e.g., Hakim and Torn 2008; Chang et al. 2013) and mesoscale convection (e.g., Bednarczyk and Ancell 2015; Torn and Romine 2015; Berman et al. 2017). The ensemble-based sensitivity technique applies linear statistics about a set of nonlinear forecast trajectories, which can make it useful for quantifying the impact of nonlinear processes, such as WCB-related diabatic heating, on downstream forecast variability. As a consequence, this study explicitly evaluates how uncertainty associated with the WCB impacts downstream variability by employing ensemble model forecasts and ensemble-based sensitivity analysis for a case study characterized by cyclogenesis and subsequent amplification of the downstream flow.

The remainder of this paper proceeds as follows. Section 2 describes the dataset and ensemble sensitivity technique while section 3 provides the synoptic overview of the case. Section 4 details the forecast uncertainty in downstream ridge amplification and the method used to isolate ridge amplitude variability. Section 5 diagnoses potential WCB sources of forecast uncertainty associated with later forecasted ridge amplification, while section 6 explores how upstream features contribute to WCB forecast uncertainty. A summary and conclusions are given in section 7.

2. Data and methodology

a. Case selection

The hypothesis that forecast variability in the downstream waveguide is sensitive to upstream WCB uncertainty is evaluated through a case study of a highly amplified ridge over the central North Atlantic at 1200 UTC 14 November 2013, which is the same period examined by Baumgart et al. (2018). This period is characterized by cyclogenesis over eastern North America and subsequent ridge amplification over the central North Atlantic, which later led to wave breaking over western Europe and heavy rainfall over Sardinia associated with a cutoff cyclone. ECMWF ensemble forecasts initialized 4–6 days prior to the ridge amplification exhibited significant errors in the amplitude and position of the ridge, which serves as motivation for investigating the sensitivity of ridge development to uncertainty in the upstream WCB at earlier times.

b. Dataset

This study uses ensemble forecasts from the 51-member ECMWF model, extracted from The Observing System
Research and Predictability Experiment Interactive Grand Global Ensemble (TIGGE; Bougeault et al. 2010). In addition, ECMWF deterministic analyses were also extracted from TIGGE. The TIGGE dataset is characterized by discrete vertical levels of 1000, 925, 850, 700, 500, 300, 250, and 200 hPa, with a horizontal resolution of 1° × 1°. The main focus of this study is on forecasts initialized at 1200 UTC 8 November 2013, which is six days prior to the North Atlantic ridge reaching maximum amplitude. In addition, a similar analysis was carried out for forecasts initialized 24 h (1200 UTC 9 November) and 48 h (1200 UTC 10 November) later.

c. Ensemble-based sensitivity analysis

The role of forecast uncertainty at earlier lead times on subsequent ridge amplitude forecasts is evaluated by using the ensemble-based sensitivity technique. For an ensemble of size $M$, the sensitivity of a forecast metric $J$ to a model state variable $x_i$ at some earlier lead time is computed via

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

where $J$ and $x_i$ are $1 \times M$ ensemble estimates of the forecast metric and $i$th state variable, respectively, $\text{cov}$ denotes the covariance, and $\text{var}$ denotes the variance. For ease of comparing various fields, the values of $x_i$ are normalized by its ensemble standard deviation prior to computing sensitivity; therefore all sensitivities have units of the forecast metric per standard deviation of the state variable (here dimensionless). Sensitivity values are considered statistically significant if the absolute value of $\partial J/\partial x_i$ is greater than the 95% confidence interval using a z-score test (e.g., Torn and Hakim 2008).

3. Case overview

The days preceding ridge development are characterized by a series of synoptic interactions that originate from cyclogenesis and Rossby wave breaking over the North Pacific Ocean. At 1200 UTC 9 November 2013 (5 day prior to the maximum ridge amplitude) the 250-hPa PV$^1$ field over the Pacific is characterized by a series of troughs and ridges, such as a ridge over Japan (hereafter “R1”), trough at 160°E (hereafter “T1”), ridge at 160°W (hereafter “R2”), and trough at 145°W (hereafter “T2”; Fig. 1a). One day later (Fig. 1b), R1 amplifies in association with surface cyclogenesis near 150°E (hereafter “C1”). Downstream of this cyclone is a region of pronounced 250-hPa divergent flow, which would be expected to amplify R1 through negative PV advection.

At 1200 UTC 11 November, R1 extends from 160°E to 160°W, with T1 extending southward along 160°W (Fig. 1c). Moreover, R2 undergoes anticyclonic wave breaking over western Canada, with a surface anticyclone (hereafter “AC”) forming near 50°N, 100°W. One day later (12 November), R2 moves eastward to 90°W while a positively tilted PV trough (hereafter “T3”) forms near 40°N, 100°W (Fig. 1d). Meanwhile AC undergoes anticyclogenesis to 1048 hPa and moves southward toward 40°N, 100°W. Farther downstream, cyclogenesis occurs at 50°N, 70°W near New Brunswick, Canada (hereafter “C3”).

At 1200 UTC 13 November, C3 deepens to 986 hPa with increasing upper-level divergent outflow and ridge amplification (hereafter “R3”) immediately to its east (Fig. 1e). Absent other factors, the cross-PV contour component of the divergent wind supports R3’s amplification. Notably, C3 reached its strongest intensity 12 h later at a minimum sea level pressure of 968 hPa (not shown). Finally at 1200 UTC 14 November, AC and T3 move eastward while C3 has moved northward to 65°N, 40°W near Greenland (Fig. 1f). More importantly, R3 reaches its maximum amplitude, with broad upper-level divergent outflow along its western side between 20°–65°N and 50°–10°W. Furthermore, a trough (hereafter “T4”) forms farther downstream over western Europe along the Prime Meridian. Given that the amplification of R3 occurs concurrently with the divergent outflow associated with C3, it is possible that the structure of the ridge could be sensitive to the divergent outflow associated with C3. Therefore, the remainder of this paper focuses on the sensitivity of the ridge structure to the divergent outflow and the upstream features that modulate the divergent outflow.

4. Waveguide forecast uncertainty

The central North Atlantic ridge (R3) at 1200 UTC 14 November was characterized by significant forecast uncertainty within ECMWF ensemble forecasts at 1–6 day lead times (Fig. 2a). As the lead time increases, the ensemble-mean ridge, denoted by the 250-hPa 2 PVU (1 PVU = 10$^{-6}$ K kg$^{-1}$ m$^2$ s$^{-1}$) contour, is characterized by a more underestimated amplitude relative to the analysis. However, these ensemble-mean amplitude differences may be due to variability in ridge position within the ensemble, which is smoothed out in the

---

$^1$To reduce small-scale noise, the PV field, for this figure and all subsequent PV figures, represents the 300-km area-averaged values.
ensemble-mean PV field. The uncertainty in the ridge amplitude and position is further elucidated by looking at individual ensemble forecasts, for forecasts initialized at 1200 UTC 8 November (144-h forecast; Fig. 2b). At this lead time, there is clearly both position and amplitude variability in the forecasted North Atlantic ridge. As a consequence, the underestimated ensemble-mean ridge amplitude in Fig. 2a could be due to ridge position or amplitude differences between members.

Given that the focus of this study is on ridge amplitude variability, it is necessary to identify a forecast metric that isolates this particular metric. Area-average metrics (e.g., 250-hPa PV) at the ridge apex are not necessarily appropriate because these metrics can take on different values due to ridge amplitude and/or ridge position. One way to separate ridge position and amplitude variability within the ensemble forecasts is to apply empirical orthogonal functions (EOF) analysis to the ensemble 250-hPa PV field over a region enclosing the ridge, and use the principal components (PCs) of the EOF pattern that most closely relate to the ridge amplitude as the forecast metric (e.g., Chang et al. 2013). Here EOF analysis is applied to the 144-h ensemble forecasts of the 250-hPa PV field over a domain enclosing the ensemble-mean ridge (domain shown in Fig. 3). Figure 3a shows the leading EOF (hereafter EOF1), which explains 30.2% of the variance within this area. This EOF pattern is characterized by a west–east negative–positive dipole centered on the ridge, which indicates that the dominant mode of variability in the PV field is the west–east position of the ridge, such that positive (negative) PCs denote a more westward (eastward) ridge (hereafter PC1). By contrast, the second EOF (hereafter EOF2; Fig. 3b) explains 16.0% of the variance and indicates a region of negative values collocated with the ridge apex and positive values with the upstream and downstream trough. As a consequence, this mode of variability corresponds with the ridge amplitude, such that positive (negative) PCs denote a more (less) amplified waveguide, including the ridge (hereafter PC2).

**FIG. 1.** ECMWF deterministic analysis of mean sea level pressure (contours, hPa), 250-hPa irrotational wind (vectors, m s$^{-1}$; ≥5 m s$^{-1}$ only), and 250-hPa potential vorticity (shading, PVU) for (a) 1200 UTC 9 Nov, (b) 1200 UTC 10 Nov, (c) 1200 UTC 11 Nov, (d) 1200 UTC 12 Nov, (e) 1200 UTC 13 Nov, and (f) 1200 UTC 14 Nov 2013.  

---

The EOF patterns shown in this study are computed by subtracting grid points of the field for each ensemble member from the ensemble mean and normalizing by the standard deviation, and then regressing the various PCs onto the respective field, resulting in EOF patterns with units of PVU.
5. Diagnosis of ridge amplitude variability

The source of ridge amplitude variability is evaluated by employing the ensemble-based sensitivity technique with the primary forecast metric \( J \) defined as PC2 (hereafter \( J_{\text{RIDGE-PC2}} \)), where positive values denote a higher-amplitude ridge and negative values denote a lower-amplitude ridge. It is noted to the reader that the sensitivity magnitude (e.g., 10 PC units per standard deviation of 250-hPa PV) for this metric is only meaningful in conveying the degree to which the ridge is amplified, such that larger (smaller) sensitivity magnitudes indicate larger (smaller) ridge amplitudes. In addition, a similar analysis has been carried out for the case when \( J \) is PC1 (i.e., ridge position). The results indicate that the ridge position is mainly sensitive to the position of upstream features at earlier lead times (not shown). In addition, the sensitive regions for ridge position variability are independent of the sensitive regions for ridge amplitude variability that are subsequently shown.

The processes associated with the uncertainty in the 144-h ridge amplitude are determined by computing the sensitivity of \( J_{\text{RIDGE-PC2}} \) to the 250-hPa PV field at earlier lead times (Fig. 4). At 144 h, \( J_{\text{RIDGE-PC2}} \) exhibits negative sensitivity to the ensemble-mean ridge (R3) over the central North Atlantic and a dipole of negative–positive sensitivity on the western–eastern sides of the upstream longwave trough (T3) over the western North Atlantic (Fig. 4a). The negative sensitivity to the ridge apex is consistent with the definition of \( J_{\text{RIDGE-PC2}} \) since the negative sensitivity indicates that the members with lower PV in the ridge have a more amplified ridge at 144 h. On the other hand, the negative–positive sensitivity dipole associated with the trough suggests that the members with a more amplified ridge are characterized by an upstream trough shifted farther east. Moving backward 12 h, the aforementioned negative sensitivity remains collocated with the ridge apex and the negative–positive sensitivity dipole straddles the trough (Fig. 4b).

![Fig. 2. The 250-hPa 2 PVU contour valid at 1200 UTC 14 Nov 2013 for (a) the ECMWF ensemble-mean forecast at different lead times and (b) the day 6 forecast of the ECMWF 51-member ensemble (blue lines) and ensemble mean (red line) compared to the analysis (black line).](image)

![Fig. 3. Regressed 144-h 250-hPa potential vorticity (shading, PVU) from the (a) first principal component of the 144-h 250-hPa potential vorticity field initialized at 1200 UTC 8 Nov 2013. (b) As in (a), but for the second principal component. The black lines denote the ensemble-mean 250-hPa potential vorticity (contours, PVU).](image)
Moreover, these coherent sensitivity patterns track the movement of the PV features at 120 h (Fig. 4c) and 108 h (Fig. 4d), while prior to 108 h the sensitive regions are small and insignificant (not shown), which could be related to sampling issues or the flow evolution. Overall, these results emphasize that the 144-h ridge amplitude is sensitive to the structure of the ridge and upstream trough at earlier times.

The sensitivity of $J_{\text{RIDGE-PC2}}$ to the 250-hPa PV becomes largest after 108 h during the intensification of C3 (cf. Fig. 1), which suggests that downstream ridge amplitude could be sensitive to cyclone-related processes. One process by which a cyclone can modify the upper-level PV distribution is via the divergent outflow associated with latent heating, which can amplify a ridge through advection of low PV air by the irrotational wind (e.g., Archambault et al. 2013). Given that the cyclone’s divergent outflow is perpendicular to the PV gradient near the ridge (cf. Figs. 1c,f), it is possible that differences in the advection of PV by the irrotational wind could lead to subsequent differences in the ridge amplitude. This hypothesis is evaluated by computing the sensitivity of $J_{\text{RIDGE-PC2}}$ to the advection of 250-hPa PV by the irrotational wind at earlier times (Fig. 5).

At 144 h, the surface cyclone located at 60°N, 52°W is associated with a region of negative sensitivity immediately to its northeast coincident with the area of ensemble-mean negative PV advection by the irrotational wind (Fig. 5a). This negative sensitivity, which can be traced backward in time to 132 h (Fig. 5b), 120 h (Fig. 5c), and 108 h (Fig. 5d), suggests that the members with largest ridge amplification are associated with greater negative PV advection by the irrotational wind immediately downstream of the cyclone during the previous 36 h. Moreover, the sensitive region becomes statistically significant when the ensemble-mean negative PV advection first appears at 108 h, which could explain why the sensitivity cannot be traced backward prior to 108 h (not shown); however, it is also possible that the lack of sensitivity is due to sampling error from the limited ensemble size. Despite this, the connection between the PV advection over the sensitive region and the amplitude of the ridge is further illustrated by computing a correlation between $J_{\text{RIDGE-PC2}}$ and the 120–144-h PV advection area averaged over the region inside the red contour shown in Figs. 5a–c, which encloses the region of largest sensitivity. The correlation coefficient between this PV advection metric and $J_{\text{RIDGE-PC2}}$ is 0.59, which is a relatively large correlation and thus further supports the connection between these two features. In addition, the sensitivity of $J_{\text{RIDGE-PC2}}$ to the advection of 250-hPa PV by the rotational wind is smaller (not shown), which suggests that the PV advection by the irrotational wind plays a larger role in modulating subsequent ridge amplification.

FIG. 4. Sensitivity of $J_{\text{RIDGE-PC2}}$ to the (a) 144-, (b) 132-, (c) 120-, and (d) 108-h 250-hPa potential vorticity (shading, PC units) initialized at 1200 UTC 8 Nov 2013. The black lines denote the ensemble-mean 250-hPa potential vorticity (contours, PVU). The stippling indicates where the sensitivity is statistically significant at the 95% confidence level. The ‘‘L’’ indicates the location of the ensemble-mean forecasted cyclone C3.
One potential reason for the lack of sensitivity to the 250-hPa PV advection by the irrotational wind near the cyclone prior to 108 h is that the uncertainty in the PV advection by the irrotational wind originates from variability in latent heat release at earlier times. Given that the magnitude of the divergent outflow of a midlatitude cyclone is often proportional to the magnitude of the latent heating within the WCB (e.g., Wernli 1997; Joos and Wernli 2012; Chagnon et al. 2013), it is possible that there may be sensitivity to the distribution of latent heating in the WCB. Unfortunately, the TIGGE dataset does not contain diabatic heating rates, so a proxy variable must be used. From the water continuity equation, latent heat release is proportional to the integrated horizontal water vapor transport (hereafter IVT) convergence. In this manner, differences in latent heat release may be associated with differences in IVT convergence. Figure 6 shows the sensitivity of $J_{\text{RIDGE-PC2}}$ to the 850–1000-hPa IVT convergence at various lead times. At 132 h (Fig. 6a) there is positive sensitivity concentrated to the east of the ensemble-mean cyclone within a region of ensemble-mean IVT convergence to the north of southerly IVT, which suggests that greater IVT convergence there, and hence greater latent heat release, would be associated with a larger ridge amplitude later on. This positive sensitivity region occurs within an area that meets the WCB criteria from Madonna et al. (2014), which is determined from trajectories computed from the ERA-Interim reanalysis dataset (Sprenger et al. 2017). 12 h earlier (Fig. 6b), the positive sensitivity region can be traced east of the cyclone’s center ahead of a region of broad ensemble-mean southerly wind to the southeast of the cyclone along 50°W. This sensitive region reduces in amplitude and rotates anticyclonically around the cyclone at 108 h (Fig. 6c) and 96 h (Fig. 6d), suggesting that members with greater ridge amplitude are associated with greater lower-tropospheric IVT convergence within the WCB to the east of the cyclone earlier on. Given that IVT convergence includes both moisture content and wind velocity, the sensitivity to each can be elucidated by computing the sensitivity of $J_{\text{RIDGE-PC2}}$ to the 120-h 850–1000-hPa meridional wind and specific humidity independently (not shown). It is found that $J_{\text{RIDGE-PC2}}$ exhibits larger sensitivity to meridional wind (10 PC units per standard deviation) than to specific humidity (8 PC units per standard deviation) within the sensitive region to the east of the cyclone, which suggests that the ridge amplitude exhibits greater sensitivity to the wind compared to the specific humidity. Other lead times are characterized by similar results (not shown).

Given the sensitivity of the ridge amplitude to the horizontal structure of the upstream WCB, it is desirable...
to test if the ridge amplitude is also sensitive to the vertical structure of the WCB. This hypothesis is evaluated by computing the sensitivity of $J_{\text{RIDGE-PC2}}$ to the 108-h vertical profiles of water vapor, meridional wind, and temperature area averaged within the black box in Fig. 6c (Fig. 7). This region encompasses the northern edge of the WCB and is colocated with large IVT convergence sensitivity shown in Fig. 6c. Within this region, the members are characterized by positive sensitivity of up to 7.5 PC units per standard deviation increase in moisture between the surface and 700 hPa, which suggests that members with large ridge amplitude are associated with greater moisture content in that layer (Fig. 7a). In addition, there is larger positive sensitivity to the meridional wind (9 PC units) from 850 hPa to the surface, particularly near 850 hPa (Fig. 7b), and positive sensitivity to temperature (up to 7.5 PC units) below 500 hPa (Fig. 7c). These sensitivity values suggest that stronger southerly winds from 850 hPa to the surface and higher temperatures below 500 hPa are associated with a more amplified ridge. Overall, the vertical profiles of sensitivity suggest that members with a large ridge amplitude are characterized by a more intense WCB, meaning stronger southerly winds, higher moisture content, and higher temperatures within the lower troposphere.

The higher lower-tropospheric moisture, temperature, and stronger southerly winds within the WCB could be due to those members having a stronger cyclone, which might be expected to have stronger geostrophic winds and therefore larger transport of heat and moisture in the WCB (e.g., Binder et al. 2016). This hypothesis of earlier variability in cyclone intensity is evaluated by computing the sensitivity of $J_{\text{RIDGE-PC2}}$ to MSLP at various lead times (Fig. 8). At 132 h (Fig. 8a) the largest amplitude sensitivity to the MSLP is located in the ensemble-mean cyclone center and to its south characterized by negative sensitivity. This result implies that the members with greater ridge amplitude are associated with lower pressure in the vicinity of the ensemble-mean cyclone at 132 h.
It is worth noting that there is little correlation between the bulk properties of the cyclone and subsequent ridge amplitude. For example, the correlation between each member’s 132-h cyclone minimum MSLP and \( J_{\text{RIDGE-PC2}} \) is 0.0001, while the correlation between each member’s 132-h cyclone position and \( J_{\text{RIDGE-PC2}} \) is 0.15, both of which are statistically insignificant for the 51-member sample size. Therefore, the aforementioned negative sensitivity result does not reflect the position and/or intensity of the cyclone alone. This negative sensitivity region can be traced backward in time to 120 h (Fig. 8b) shifting west of the cyclone center. However, by 108 h (Fig. 8c) the negative sensitivity region is much reduced in magnitude and concentrates farther west and south of the cyclone center, suggesting a position shift of the cyclone center to the southwest and/or a decrease in pressure along the cold front to the south of the cyclone. The region south of the cyclone center is coincident with the cold front position, which combined with the negative sensitivity signature suggests that members with greater ridge amplitude are associated with lower pressure near the ensemble-mean cyclone only at 120 h and later. Instead, \( J_{\text{RIDGE-PC2}} \) is sensitive to the pressure field along the cold front pressure trough. Moreover, the observed intensity of the cyclone at the aforementioned lead times falls within the forecast ensemble envelope of cyclone intensity (Table 1), which supports the notion that the uncertainty in baroclinic development of the cyclone had little connection with subsequent ridge amplitude forecast variability.

Later initialization times exhibit similar sensitivity of the amplitude of R3 to the upstream WCB. For example, Fig. 9 illustrates the sensitivity of the central North Atlantic ridge amplitude\(^3\) valid at 1200 UTC 14 November to earlier forecasts of 250-hPa PV advection by the irrotational wind, 850–1000-hPa IVT convergence, and MSLP initialized 24 h later (1200 UTC 9 November; left column) and 48 h later (1200 UTC 10 November; right column). Both initializations are characterized by negative sensitivity to PV advection by the irrotational wind to the northeast of the cyclone (Figs. 9a,b), which is the same location of largest negative sensitivity for the forecast initialized at 1200 UTC 8 November (cf. Fig. 5b). Moreover, both initializations indicate positive sensitivity to 850–1000-hPa IVT convergence east of the cyclone 24 h earlier (Figs. 9c,d), which suggests that the members with a more amplified

\( ^3 \)The forecast metric is computed independently for each initialization by selecting the EOF pattern representing ridge amplitude variability at 1200 UTC 14 November. Specifically, EOF1 for the 1200 UTC 9 November initialization and EOF2 for the 1200 UTC 10 November initialization.

![Fig. 7. Vertical profile of the sensitivity of \( J_{\text{RIDGE-PC2}} \) (PC units; top x axis) to the 108-h (a) water vapor mixing ratio, (b) meridional wind, and (c) temperature averaged over the black metric box shown in Fig. 6c as a function of pressure (red line) initialized at 1200 UTC 8 Nov 2013. The dots indicate vertical levels where the sensitivity is statistically significant at the 95% confidence level. The black line denotes the ensemble-mean profile (bottom x axis) [kg kg\(^1\), m s\(^{-1}\), and K in (a)–(c), respectively].](image-url)
ridge at 1200 UTC 14 November are associated with stronger IVT convergence (and implied larger latent heat release) east of the cyclone within the WCB. In addition, both initializations show negative sensitivity to the pressure trough to the south of the cyclone as early as 1200 UTC 12 November (Figs. 9e,f), which supports the importance of this feature. However, it is also noted that the two initializations indicate opposite sensitivity values to the MSLP field east of the cyclone, such that the 1200 UTC 9 November initialization exhibits positive sensitivity while the 1200 UTC 10 November initialization exhibits negative sensitivity. Overall, these results suggest that the sensitivity of the ridge amplitude to the upstream WCB is mainly consistent among several initialization times.

6. Diagnosis of WCB uncertainty

The aforementioned results suggest that uncertainty associated with the pressure trough can modulate the southerly winds that transport moisture and heat within the WCB, which in turn would yield variability in the subsequent downstream ridge. However, it is difficult to determine the source of the pressure trough uncertainty using the ridge amplitude forecast metric because the pressure trough sensitivity disappears at times prior to 96 h (not shown). As a consequence, it is of interest to determine what features modulate the uncertainty within the pressure trough by defining a new forecast metric: the 96-h MSLP averaged over the pressure trough region denoted by the red outlined area in Fig. 8d (hereafter $J_{\text{MSLP}}$). It is worth noting that the correlation between $J_{\text{RIDGE-PC2}}$ and $J_{\text{MSLP}}$ is 0.35. As a result, sources of uncertainty for the pressure trough are partially linked to sources of uncertainty for the ridge amplitude.

One hypothesis for variability in the pressure along the cold front is that it reflects uncertainty in the position of upstream surface features at earlier times such as AC and C3 (cf. Fig. 1). This hypothesis is tested by computing the sensitivity of $J_{\text{MSLP}}$ to the MSLP field at

---

4 Higher $J_{\text{MSLP}}$ denotes lower MSLP values.
earlier times (Fig. 10). At 96 h, there is a broad region of negative sensitivity west and south of C3 (located at 50°N, 70°W), extending southward ahead of the trailing cold front across the eastern United States (Fig. 10a).

This negative sensitivity suggests that lowering the pressure along the front is associated with a more amplified pressure trough (i.e., high $J_{MSLP}$) at the same valid time, which is to be expected. In addition, there is a region of

![Fig. 9](image-url)  
**FIG. 9.** Sensitivity of $J_{RIDGE-PC1}$ to the (a) 108-h advection of 250-hPa potential vorticity by the 250-hPa irrotational wind, (c) 84-h 850–1000-hPa integrated horizontal water vapor transport convergence, and (e) 72-h MSLP (shading, PC units) for forecasts initialized at 1200 UTC 9 Nov 2013. (b),(d),(f) As in (a),(c),(e), but for the sensitivity of $J_{RIDGE-PC2}$ to the 84-, 60-, and 48-h forecasts, respectively, initialized 1200 UTC 10 Nov 2013. The black lines denote the ensemble-mean values [PVU day$^{-1}$, kg s$^{-1}$, and hPa in (a),(c),(e), respectively]. The vectors in (a),(b) represent the 250-hPa irrotational winds (units: m s$^{-1}$), while the vectors in (c),(d) represent the wind components of the 850–1000-hPa integrated horizontal water vapor transport (units: kg m$^{-1}$ s$^{-1}$). The stippling indicates where the sensitivity is statistically significant at the 95% confidence level. The “L” indicates the location of the ensemble-mean forecasted cyclone C3.

### TABLE 1. Distribution of predicted MSLP for cyclone C3 by the ECMWF ensemble initialized at 1200 UTC 8 Nov 2013 compared with NWS surface analyses of C3 (NWS 2019).

<table>
<thead>
<tr>
<th>Valid time</th>
<th>Mean</th>
<th>Standard deviation</th>
<th>Maximum</th>
<th>Minimum</th>
<th>NWS analysis</th>
</tr>
</thead>
<tbody>
<tr>
<td>0000 UTC 13 Nov</td>
<td>991.70 hPa</td>
<td>5.18 hPa</td>
<td>1002.94 hPa</td>
<td>978.35 hPa</td>
<td>997 hPa</td>
</tr>
<tr>
<td>1200 UTC 13 Nov</td>
<td>983.96 hPa</td>
<td>6.85 hPa</td>
<td>1001.31 hPa</td>
<td>964.61 hPa</td>
<td>984 hPa</td>
</tr>
<tr>
<td>0000 UTC 14 Nov</td>
<td>978.78 hPa</td>
<td>9.65 hPa</td>
<td>1000.02 hPa</td>
<td>958.44 hPa</td>
<td>966 hPa</td>
</tr>
</tbody>
</table>
positive sensitivity associated with the upstream anticyclone AC on its southeastern side over Missouri at 96 h, which suggests that shifting the anticyclone south-eastward is also associated with a deeper pressure trough. Moreover, the west–east positive–negative sensitivity dipole at 96 h implies a stronger pressure gradient across the cold front. Going 12 h backward in time, the region of positive sensitivity is located on the anticyclone’s southern side over the northern Great Plains while the region of negative sensitivity remains south of the cyclone (Fig. 10b). At earlier times, both sensitive regions reduce in magnitude and spatial coverage as the negative region remains west and south of the cyclone while the positive region rotates anticyclonically around the anticyclone such that it is located southwest of the westward-shifted cyclone over western Canada and Washington at 60 h (Fig. 10d). Overall, these results suggest that uncertainty in the pressure trough along the cold front is linked to the location and extent of the upstream anticyclone and the pressure gradient across the trailing cold front.

The pressure trough is collocated with the cold front, so it is likely that variability in the pressure trough is tied to the movement of air masses on either side of lower-tropospheric fronts. The connection between the pressure trough and lower-tropospheric fronts can be evaluated by computing the sensitivity of $J_{\text{MSLP}}$ to the 850–1000-hPa $\theta$ (Fig. 11). This layer averaging of $\theta$ helps to remove variability related to small differences in the vertical position of fronts; however, the results of using one pressure level are similar (not shown). At 96 h, there is positive sensitivity to $\theta$ ahead of the cold front over the northeastern United States and negative sensitivity along the cold front extending from Kansas to western Pennsylvania (Fig. 11a), which suggests that higher $\theta$ over the northeastern United States and lower $\theta$ along the cold front over the Midwest is associated with a stronger pressure trough at the same valid time of 96 h. This negative sensitivity along the advancing cold front suggests that the members with a higher $J_{\text{MSLP}}$ are associated with a stronger cold front in this location, which is consistent with the location of higher pressure in the surface anticyclone (cf. Fig. 10a). 12 h earlier, these sensitive regions shift westward and decrease in magnitude; however, the positive sensitivity extends over the northeastern United States and farther south over the southern United States, while the negative sensitivity is shifted westward between Nebraska and Illinois (Fig. 11b). By 72 h (Fig. 11c) and 60 h (Fig. 11d) the positive sensitivity region remains located over the southern United States and located southwest of the westward-shifted cyclone while the negative sensitivity region becomes much smaller in spatial scale and is localized over western

---

**FIG. 10.** Sensitivity of $J_{\text{MSLP}}$ to the (a) 96-, (b) 84-, (c) 72-, and (d) 60-h MSLP (shading, Pa) initialized at 1200 UTC 8 Nov 2013. Black stippled regions indicate where the sensitivity is statistically significant at the 95% confidence level. Contours are the ensemble-mean MSLP (hPa). The ‘L’ indicates the location of the ensemble-mean forecasted cyclone C3.
Montana. Overall, the combination of negative and positive sensitivity located on the western and eastern sides of the \( \theta \) gradient at 96 h, respectively, would imply that a stronger \( \theta \) gradient (i.e., higher baroclinicity along the cold front) is associated with a stronger pressure trough at 96 h.

The movement of the cold air behind the cold front, and hence the pressure trough’s intensity, may be modulated by the position of upper-level features (e.g., Torn and Romine 2015; Berman et al. 2017). This hypothesis is evaluated by computing the sensitivity of \( J_{\text{MSLP}} \) to the 250-hPa PV at earlier lead times (Fig. 12). At 96 h, the ensemble-mean PV is characterized by a trough over the Midwest (Fig. 12a). Trough T3 (labeled in Fig. 12a) is characterized by a negative–positive dipole oriented west–east across its southern extent, with a small negative sensitivity region located at 41°N, 98°W and a broad positive sensitivity region extending from 40°N, 90°W to 45°N, 75°W. This sensitivity dipole suggests that shifting T3 southeastward would be associated with a deeper surface pressure trough at 96 h. Going backward in time, the sensitivity dipole remains along the base of T3 as T3 moves northwestward at the earlier times of 84 h (Fig. 12b), 72 h (Fig. 12c) and 60 h (Fig. 12d). However, prior to 72 h, the sensitivity dipole changes such that the negative sensitivity increases in magnitude and spatial extent while the positive sensitivity diminishes in area. Nonetheless, the sensitivity of the surface pressure trough to T3 suggests that the cold air behind the cold front may also be sensitive to T3. This hypothesis is evaluated by redefining the forecast metric to be the 850–1000-hPa \( \theta \) within the black box in Fig. 11b (hereafter \( J_{\text{THETA}} \)). For this metric, higher values of \( J_{\text{THETA}} \) are defined as lower area-averaged \( \theta \). If the cold front is also sensitive to T3, then the sensitivity of \( J_{\text{THETA}} \) to the 84-h 250-hPa PV (Fig. 13a) should be maximized with the trough. Indeed, the trough is associated with the same negative–positive dipole sensitivity on the north–south sides of the trough as in Fig. 12b, suggesting that a more southeastward trough is associated with decreased \( \theta \) in the box and an equatorward shift in the cold air behind the front. As a consequence, it appears that the trough modulates both the movement of the cold air and the subsequent pressure trough intensity.

One potential explanation for the sensitivity of both the pressure trough and the cold air behind the front to the position of the upstream trough is that the trough modulates the position of the cold air through “PV action at a distance” (Hoskins et al. 1985). The mechanism by which the trough modulates \( \theta \) is assessed by applying...
statistical PV inversion (e.g., Hakim and Torn 2008) to the 250-hPa PV denoted by positive sensitivity in Fig. 13a. More precisely, statistical PV inversion is applied to the 250-hPa PV anomaly, which is defined here as the difference in PV between the 10 highest and 10 lowest $J_{\text{TTHETA}}$ members. Figure 13b shows the 84-h 250-hPa PV anomaly and the domain over which the statistical inversion is applied, along with the inverted 850–1000-hPa winds and $\theta$ associated with the PV anomaly. South of the PV anomaly, the inverted winds are northeasterly and oriented normal to the cold front located across the Midwest. The northeasterly inverted winds also coincide with negative perturbation $\theta$, suggesting that a positive PV anomaly is associated with colder air behind the front. Therefore, this result suggests that uncertainty in the structure of T3 modulates the position of the cold air, which in turn impacts the pressure trough’s intensity at 96 h.

Given the importance of the trough’s position in modulating the cold front, the remainder of the paper focuses on identifying the source of the trough’s position variability. For example, the position variability of the trough could be due to earlier uncertainty in the phasing of upstream upper-level PV features (e.g., Langland et al. 2002; Lamberson et al. 2016). This hypothesis is evaluated by defining a forecast metric that represents the 84-h trough position. The forecast metric is selected by applying an EOF analysis to the 84-h 250-hPa PV field over a domain enclosing the ensemble-mean trough and defining the forecast metric as the PCs of the EOF pattern that is related to the trough position. For the EOF analysis applied to the trough, the first EOF (Fig. 14), which explains 38.4% of the variance, exhibits a negative–positive dipole that straddles the northern and southern sides of the trough; therefore it can be used to quantify the north–south position variability of the trough. Given that the EOF in Fig. 14 matches the sensitivity dipole shown in Figs. 12b and 13a, the PCs of this EOF are used as the trough position forecast metric (hereafter $J_{\text{TROUGH-PC1}}$).

The sensitivity of the trough’s position to earlier upper-level features is evaluated by computing the sensitivity of $J_{\text{TROUGH-PC1}}$ to 250-hPa PV at earlier times (Fig. 15). At 84 h, T3 is straddled on its northwestern and southern sides by a negative–positive sensitivity dipole, which suggests that a more southern trough is associated with higher $J_{\text{TROUGH-PC1}}$ at the same valid time (Fig. 15a). This sensitive region is also apparent between 72 and 60 h (Figs. 15b,c) although the positive sensitivity region

---

3 Here positive (negative) $J_{\text{TROUGH-PC1}}$ denotes a more southward (northward) trough.
diminishes in magnitude. During this time, the ridge R2 is undergoing anticyclonic wave breaking and the negative sensitivity region, located at 60 h at 60°N, 120°W, is concentrated on R2’s eastern edge where it interacts with the western edge of T3 over central Canada. In particular, one might expect that this negative sensitivity region implies a stronger negative PV anomaly, which would be associated with more northerly winds on its eastern side, and subsequently helps advect T3 farther southward at later times. The negative sensitivity region follows the apex of the ridge between 48 and 36 h (Figs. 15d,e) such that at 24 h (Fig. 15f) the negative sensitivity centered at 60°N, 160°W suggests that a more amplified ridge at 24 h is associated with a more southward trough at 84 h. Overall, these results suggest that the position of the trough is modulated by the details of R2’s anticyclonic wave breaking, such that a more amplified ridge is associated with a more southward trough in the forecast.

FIG. 13. (a) Sensitivity of $J_{\Theta}$ to the 84-h 250-hPa potential vorticity (shading, K) initialized at 1200 UTC 8 Nov 2013. Black stippled regions indicate where the sensitivity is statistically significant at the 95% confidence level. Contours are the ensemble-mean potential vorticity (PVU). (b) The 84-h inverted 850–1000-hPa potential temperature (shading, K) and 850–1000-hPa winds (vectors, m s$^{-1}$) from performing a potential vorticity inversion of the 84-h 250-hPa positive potential vorticity anomaly difference between the high and low $J_{\Theta}$ members for the forecast initialized at 1200 UTC 8 Nov 2013 (green bolded contour, PVU). The black thin contours are the ensemble-mean 850–1000-hPa potential temperature at the corresponding time (K). The “L” indicates the location of the ensemble-mean forecasted cyclone C3.
7. Summary and conclusions

This study evaluates the role of earlier WCB forecasts on ensuing ridge amplitude forecasts for a case study of a cyclogenesis event in eastern North America and the subsequent development of a highly amplified ridge over the central North Atlantic at 1200 UTC 14 November 2013. For this case, the 51-member ECMWF ensemble forecasts initialized at 1200 UTC 8 November 2013 were characterized by uncertainty in the ridge amplitude and position. The ridge amplitude is isolated by employing EOF analysis on the 144-h 250-hPa PV centered on the ridge. The source of ridge amplitude uncertainty is evaluated by applying the ensemble-based sensitivity technique to the principal components of this EOF and earlier forecast fields.

The sensitivity analysis indicates that members with a more amplified ridge are associated with greater negative PV advection by the irrotational wind immediately downstream of the deepening surface cyclone. The magnitude of the PV advection is modulated by the upstream cyclone’s divergent outflow, which is itself modulated by the cyclone’s latent heating. The downstream ridge amplitude is sensitive to the lower-tropospheric transport of heat and moisture to the east of the cyclone, such that members with a stronger WCB are associated with greater ridge amplitude. In addition, the members with greater ridge amplitude are characterized by a more intense pressure trough along the cold front to the south of the cyclone, which implies stronger geostrophic southerly flow in the WCB. Furthermore, the pressure trough’s intensity is modulated by the position of southward-moving cold air, such that a more southward movement of cold air is associated with greater baroclinicity across the cold front and thus associated with a deeper pressure trough. Finally, the position of the cold air appears to be modulated by uncertainty in upstream wave breaking, such that a more amplified upstream ridge would result in the cold air moving farther south.

This process of uncertainty in upper-level features modulating the positions of lower-tropospheric boundaries is consistent with Torn and Romine (2015) and Berman et al. (2017). However, in contrast to those studies, this case highlights a highly nonlinear manner by which the ensemble variability can grow, as demonstrated by the small sensitivity values associated with the main ridge and trough that grow faster at later times compared to earlier times. Uncertainty in the position of upper-level PV features originating over the Pacific can lead to uncertainty in subsequent wave breaking and in the position of upper-level PV features over North America. In turn, the uncertainty in these tropopause-based PV features is associated with differences in the induced lower-level flow over North America. Uncertainty in the lower-level flow modulates the speed of cold air advancing behind a cold front, which in turn modulates the baroclinicity and associated pressure trough along the front. The pressure trough modulates the strength of geostrophic southerly flow into the nearby WCB over the eastern United States and in turn the amount of moisture flux convergence. Finally, differences in the moisture flux convergence modulate the amount of latent heat release and associated upper-
tropospheric divergent outflow, and in turn the PV advection near the midlatitude waveguide. This result is consistent with the findings of Teubler and Riemer (2016) that ridge amplitude is strongly modulated by upper-tropospheric divergent outflow, and consistent with the hemisphere-scale error analysis in Baumgart et al. (2018), which highlights the contribution of forecast errors associated with upper-tropospheric divergent outflow toward subsequent ridge amplitude errors. However, this study’s investigation of localized ridge amplitude variability emphasizes the large sensitivity to both upstream PV anomalies as well as to upstream upper-tropospheric divergent outflow, especially at later times (e.g., 108 h) when the cyclone deepens and the PV ridge forms, which suggests that the contribution of upper-tropospheric divergent outflow to PV error growth can be particularly significant.

Future work should compare these results to other case studies to test the representativeness of these error growth mechanisms to other high-amplitude ridge events near upstream WCBs. Additionally, given the large sensitivity to the details of the WCB it may be prudent to identify additional observations in this region. Overall, these results suggest that errors within the WCB of cyclones can play a pivotal role in the amplification of errors along the downstream Rossby waveguide.

Acknowledgments. The first author would like to thank Drs. Dan Keyser and Ben Moore for insightful discussion of this work. The first author also thanks the scientists, and in particular early career scientists, of the Transregional Collaborative Research Center SFB/TRR 165 “Waves to Weather” (W2W), which is funded by the German Research Foundation (DFG), for their excellent feedback during a visit of the author sponsored by W2W. In particular, the first author would like to thank Drs. Michael Riemer, Christian Grams, George Craig, Heini Wernli, and Olivia Martius for many fruitful discussions. We also thank three anonymous reviewers for their insightful comments on an earlier version of this manuscript.

FIG. 15. Sensitivity of $J_{\text{TROUGH-PC1}}$ to the (a) 84-, (b) 72-, (c) 60-, (d) 48-, (e) 36-, and (f) 24-h 250-hPa potential vorticity (shading, PC units) initialized at 1200 UTC 8 Nov 2013. Black stippled regions indicate where the sensitivity is statistically significant at the 95% confidence level. Contours are the ensemble-mean potential vorticity (PVU). The “L” indicates the location of the ensemble-mean forecasted cyclone C3.
This work was supported by National Science Foundation Grant 1461753.

REFERENCES


